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Detailed tectonic reconstructions of the Western Mediterranean region for the last 35 Ma, insights on driving mechanisms

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Abstract – Slab retreat, slab tearing and interactions of slabs are first-order drivers of the deformation of the overriding lithosphere. An independent description of the tectonic evolution of the back-arc and peripheral regions is a pre-requisite to test the proposed conceptual, analogue and numerical models of these complex dynamics in 3-D. We propose here a new series of detailed kinematics and tectonic reconstructions from 35 Ma to the Present shedding light on the driving mechanisms of back-arc rifting in the Mediterranean where several back-arc basins all started to form in the Oligocene. The step-by-step backward reconstructions lead to an initial situation 35 Ma ago with two subduction zones with opposite direction, below the AlKaPeCa block (*i.e.* belonging to the Alboran, Kabylies, Peloritani, Calabrian internal zones). Extension directions are quite variable and extension rates in these basins are high compared to the Africa-Eurasia convergence velocity. The highest rates are found in the Western Mediterranean, the Liguro-Provençal, Alboran and Tyrrhenian basins. These reconstructions are based on shortening rates in the peripheral mountain belts, extension rates in the basins, paleomagnetic rotations, pressure-temperature-time paths of metamorphic complexes within the internal zones of orogens, and kinematics of the large bounding plates. Results allow visualizing the interactions between the Alps, Apennines, Pyrenean-Cantabrian belt, Betic Cordillera and Rif, as well as back-arc basins. These back-arc basins formed at the emplacement of mountain belts with superimposed volcanic arcs, thus with thick, hot and weak crusts explaining the formation of metamorphic core complexes and the exhumation of large portions of lower crustal domains during rifting. They emphasize the role of transfer faults zones accommodating differential rates of retreat above slab tears and their relations with magmatism. Several transfer zones are identified, separating four different kinematic domains, the largest one being the Catalan-Balearic-Sicily Transfer Zone. Their integration in the wider Mediterranean realm and a comparison of motion paths calculated in several kinematic frameworks with mantle fabric shows that fast slab retreat was the main driver of back-arc extension in this region and that large-scale convection was a subsidiary driver for the pre-8 Ma period, though it became dominant afterward. Slab retreat and back-arc extension was mostly NW-SE until ~ 20 Ma and the docking of the AlKaPeCa continental blocks along the northern margin of Africa induced a slab detachment that propagated eastward and westward, thus inducing a change in the direction of extension from NW-SE to E-W. Fast slab retreat between 32 and 8 Ma and induced asthenospheric flow have prevented the transmission of the horizontal compression due to Africa-Eurasia convergence from Africa to Eurasia and favored instead upper-plate extension driven by slab retreat. Once slab retreat had slowed down in the Late Miocene, this N-S compression was felt and recorded again from the High Atlas to the Paris Basin.

Keywords: Western Mediterranean / plate reconstructions / slab retreat / transfer zones / exhumation / weak crust

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Résumé – Reconstructions détaillées de la Méditerranée occidentale depuis 35 Ma, implications en terme de mécanismes moteur. Le retrait des panneaux plongeants dans les zones de subduction, leurs éventuelles déchirures et les interactions entre ces panneaux sont des moteurs de premier ordre de la déformation lithosphérique. Une description indépendante de l'évolution tectonique des régions arrière-arc et de leur périphérie est un prérequis indispensable pour tester les différents modèles conceptuels, analogiques ou numériques de ces interactions complexes en 3-D. Nous proposons une série de nouvelles reconstructions cinématiques et tectoniques de la Méditerranée occidentale depuis 35 Ma qui éclairent les mécanismes moteurs de l'extension arrière-arc. La Méditerranée est caractérisée par plusieurs domaines arrière-arc qui ont tous commencé à se former pendant l'Oligocène inférieur. La reconstruction pas-à-pas en remontant le temps conduit à une situation où deux zones de subduction à vergence opposées sont actives sous le bloc AlKaPeCa. Les directions d'extension sont variables selon les bassins et les taux d'extension sont élevés par rapport à ceux de la convergence Afrique-Eurasie. Les taux les plus élevés sont observés en Méditerranée occidentale dans le Bassin Liguro-Provençal, la Mer d'Alboran et la Mer Tyrrhénienne. Nos reconstructions sont basées sur les taux de raccourcissement dans les chaînes de montagnes qui entourent ces bassins, les taux d'extension dans les bassins, les rotations paléomagnétiques, les chemins pression-température-temps des complexes métamorphiques dans les zones internes des orogènes et la cinématique des grandes plaques alentour. Les résultats permettent de visualiser les interactions entre les Alpes, les Pyrénées, les Apennins, les Cordillères Bétiques et le Rif, et les bassins arrière-arc. Ces derniers se forment à l'emplacement des chaînes de montagnes éocènes auxquelles sont superposés des arcs volcaniques et donc dans des domaines de croûte chaude et épaisse expliquant la genèse de *metamorphic core complexes* et l'exhumation de larges portions de croûte inférieure pendant le rifting. Les reconstructions montrent l'importance des zones de transfert au-dessus des déchirures des panneaux plongeants et leurs relations avec les arcs magmatiques. Plusieurs zones de transfert sont identifiées, la plus importante étant la zone de transfert Catalogne-Baléares-Sicile. L'intégration de ces reconstructions dans le cadre plus large de la Méditerranée et une comparaison des chemins cinématiques calculés dans des repères cinématiques différents montrent que le retrait des panneaux plongeants est le moteur principal de la déformation arrière-arc et que la convection à grande échelle et la convergence Afrique-Eurasie sont des facteurs de deuxième ordre pour la période antérieure à 8 Ma et qu'elles redeviennent primordiale ensuite. Le retrait des panneaux plongeants est essentiellement N-S ou NW-SE jusqu'à environ 20 Ma, moment où se produit l'accostage des blocs du domaine AlKaPeca sur la marge nord-africaine, induisant une déchirure qui se propage ensuite vers l'ouest et vers l'est et un changement de la direction d'extension de N-S à E-W. Le retrait des panneaux plongeants entre 32 et 8 Ma a empêché la transmission des contraintes horizontales dues à la convergence Afrique-Eurasie et favorisé au contraire l'extension arrière-arc. Quand ce retrait ralentit au cours du Miocène supérieur à l'ouest, la compression N-S est à nouveau enregistrée et c'est elle qui préside aux déformations actives dans la région depuis le Haut-Atlas jusqu'au bassin Parisien.

Mots clés : Méditerranée occidentale / reconstructions cinématiques / retrait du panneau plongeant / zones de transfert / exhumation / croûte faible

1 Introduction

Slab dynamics is a first-order driver of the deformation of the overriding plates of subduction zones, especially when the subduction zone is constrained within a narrow space like the Mediterranean region (Carminati *et al.*, 1998a; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a, 2001b, 2003; Spakman and Wortel, 2004; Jolivet *et al.*, 2009). The geometry of slabs at depth in this area is now well constrained thanks to seismic tomography and the mantle flow associated with slab retreat can be described using seismic anisotropy (Carminati *et al.*, 1998a, 1998b; Wortel and Spakman, 2000; Piromallo and Morelli, 2003; Spakman and Wortel, 2004; Jolivet *et al.*, 2009; Bezada *et al.*, 2013; Faccenna *et al.*, 2014; Fichtner and Villaseñor, 2015; Villaseñor *et al.*, 2015). Beside the rather simple back-arc extension related to slab retreat in the Liguro-Provençal basin and Tyrrhenian Sea, the most complex regions, such as the tight Betic-Rif Arc or the Alps-Apennines junction have been described with different dynamic models involving slab tearing and interactions between slabs (Platt and Vissers, 1989; Lonergan and White, 1997; Jolivet *et al.*, 2006, 2008; Vignaroli *et al.*, 2008; van Hinsbergen *et al.*, 2014). Several key-events have been recognized in the Western Mediterranean such as the initiation of back-arc extension at ~32 Ma or the Late Miocene resumption of N-S compression in the Betic Cordillera that can be associated to the behavior of slabs at depth (Faccenna *et al.*, 2001a, 2001b; Vignaroli *et al.*, 2008; Spakman *et al.*, 2018). In order to test and discuss these conceptual, analogues and numerical models that address the interactions between slab dynamics and crustal deformation in the Mediterranean region, we propose a new set of detailed kinematic and tectonic reconstructions, obtained independently from these models.

Whether crustal deformation is driven by far-field forces guided by the resistant lithospheric mantle (lithospheric stress-guide) or by the mantle flowing underneath has been a debated question since the early days of the plate tectonics theory (Elsasser, 1968; McKenzie, 1969). This question is crucial in regions where different sources of forces are likely to be at work such as the complex back-arc regions of the Mediterranean realm. The Aegean region was used to propose different types of conceptual models, tested numerically, involving one or several of these different forces, mantle flow *vs* lithospheric stress-guide (McKenzie, 1972, 1978; Le Pichon and Angelier, 1981a, 1981b; Taymaz *et al.*, 1991; Armijo *et al.*, 1999; Wortel and Spakman, 2000; Spakman and Wortel, 2004; Govers and Wortel, 2005; Jolivet *et al.*, 2009, 2013; Faccenna and Becker, 2010; Becker and Faccenna, 2011; Capitanio, 2014; Magni *et al.*, 2014; Sternai *et al.*, 2014). The localization and propagation of the North Anatolian Fault and the concomitant extension in the Corinth Rift extension in the overriding plate of the Hellenic subduction have for instance been diversely interpreted as (i) the result of rigid extrusion of Anatolia caused by the Arabia-Eurasia collision (McKenzie, 1972, 1978; Armijo *et al.*, 1999), or (ii) a consequence of slab detachment and tearing, involving a component of basal shear by the asthenospheric mantle due to slab retreat (Jolivet *et al.*, 2009, 2013; Sternai *et al.*, 2014; Menant *et al.*, 2016b). The possible role of large-scale convection with the interference of the Afar plume was also discussed (Faccenna *et al.*, 2013b).

Similar questions can be addressed in the Western Mediterranean (Fig. 1) where large and fast displacements have been recorded since about 35 Ma and several drastic changes in the tectonic regime were documented. There, too, several models were proposed, involving slab retreat and associated small-scale asthenospheric flow as a main driver (Carminati *et al.*, 1998a; 1998b; Rosenbaum *et al.*, 2002; Faccenna *et al.*, 2004; Rosenbaum and Lister, 2004; Spakman and Wortel, 2004; Faccenna *et al.*, 2007; Jolivet *et al.*, 2009; Vignaroli *et al.*, 2009), stress transmission within the lithosphere causing buckling and strain localization (Casas Sainz and Faccenna, 2001; Dielforder *et al.*, 2019), or control of crustal deformation by the flow of mantle due to large-scale convection dragging the slab northward (Spakman *et al.*, 2018). These different drivers were probably all active at some period, but not all at the same time and their respective roles might have been more or less important through time. So far, several drastically different tectonic models were proposed for the long-term evolution of the Western Mediterranean: (i) lithospheric delamination (Platt and Vissers, 1989; Platt *et al.*, 2003a, 2013), (ii) a single north-dipping subduction progressively retreating (Lonergan and White, 1997; Rosenbaum *et al.*, 2002; Faccenna *et al.*, 2004; Jolivet *et al.*, 2006), (iii) two opposite subductions (Chalouan *et al.*, 2001; Michard *et al.*, 2002; Schettino and Turco, 2006; Vergés and Fernàndez, 2012; Leprêtre *et al.*, 2018) or different combinations of some aspects of these models, with more or less slab retreat in the westernmost Mediterranean. Figure 1B shows four different configurations before the inception of back-arc extension (Michard *et al.*, 2002; Rosenbaum *et al.*, 2002; Lacombe and Jolivet, 2005; Vergés and Fernàndez, 2012; van Hinsbergen *et al.*, 2014; Leprêtre *et al.*, 2018). The differences pertain to the initial length of the subduction zone, either restricted to the latitude of the Baleares or continuing to the future Alboran region and the polarity of subduction, either a single NW-ward subduction or two subduction zones with opposite polarities. This initial situation has important consequences in terms of the dynamics of slab retreat as explored by Chertova *et al.* (2014) with 3-D numerical modelling. One of the goals of this paper is to investigate this question through a series of kinematic reconstructions with data independent of any dynamic model.

The only independent long-term record of these complex interactions is the geological record, and more specifically tectonics and magmatism. In such a complex 3-D environment, with fast displacements and sudden kinematic changes, a prerequisite to any discussion of the dynamic interactions is to possess a detailed account of the tectonic evolution through time. This is why we propose here a series of detailed tectonic reconstructions of the whole Western Mediterranean region, made with GPlates (Boyden *et al.*, 2011), from the Apennines subduction to the Gibraltar arc, since the first evidence of back-arc extension in the Mediterranean some 35 Ma ago.

Our reconstructions are made backward step-by-step and built on independent geological and tectonic constraints, including published estimates of field-based shortening rates in convergence zones, estimates of extension rates in back-arc basins, paleomagnetic rotations, pressure-temperature-time history of metamorphic complexes, as well as information on the timing derived from stratigraphic and radiometric studies. The reconstructions are then integrated within the whole Mediterranean realm with the addition of the eastern Mediterranean reconstructions of Menant *et al.* (2016a).

We first describe the reconstructions and then discuss their geodynamic implications. The results show complex interactions between the Alps, the Apennines, the Pyrenees, the Betic Cordillera and the Rif, as well as the opening of back-arc basins. Back-arc basins formed at the emplacement of mountain belts that were locally superimposed by volcanic arcs and extensional basins, and were thus supported by thick, hot and weak continental crusts, leading to the formation of metamorphic core complexes and the exhumation of lower crustal domains during rifting. The initial configuration we obtain before the inception of slab retreat is intermediate between that of Vergés and Fernàndez (2012) and Michard *et al.* (2002) with two subduction zones with opposite polarities. The reconstructions also emphasize the role and timing of transfer fault zones accommodating tears in the slabs and their relations with magmatism. We discuss the different styles of transfer zones in the overriding plates above slab tears. Their integration in the wider Mediterranean framework and comparison of motion paths calculated in several kinematic frameworks with the mantle fabric deduced from seismic anisotropy shows that fast slab retreat was the main driver of back-arc extension in this region and that it prevented the transmission of compressional stresses across the Eurasia-Africa plate boundary zone. The situation changed at ~8 Ma when the slab retreating below the Gibraltar arc had migrated far enough westward for the N-S compression be felt again in the whole Western Mediterranean region.

2 Geodynamic context

Although the present-day setting of the Western Mediterranean (Fig. 1) is mostly characterized by compressional deformation along a N-S direction (Billi *et al.*, 2011; Faccenna *et al.*, 2014), the bulk geometry of this region was formed in a back-arc extensional environment between 35 and 8 Ma (Réhault *et al.*, 1984; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a, 2001b; Rosenbaum *et al.*, 2002; Spakman and Wortel, 2004). Extension is here the overriding plate response to the eastward or westward retreat of lithospheric slabs, which were subsequently fragmented in several segments subducting nowadays below Calabria, the Apennines and the Gibraltar

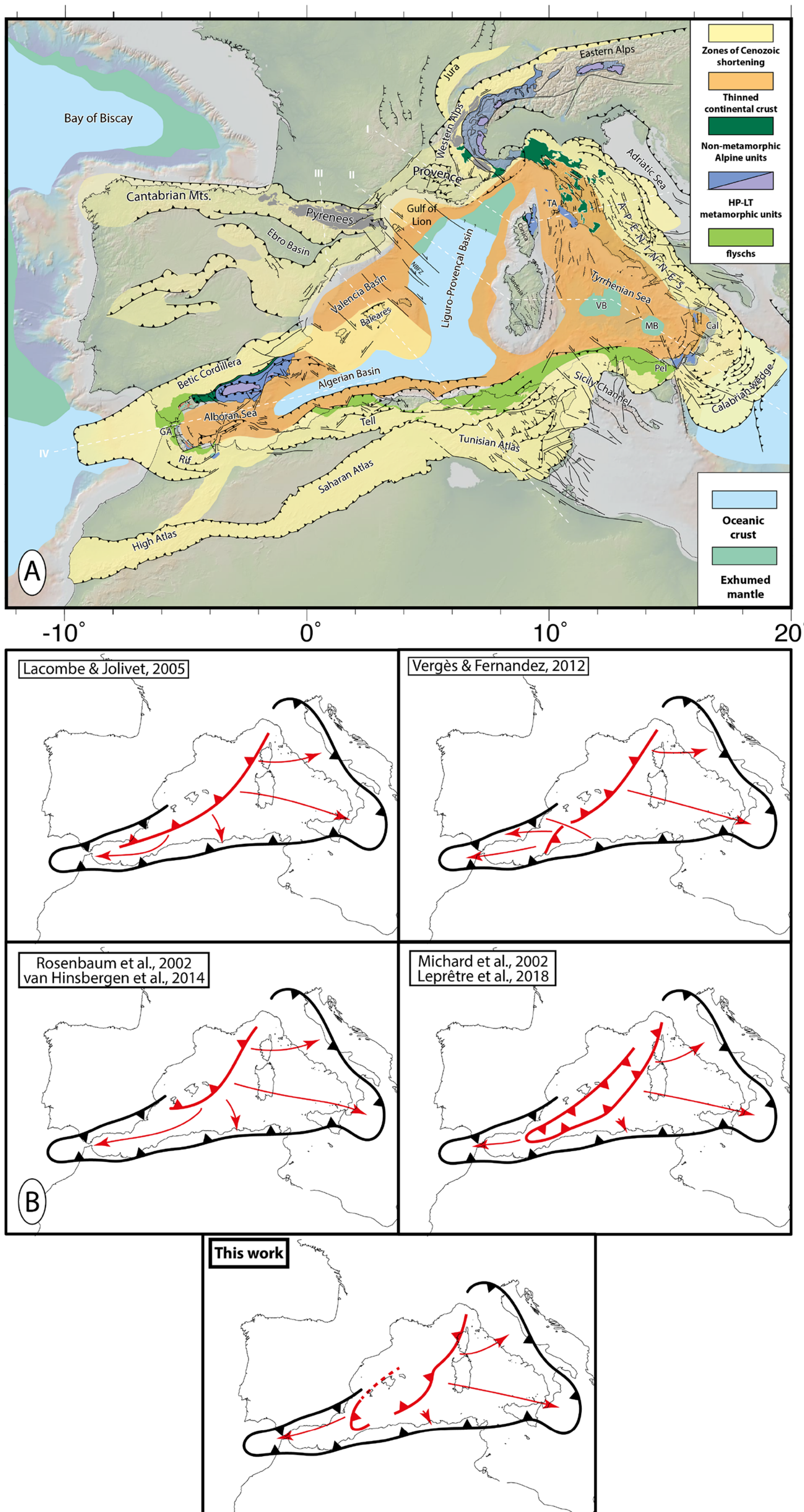


Fig. 1. Tectonic map of the Western and Central Mediterranean region (A). White dashed lines stand for the reconstructed cross-sections of Figure 3. Cal: Calabria, GA: Gibraltar Arc, MB: Marsili Basin, Pel: Peloritani, TA: Tuscan Archipelago, VB: Vavilov Basin. (B) Schematic presentation of four recent published kinematic hypotheses (Rosenbaum *et al.*, 2002; Lacombe and Jolivet, 2005; Vergés and Fernàndez, 2012; van Hinsbergen *et al.*, 2014) compared to the new model described in this paper.

arcs (Malinverno and Ryan, 1986; Dewey *et al.*, 1989; Doglioni *et al.*, 1997; Carminati *et al.*, 1998a; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a, 2001b; Rosenbaum and Lister, 2004).

Several basins were formed after 35 Ma, starting with rifting between Provence and Sardinia, shaping the Gulf of Lion passive margin and the oceanic domain of the Liguro-Provençal Basin, associated with the counterclockwise rotation of the more or less rigid Corsica-Sardinia block during the Oligocene and early Miocene (Réhault *et al.*, 1984). Extension then jumped east of Corsica and Sardinia, forming the Tyrrhenian Sea (Réhault *et al.*, 1987; Kastens *et al.*, 1988). Crustal thinning was extreme in the southern Tyrrhenian Sea where the emplacement of oceanic crust in the Pliocene and Quaternary was classically proposed (Kastens and Mascle, 1990; Patacca *et al.*, 1990; Jolivet *et al.*, 1998). More recent studies however suggest that true oceanic spreading did not happen in the Marsili and Vavilov basins (Fig. 1), where large volcanoes were emplaced within highly thinned continental crust or exhumed mantle (Prada *et al.*, 2014, 2018). Extension was also active above the retreating Gibraltar slab from the late Oligocene-early Miocene to the Late Tortonian, forming the Alboran Sea between Iberia and the northern coast of Africa (Comas *et al.*, 1992, 1999; Frizon de Lamotte *et al.*, 2004; Mauffret *et al.*, 2007; Medaouri *et al.*, 2014; Do Couto *et al.*, 2016). During the first stages (27–20 Ma) extension in this region was mostly N-S and then changed abruptly some 20 Ma from N-S to E-W (Crespo-Blanc *et al.*, 1994; Crespo-Blanc, 1995; Martinez-Martinez and Azañon, 1997; Jolivet *et al.*, 2008; Crespo-Blanc *et al.*, 2016). 8 Ma ago, extension slowed down drastically and the main tectonic regime became compressional (Weijermars *et al.*, 1985; Mora, 1993; Augier *et al.*, 2005c; Meijninger and Vissers, 2006; Billi *et al.*, 2011).

The internal zones of the Western Mediterranean mountain belts are generally thought to belong to a former independent block, named AlKaPeCa, now dispersed by later back-arc extension (Bouillin *et al.*, 1986). Its remnants are found at the periphery of the Alboran domain in the Betic Cordillera (Spain) and the Rif (Morocco), the Kabylies in the Tell (Algeria and Tunisia), the Peloritani range in Sicily and in Calabria (Bouillin *et al.*, 1986; van Hinsbergen *et al.*, 2014). Whether it constituted a single block or a series of individual smaller blocks is in fact unknown.

During this period dominated by back-arc extension, mountain belts were formed above the retreating subduction zones, the Apennines in the east, the Maghrebides in the south (Tell and Rif), and the Betics along the southern margin of Iberia. Slab retreat and coeval back-arc extension proceeded at high rates, ranging from 2 cm/yr to more than 10 cm/yr (Nicolosi *et al.*, 2012; van Hinsbergen *et al.*, 2014).

During the formation of these mountain belts (syn-orogenic stage) and during subsequent extension (post-orogenic stage), metamorphic core complexes (MCCs) were exhumed. The syn-orogenic period is characterized by a good preservation of high-pressure and low-temperature (H*P*-L*T*) metamorphic parageneses, while the late-orogenic period sees intense retrogression and often reheating, resulting in a more or less complete overprint of the H*P*-L*T* parageneses in high-temperature and low-pressure conditions (H*T*-L*P*) (Jolivet *et al.*, 1998; Platt *et al.*, 2013), an evolution resembling that of the Aegean region (Wijbrans and McDougall, 1988; Wijbrans *et al.*, 1993; Trotet *et al.*, 2001; Jolivet *et al.*, 2003; Ring *et al.*, 2010; Grasemann *et al.*, 2012; Laurent *et al.*, 2016; Roche *et al.*, 2016). The exhumation of these MCCs is associated with intense shearing deformation and the development of km-thick strain gradients that affect large portions of the middle and lower crust. A good example is the case of the Betic Cordillera where exhumation was mostly associated with N-S-trending stretching lineations and top-to-the north sense of shear before 20 Ma and by E-W-trending lineations and top-to-the west sense of shear afterward (Azañón *et al.*, 1997; Balanya *et al.*, 1997; Jolivet *et al.*, 2003, 2008; Williams and Platt, 2018). This change of direction is first observed in coeval basins deposited on top of the exhumed domains (Crespo-Blanc *et al.*, 1994; Crespo-Blanc, 1995; Azañon and Crespo-Blanc, 2000; Augier *et al.*, 2013).

This first-order evolution results from the convergence between Africa and Eurasia, but the driving mechanisms of crustal deformation can be understood in various ways. Stresses can be transmitted through the lithospheric stress-guide or from below by the asthenospheric mantle flowing underneath. Mantle flow can be caused by the large-scale convection or more local mantle flow related to slab retreat (Jolivet *et al.*, 2009; Faccenna and Becker, 2010; Faccenna *et al.*, 2013a; Faccenna *et al.*, 2014; Sternai *et al.*, 2014; Menant *et al.*, 2016b). Mountain building and back-arc extension are often coeval along a given transect (Jolivet *et al.*, 1994, 1996, 1998) but also in 3-D at the junction between two subduction zones such as the Alps and the Apennines, where the westward migration of the Alps thrust front and the eastward retreat of the Apennines front and opening of the Liguro-Provençal and Tyrrhenian basins are coeval, probably linked with toroidal flow underneath (Maffione *et al.*, 2008; Vignaroli *et al.*, 2008, 2009).

During the last 35 Ma, several major tectonic changes are recorded. The first one is the progressive ending of compressional deformation in the Pyrenees and Central Iberian Range while it was still active in the Rif and the Betics, as well as in the Apennines (Comas *et al.*, 1999; Vergés and Sàbat, 1999; Jolivet and Faccenna, 2000; Rosenbaum *et al.*, 2002; Vergés *et al.*, 2002; Mouthereau *et al.*, 2014). A second one is the inception of slab retreat all around the Mediterranean realm between 35 and 30 Ma (Jolivet and Faccenna, 2000). The third one is the end of back-arc extension some 8 Ma ago in the Alboran region (Augier *et al.*, 2005c; Meijninger and Vissers, 2006; Billi *et al.*, 2011; Augier *et al.*, 2013; Janowski *et al.*, 2017). All these events require a dynamic explanation. Are these modifications of the strain regime due to local events or parts of a single long-term mechanism, intrinsic to the convergence process?

In order to address these questions we first performed detailed tectonic reconstructions from 35 Ma to the Present and then discuss their implications in different kinematic frameworks in order to emphasize the individual drivers.

3 Previous reconstructions

Several reconstructions were published for the same region. After early proposals involving the rotation of Corsica and Sardinia (Argand, 1924; Carey, 1958), Alvarez *et al.* (1974) and Boccaletti and Guazzone (1974) proposed models showing the successive opening of the Liguro-Provençal Basin and then the Tyrrhenian Sea. The first attempt using a plate tectonics approach was by Bayer *et al.* (1973) who proposed an identification of oceanic magnetic anomalies based on the compilation of two airborne surveys and made reconstructions of the Liguro-Provençal Basin evolution where the Corsica-Sardinia block was however not rotating. Cohen (1980) proposed a series of reconstructions of the same region based on the geometry of transform faults, paleomagnetic data obtained in the margins of the domain and a compilation of geological data. These reconstructions involve the counter-clockwise rotation of a rigid Corsica-Sardinia block. Réhault *et al.* (1984) used the geometry of magnetic anomalies to infer the kinematics of opening and the position of transform faults, as well as paleomagnetic data supporting the rotation of Sardinia, the structure of the basin and its margins, and the timing of subsidence and heat-flow measurements to describe the evolution of the Liguro-Provençal Basin as a back-arc basin. Dewey *et al.* (1989) used a new study of the Atlantic fracture zones based on SEASAT data to propose an updated history of the Africa-Europe convergence. Within this large-scale kinematic framework they proposed detailed reconstructions of the Western Mediterranean including the Liguro-Provençal Basin, the Tyrrhenian Sea and the Apennines. They used a new interpretation of the rotation of Corsica and Sardinia and the geological history of the basins and mountain belts around. Gueguen *et al.* (1998) proposed a more detailed tectonic evolution of the entire Western Mediterranean region back to the early Miocene based on the geometry of transform faults and the balancing of sections across the basins based on available seismic profiles. Rosenbaum *et al.* (2002) made reconstructions of the entire Western Mediterranean region since 30 Ma, using a variety of data types, from structural geology, metamorphic petrology, magmatic history, sedimentary basins, paleomagnetic data and geophysics (Fig. 1B). This was the first attempt using an integrated plate kinematics software (PLATYPLUS). Michard *et al.* (2002) proposed a different scenario where two subductions are facing each other (Fig. 1B), one being the future Apennines subduction and the second one, dipping to the SW the southern extension of the Alpine ocean. Jolivet *et al.* (2003) proposed reconstructions of the whole Mediterranean region to discuss the evolution of H*P*-L*T* metamorphic units and Lacombe and Jolivet (2005) showed a focus on the Western Mediterranean to discuss the relations between Corsica and Provence. Dèzes *et al.* (2004) showed reconstructions of the same region to discuss the evolution of the West European Rift. Schettino and Turco (2006) focused their study on the Liguro-Provençal Basin. They did a detailed kinematic restoration using the identification of transform faults on magnetic anomaly maps to obtain the flow lines and position of rotation poles, an interpretation of the magnetic anomalies to date the opening and a balancing of crustal profiles. Mantovani *et al.* (2009) published detailed reconstructions of the evolution of the Tyrrhenian Sea and the Apennines since the Middle Miocene. Van Hinsbergen *et al.* (2014) performed detailed reconstructions of the Western Mediterranean made with *GPlates* with the aim of addressing the cause of initiation of slab rollback in the Oligocene. These reconstructions are tested in particular with the amount of subducted lithosphere seen in seismic tomography data sets. Van Hinsbergen *et al.*'s (2014) work is the closest to ours in terms of methodology, except that we do not use seismic tomography to test the model and rely only upon the surface geological record as constraints. In the following we discuss the differences between the two models. More recently, Leprêtre *et al.* (2018) proposed new reconstructions of the westernmost Mediterranean based on a synthesis of geological studies in the Rif and Tell.

These reconstructions focused on the Western Mediterranean or only on the Liguro-Provençal Basin were published at the same time as reconstructions encompassing much larger regions such as all the belts born from the Tethys Ocean, including the Mediterranean realm (Dercourt *et al.*, 1986; Ricou *et al.*, 1986; Dercourt *et al.*, 1993; Stampfli *et al.*, 2002; Gaetani *et al.*, 2003) or the whole of Western Europe, Arctic and Western Tethys (Ziegler, 1999). The most recent of these large-scale reconstructions was recently published by van Hinsbergen *et al.* (2019). It shows a detailed evolution of the Western Tethys and Mediterranean region since 200 Ma. These large-scale reconstructions show the long-term evolution of the Western Mediterranean but not the detail we need for discussing the driving parameters of crustal deformation since 35 Ma.

In all these reconstructions, paleomagnetic data gathered in the continental areas surrounding the basins are crucial constraints, together with the large geological and geophysical data sets. At variance with large oceanic domains, the use of oceanic magnetic anomalies is much less powerful because the classical Vine and Matthews's (1963) type of anomalies are difficult to recognized in the oceanic domains of the Western Mediterranean, yet they have been used in some of these reconstructions (Bayer *et al.*, 1973; Schettino and Turco, 2006). The nature of the crust in the "oceanic" domain is also quite disputed and several studies conclude to an atypical oceanic crust, if not only exhumed mantle. This is true of the Tyrrhenian Sea (Prada *et al.*, 2014, 2018) and of the Liguro-Provençal Basin (Rollet *et al.*, 2002; Dannowski *et al.*, 2020). Magnetic anomalies are indeed short and their symmetry on either side of a potential spreading axis is particularly unclear. Some of these anomalies clearly correspond to local large volcanoes and not to a typical oceanic crust. Similarly, using these anomalies to determine the geometry of transfer or transform faults is difficult. We thus do not rely on the magnetic anomaly pattern for our reconstructions.

4 Reconstruction method and model inputs

Reconstructions were made with the help of *GPlates*, a free plate kinematics software developed by EarthByte (http://www.gplates.org; Boyden *et al.*, 2011). GPlates works with "features", polygones or polylines (mostly polylines in our case), that are rotated about Eulerian rotation poles. For reconstructing the Western Mediterranean, and integrating these reconstructions in the whole Mediterranean framework, we used more than 400 such features (Fig. 2). Several tables are provided in the Supplementary Materials of this paper. Supplementary Material #1 contains (i) the geological, geophysical and paleomagnetic constraints used for the reconstructions, (ii) the associated list of references and (iii) the list of features (rigid polygons) used for the reconstructions. Supplementary Material #2 gives the GPplates rotation file with the rotation poles and angles of rotation, assorted with the bibliographic references (all included in the list of

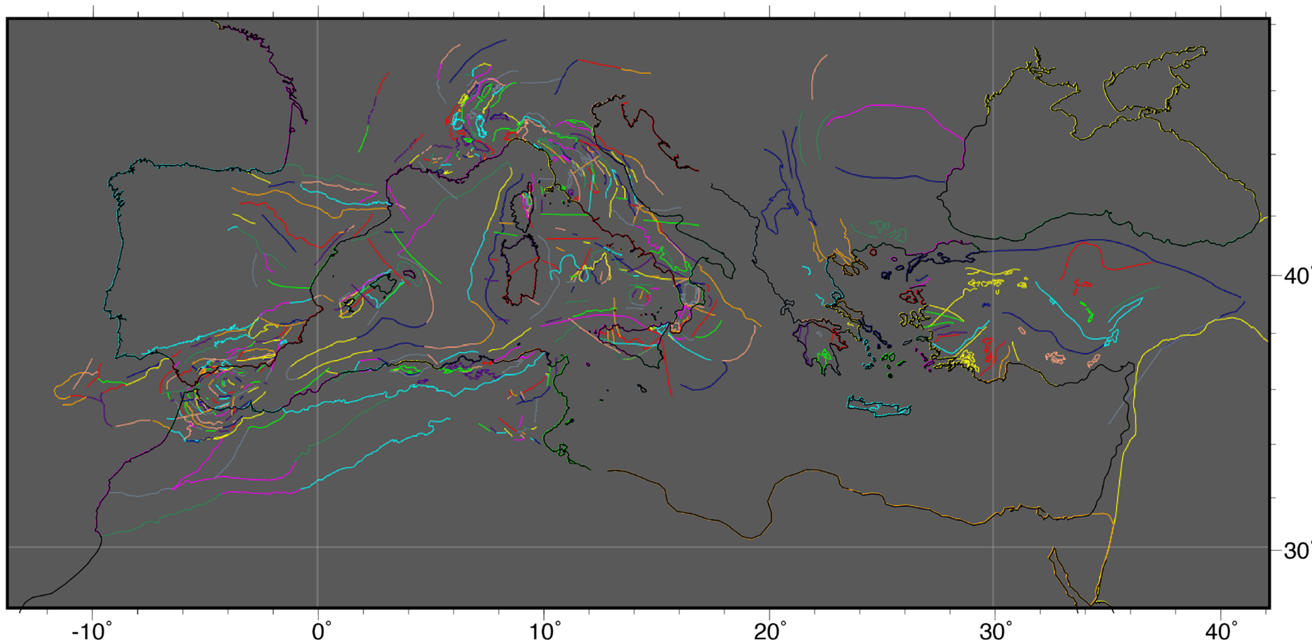


Fig. 2. Features (rigid polygons or polylines) used in the GPlates reconstructions shown with various colors. The list of these features is given in Supplementary Material #1 and the rotation file in Supplementary Material #2.

references at the end of this paper). The angles of rotation have been calculated to fit the amount of shortening or extension extracted from the listed reference. Supplementary Material #3 is an additional figure to discuss the significance of crossing motion-paths. The movie *reconstructions-Medit-GPlates.mp4* shows the displacement of all blocks used for reconstructing this area.

Although the rotating features are rigid bodies, we take into account internal deformation of blocks by using polylines at their limits that rotate with different angles (Fig. 3A), thus accommodating shortening or extension like in van Hinsbergen *et al.* (2014). The general reconstruction procedure is also similar to that used by Menant *et al.* (2016a) for the eastern Mediterranean. Because we use polylines, the blocks are deformable at will. Adjusting the rotation angle simply increases or decreases the amount of shortening or extension. In between the two rotating block limits, which converge or diverge, the deformation is supposed to be pure shear in map view, homogeneously distributed across the deforming block. Converging block limits lead to crustal thickening and diverging block limits to extension. The amount of extension or shortening is a direct consequence of the imposed rotation angle. Thus, rifted margins such as the Gulf of Lion are reconstructed using balanced restoration as in Jolivet *et al.* (2015a, 2015b) and mountain belts in convergence zones are zones where crustal thickening is accommodated based on published balanced cross-sections.

The reconstructions are contained within the Mediterranean region that is surrounded by large plates, Africa, Eurasia, Iberia, and Arabia. We use the model recently published by Nirrengarten (2016) and Nirrengarten *et al.* (2018) and references therein for the motion of these large plates, an adaptation of Seton *et al.* (2012). This model uses oceanic magnetic anomalies in large oceans and takes into account the deformation of the margins during rifting. An alternative model was published by Macchiavelli *et al.* (2017), which does not differ much for the 35–0 Ma period. Besides the kinematic constraints imposed by the motion of the large plates around, we calculate kinematic parameters for the Western and Central Mediterranean domains using geological and paleomagnetic data (see Supplementary Materials #1 and #2 and references therein).

The kinematic history of the deforming region is constrained by the timing and rates of extension in basins and shortening in mountain belts, the activation and inactivation times of thrusts and detachments, the age of emplacement of oceanic crust, information on the migration of the magmatic arc during slab retreat (Fig. 3B). It is also constrained by the geometrical fit of oceanic domain margins like the Liguro-Provençal basin and by paleomagnetic data used to constrain the rotation of individual blocks such as Corsica-Sardinia (Gattacceca *et al.*, 2007; Advokaat *et al.*, 2014), oroclinal bending of Apennines (Sagnotti *et al.*, 2000; Caricchi *et al.*, 2014), rotations in Sicily (Speranza *et al.*, 2003), rotations in the Gibraltar arc (Mattei *et al.*, 2006; Crespo-Blanc *et al.*, 2016) and a recent synthesis of formation of the tight Western Mediterranean arcs based on paleomagnetic data (Cifelli *et al.*, 2016). Additional information comes from the pressure-temperature-time constraints obtained on exhumed metamorphic rocks that are used as a proxy for burial as a function of time, bringing some insights on the thermal regime and active geodynamic processes. Hence, H*P*-L*T* and H*T*-L*P* metamorphic events are diagnostic of syn- and post-orogenic environments, respectively. Kinematic indicators in the exhumed MCCs are also used to construct the kinematic model. We avoided using any *a priori* dynamic ideas to construct the kinematic model, using only published kinematic data for the large plates surrounding our study area and geological and paleomagnetic data for the internal kinematics. We thus consider that our kinematic model

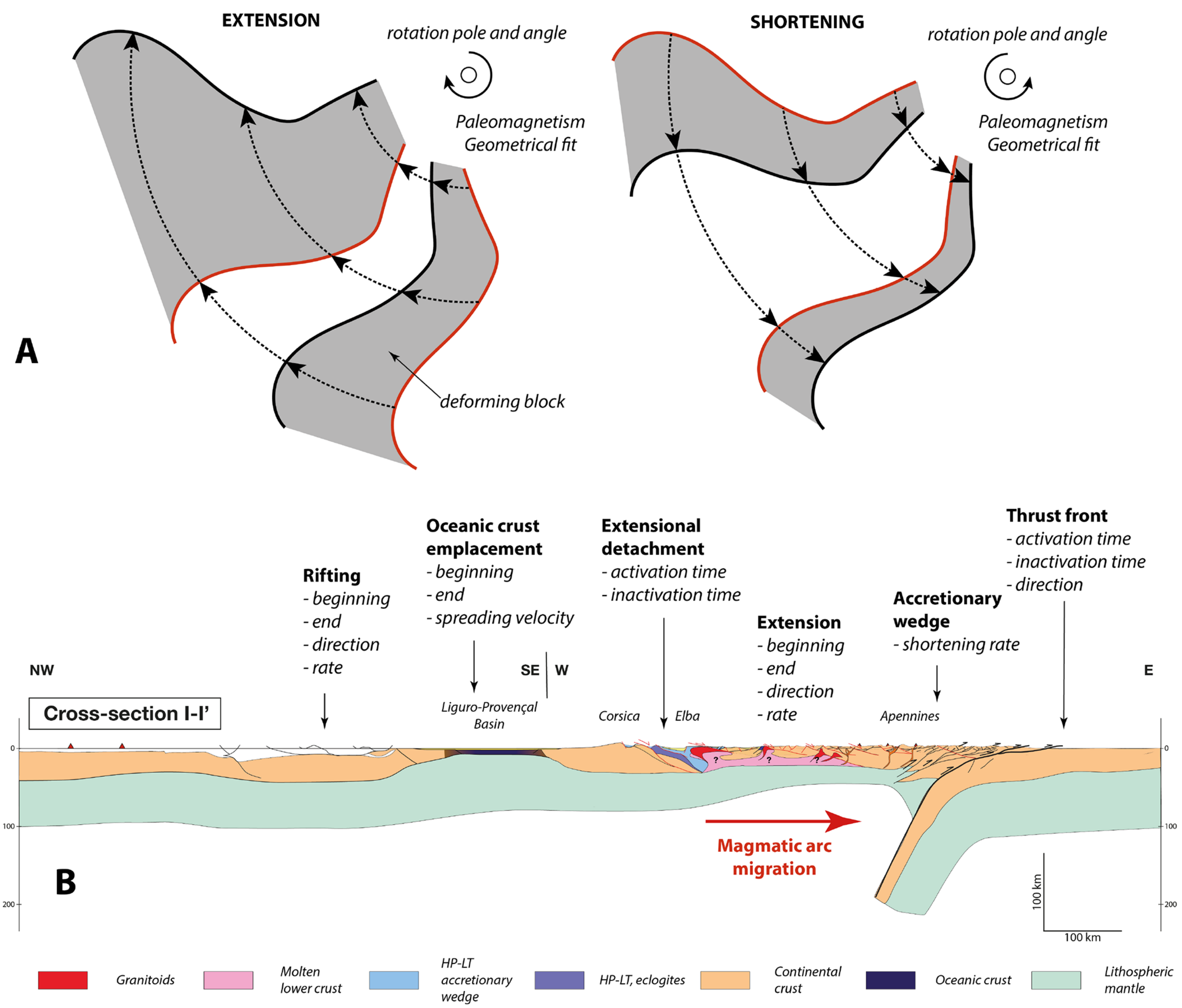
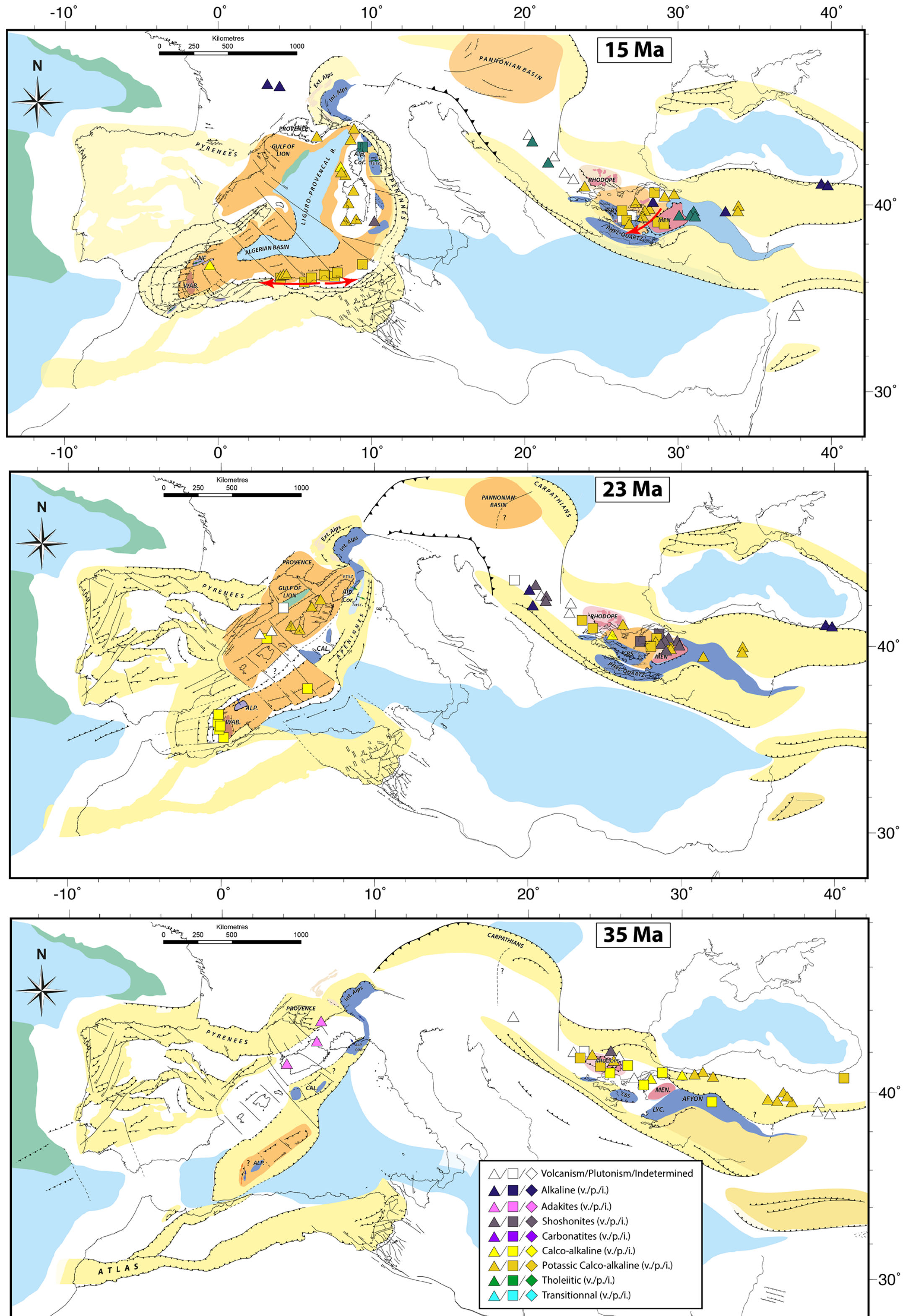


Fig. 3. Schematic presentation of the method used for reconstructions. (A) The various constraints used in the reconstructions along a cross-section from Provence to the Apennines (see location on Fig. 1). (B) Methodology to account for internal deformation of blocks during their displacement.

is as independent as possible of any dynamic considerations and can then be used to discuss dynamic processes.

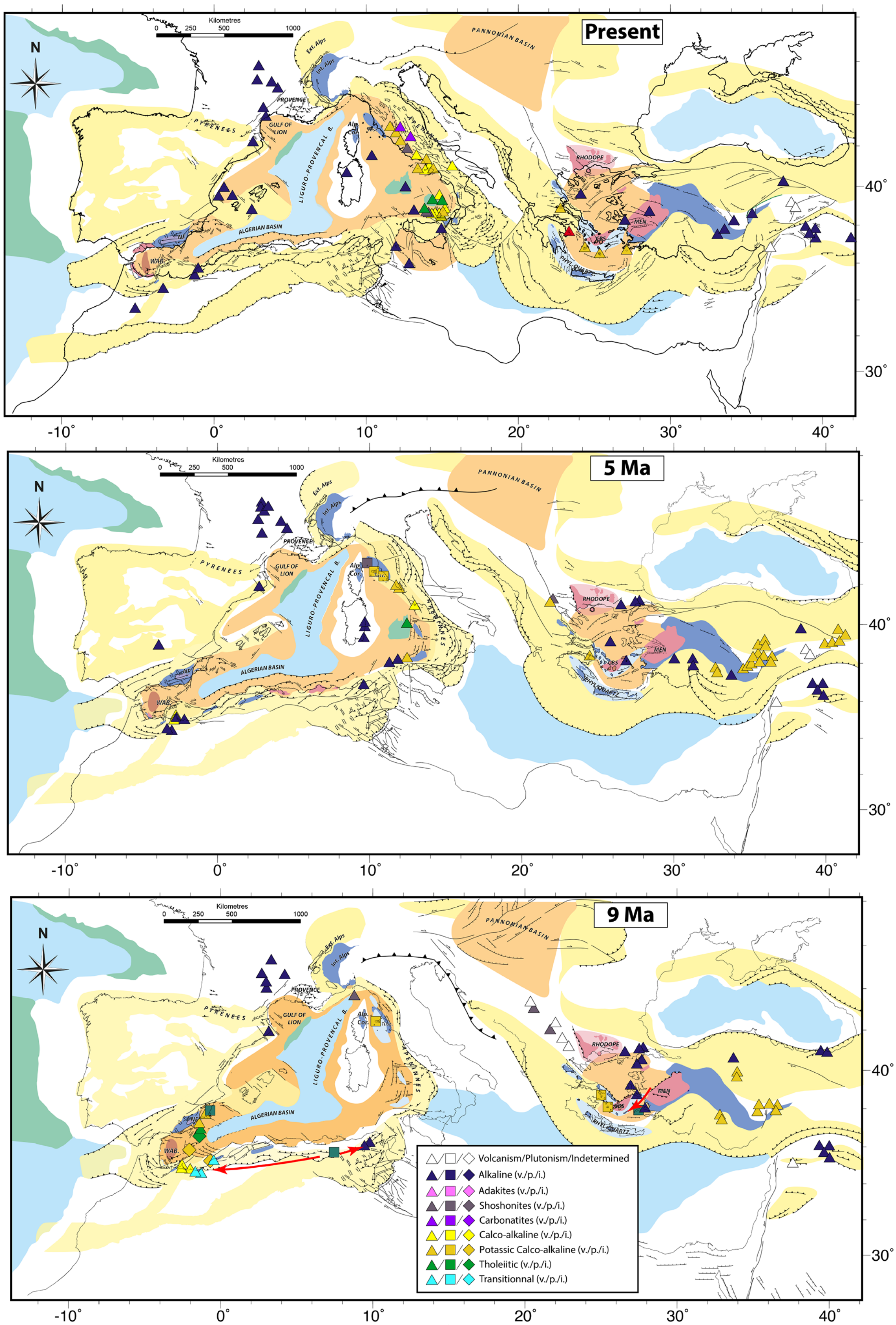
Detailed geological cross-sections across mountain belts, their history of deformation (timing, shortening directions) and evolution of their foredeeps were used to constrain shortening at the periphery of the deforming region: Jura and Alps (Giannerini, 1980-1981; Sommaruga, 1997; Burkhard and Sommaruga, 1998; Lickorish and Ford, 1998; Scharf *et al.*, 2013; Jourdon *et al.*, 2014), Apennines (Pauselli *et al.*, 2006; Boccaletti *et al.*, 2011), Rif and Atlas (Gomez *et al.*, 1998; Zarki *et al.*, 2004; Benaouali-Mebarek *et al.*, 2006; Chalouan *et al.*, 2006; El Kadiri *et al.*, 2006; Chalouan *et al.*, 2008; Di Staso *et al.*, 2010; Vitale *et al.*, 2014; Capella *et al.*, 2017), Gibraltar arc (Rosenbaum and Lister, 2004; Crespo-Blanc and Frizon de Lamotte, 2006; Lujan *et al.*, 2006; Pedrera *et al.*, 2014; Sanz de Galdeano *et al.*, 2015; Crespo-Blanc *et al.*, 2016), Calabria and Peloritani (Bonardi *et al.*, 2001; Heymes *et al.*, 2010; Cirrincione *et al.*, 2015), Baleares (Sàbat *et al.*, 1997; Etheve *et al.*, 2016), shortening within Iberia (Vergés and Sàbat, 1999; Quintana *et al.*, 2015), and Pyrenees (Muñoz, 1992; Meigs *et al.*, 1996; Muñoz, 2002; Mouthereau *et al.*, 2014; Teixell *et al.*, 2018).

The tectonic history of back-arc basins was also a major constraint with in particular exhumation kinematics and *P-T*-time evolution of MCCs, paleostress studies and structure of passive margins: Tyrrhenian (Cocchi *et al.*, 2009), Corsica (Jolivet *et al.*, 1990; Fournier *et al.*, 1991; Jolivet *et al.*, 1998; Brunet *et al.*, 2000; Molli *et al.*, 2006; Vitale Brovarone and Herwartz, 2013; Beaudoin *et al.*, 2017), Tuscan archipelago (Jolivet *et al.*, 1998), Valencia Basin (Roca *et al.*, 1999; Etheve *et al.*, 2016; Etheve *et al.*, 2018), Gulf of Lion (Bache *et al.*, 2010; Jolivet *et al.*, 2015a; Moulin *et al.*, 2015), Alboran region (Jabaloy *et al.*, 1993; Platt *et al.*, 2003a, 2003b; Booth-Rea *et al.*, 2004, 2005, 2015; Augier *et al.*, 2005a, 2005b;



(a)

Fig. 4. Reconstructions of the Mediterranean domain since 35 Ma. Colored symbols represent magmatic bodies (plutonic, volcanic and unspecified) and their magmatic series. The details of the Western Mediterranean is shown in Figure 5. For details on the Aegean-Anatolia region the reader is referred to Menant *et al.* (2016a, 2016b). Alps and Carpathians not precisely reconstructed, only indicative. Light blue: oceanic crust; green: exhumed mantle; orange: thinned continental crust; pink: high-temperature metamorphic core complexes; deep blue: high-pressure and low-temperature metamorphic complexes; yellow: areas with active shortening.



(b)

Fig. 4. Continued

Jolivet et al., 2006, 2008; Martinez-Martinez et al., 2006; Iribarren et al., 2007; Mauffret et al., 2007; Augier et al., 2013; Do Couto et al., 2014, 2016).

Reconstructions were built progressively backward in time and are represented at key time steps, 5 Ma, 9 Ma, 15 Ma, 23 Ma, 28 Ma and 35 Ma (Figs. 4 and 5). Reconstructions of the Western Mediterranean were then integrated within a model encompassing the whole Mediterranean with the addition of the reconstructions of the Eastern Mediterranean published by Menant et al. (2016a). Note that the Alps and Carpathians were not reconstructed in this work and are shown only schematically, except for a part of the western Alps. On these reconstructions, the evolution of magmatic production was added with a differentiation of the magmatic series in order to discuss the interaction with geodynamic processes at mantle scale. The reader is referred to the work of Menant et al. (2016a) for the methodology of associating magmatic events to the reconstructions and for the compilation of original data for the Eastern Mediterranean. For the Western Mediterranean, the magmatic events were taken from the detailed compilations of magmatic events in the whole Western Mediterranean of Savelli (1988, 2002a, 2002b, 2015). These were completed by Serri et al. (1993) and Avanzinelli et al. (2009) for the evolution of magmatism in the Italian peninsula and periphery of the Tyrrhenian Sea. Ages of the Malaga dykes are from Esteban et al. (2013). The age of granitic bodies around Beni Bousera was taken from Rossetti et al. (2010). Réhault et al. (2012) detailed the Late-Eocene to Early Miocene magmatism in the northern part of the Liguro-Provençal Basin and more recent data were acquired by Lustrino et al. (2017). Maury et al. (2000), Coulon et al. (2002) and Abbassene et al. (2016) presented the evolution of magmatism along the northern margin of Africa and its relation to slab tearing. Duggen et al. (2004, 2005, 2008, 2009) detailed the evolution of magmatism in the westernmost Mediterranean in the Atlas, Rif and Alboran region, discussing the respective roles of subduction and hot spot volcanism. The ages of magmatism of the French Massif Central were taken from Nehlig et al. (2003).

5 Tectonic and geodynamic evolution from 35 Ma to the present

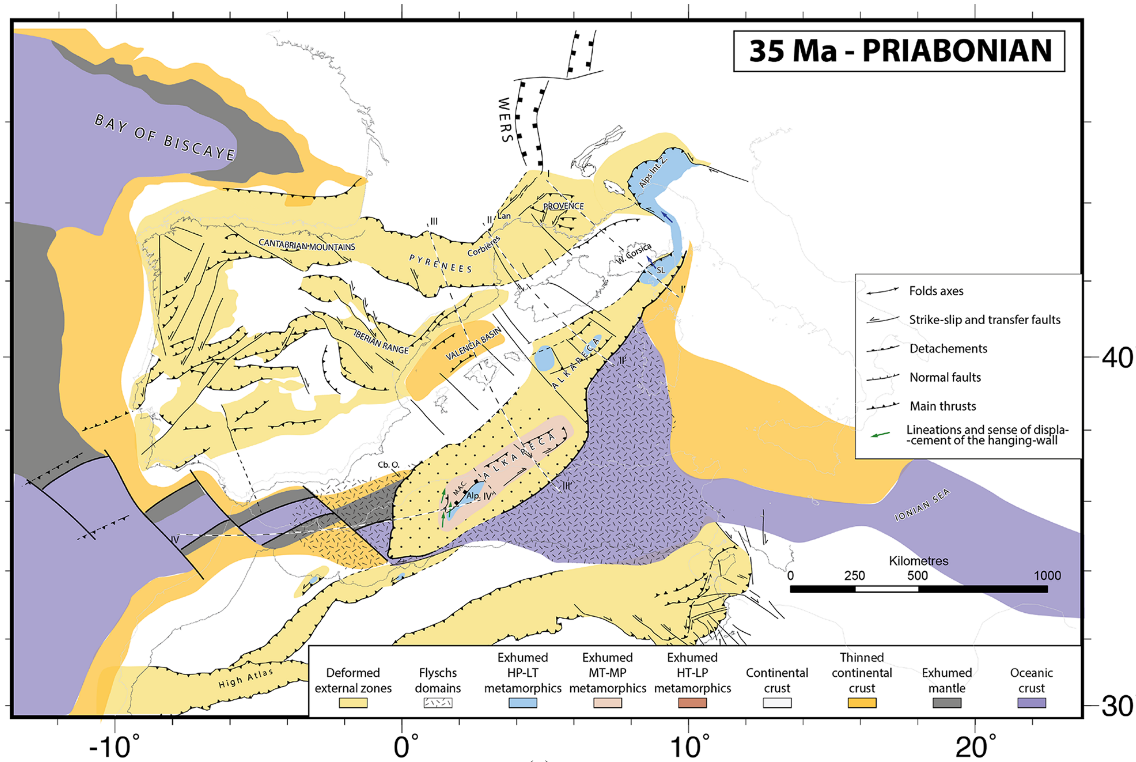
The reconstructed paleo-tectonic maps at key-periods come with lithospheric-scale cross-sections along four major transects (I to IV, Figs. 6A–6D) that represent the main geodynamic processes at work since the end of the Eocene; their locations are shown in Figure 1. In the following, we describe these key-periods and insist on the transitions in between. The chosen time-steps are just snapshots in a continuous evolution and the transition between these is equally important.

5.1 35–28 Ma, from late Eocene to middle Oligocene (Figs. 5a and 5b)

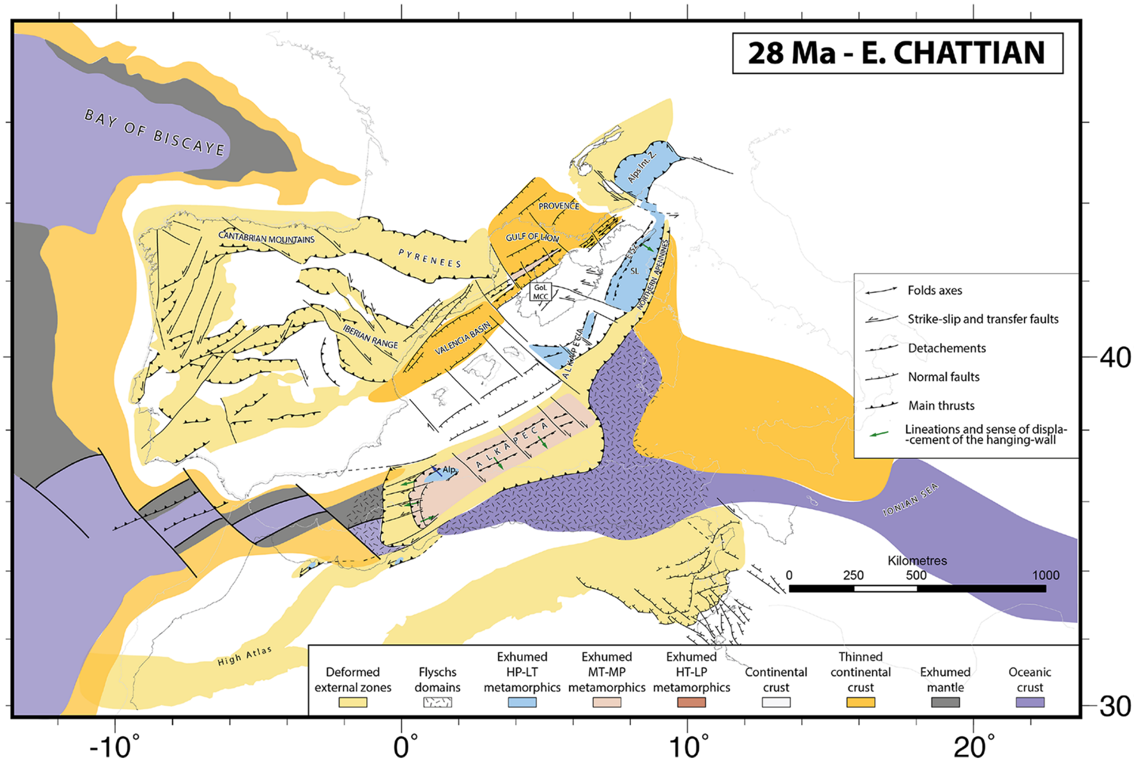
The earliest stage of our reconstructions (35 Ma) corresponds to the Priabonian. The backward step-by-step reconstruction leads to a geometry of subduction zones where the Mesozoic Tethyan lithosphere is consumed in two different trenches. The longest one runs from the zone of transition from the Alps to the future Apennines all the way to the southeast of the Baleares. The Tethyan oceanic lithosphere sinks northwestward underneath Corsica, Sardinia and the AlKaPeCa block. The second subduction zone has an opposite vergence. The subducting plate dips underneath the westernmost part of AlKaPeCa along a NS-trending trench in the Alboran domain. The trench probably extends further toward the northeast but its possible connection with the then-closed Ligurian Ocean as in Michard et al. (2002) and Chalouan and Michard (2004) is uncertain. This subduction zone has consumed the partly oceanized basin preserved in the Nevado-Filábride complex in the Bédar Macael unit (Puga et al., 2017).

The oceanic domain between Iberia and Africa is represented on the reconstructions based on the hypotheses proposed by Frizon de Lamotte et al. (2011), Vergés and Fernàndez (2012), Leprêtre et al. (2018), Fernàndez et al. (2019) or Pedrera et al. (2020). Because the relative motion during the Jurassic and the early Cretaceous between Africa and Eurasia was highly oblique with a left-lateral component, these authors show a series of short spreading centers, partly oceanized and with exhumed mantle in the ocean-continent transition and separated by transform faults. This domain is now partly included in the Betic nappe stack. Portions with partly oceanized crust or exhumed mantle are found in the Bédar-Macaël unit of the Nevado-Filábride complex or the Ronda and Beni Bousera peridotite massifs on the Alpujárride complex. It is progressively subducted and partly accreted through the subsequent stages of the reconstructions.

The Priabonian sees a major transition in the Pyrenees and Languedoc (Séranne, 1999; Grool et al., 2018), characterized by the last thrusting episodes along the northern Pyrenean thrust front that is reworked by left-lateral strike-slip faults and associated basins between the Pyrenees and Provence fold-and-thrust belt. These strike-slip faults connect further north with the West European Rift System (WERS) (Séranne, 1999). We assume that the overall crustal-scale structure seen on seismic profiles of the Pyrenees today (Chevrot et al., 2018) was to the first order established at the end of the Eocene, despite some more shortening until the Late Oligocene – Early Miocene in the South Pyrenean Zone (Muñoz et al., 2018). We also assume that this structure was present in the eastern Pyrenees at the time the Gulf of Lion passive margin began to form (see cross-section II-II' and section III-III' in Fig. 6B). During the Oligocene, rifting in the Gulf of Lion led to the collapse of the eastern part of the Pyrenees. Jolivet et al. (2020) proposed that this rifting episode involved extraction of upper mantle and lower crust to form the Gulf of Lion's metamorphic core complex. During this period, Corsica belongs to an active compressional orogen sandwiched between the future Apennines subduction, then dipping toward the NW, and the Provence fold-and-thrust belt (Vially and Tremolières, 1996; Lacombe and Jolivet, 2005) (see cross-section I-I' on Fig. 6A). The alternative solution of a flip of subduction from a south-dipping subduction in Provence to a north-dipping subduction south of Corsica and Sardinia as proposed by Schettino and Turco (2006) would make slab retreat almost coeval with subduction initiation, which is not likely. The Paleozoic continental basement of Western Corsica overthrusts the deformed cover of Provence, shortened by a series of north- and south-vergent thrusts, such as in the Sainte Baume and Sainte Victoire massifs (Le Pichon et al., 2010; Oudet et al.,

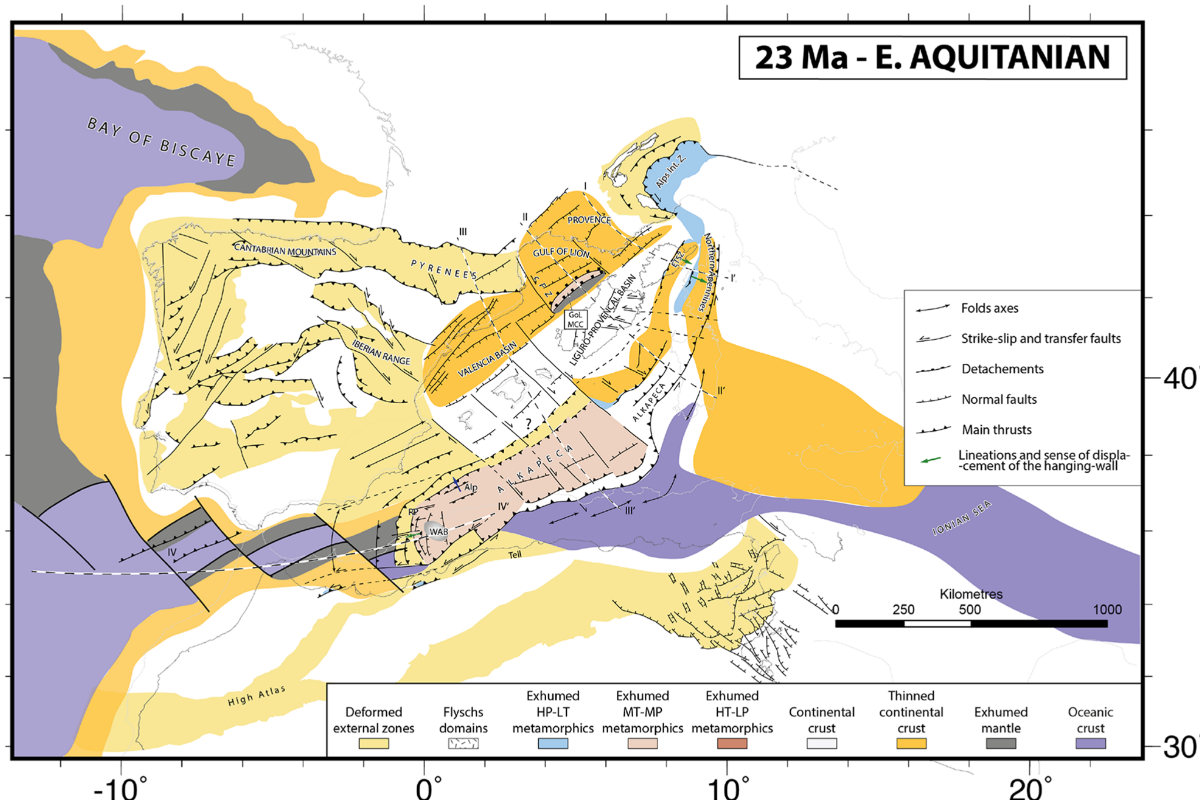


(a)

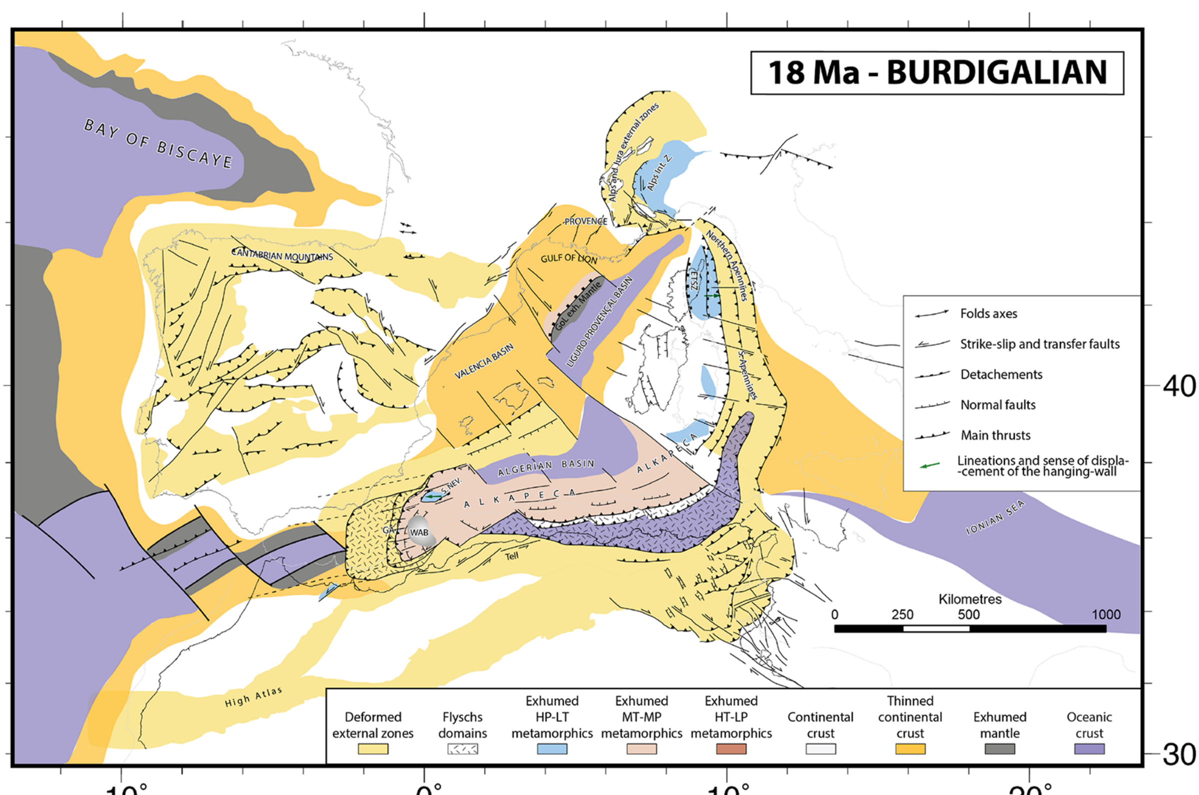


(b)

Fig. 5. Reconstructions in map view at selected periods since the end of the Eocene. 5a: Priabonian (35 Ma), 5b: Late Rupelian-early Chattian (28 Ma); 5c: Early Aquitanian (23 Ma, 5d: Burdigalian (18 Ma), 5e: Langhian (15 Ma), 5f: Tortonian (9 Ma), 5 g: Zanclean (5 Ma), 5h: Present-day. Alp: Alpujárride, Cb. O.: Cobdar Ocean, Elb: Elba, ETSZ: East Tenda Shear Zone, GB: Granada Basin, Gi: Giglio, GoL MCC: Gulf of Lion Metamorphic Core Complex, HO: Huercal Overa Basin, Lan: Languedoc, MB: Marsili Basin, MC: Monte Cristo, NF: Nevado-Filábride, RP: Ronda peridotite, SA: Sierra Alhamilla, SG: Sierra de Gador, SL: Schistes Lustrés, SN: Sierra Nevada, STB: Sorbas-Tabernas Basin, VB: Vavilov Basin, WAB: West Alboran Basin. I-I', II-II', III-III' and IV-IV' are the traces of the cross-sections of Figure 6, also positioned on the map of Figure 1 for the present-day situation.

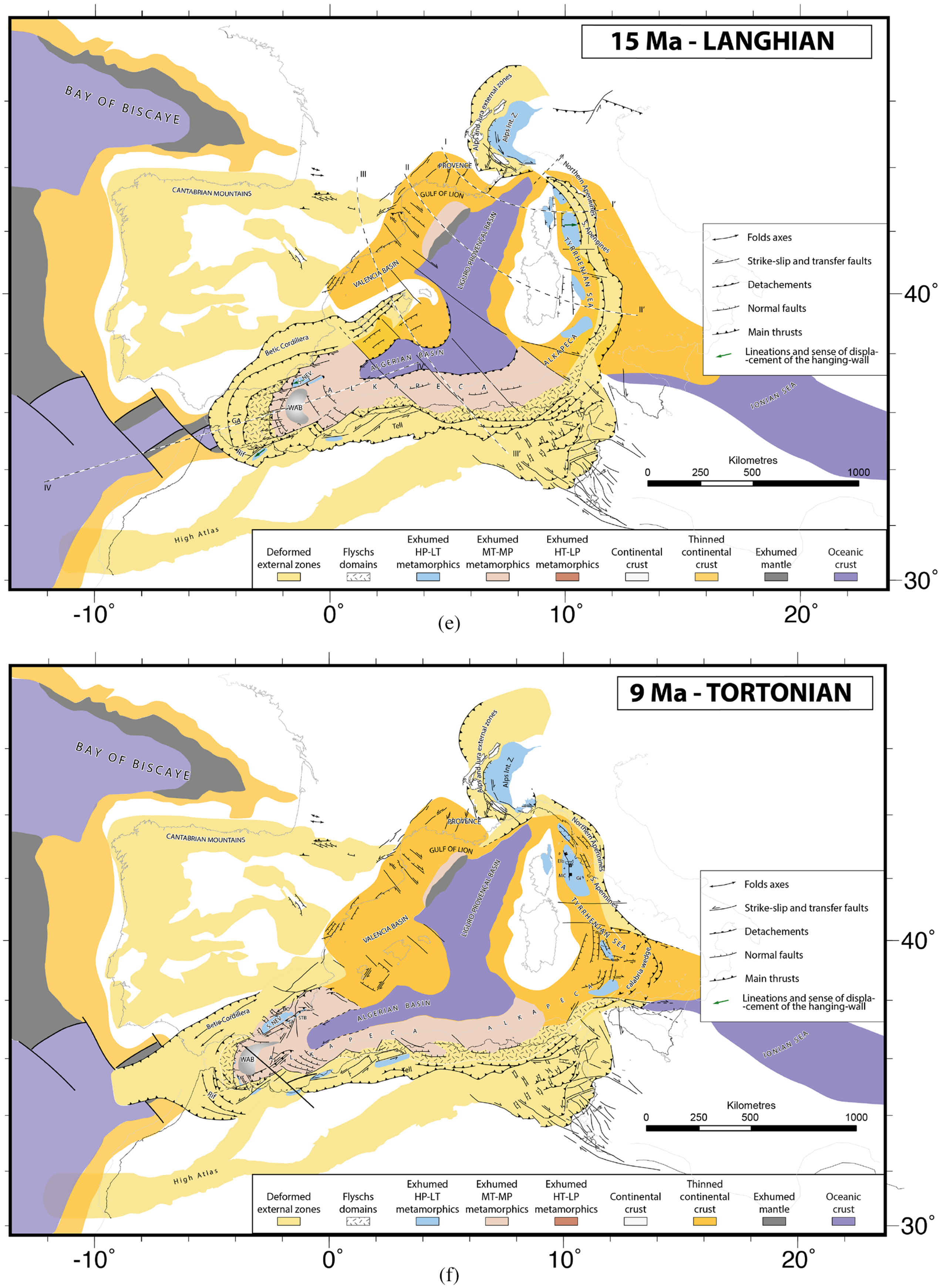


(c)



(d)

Fig. 5. Continued



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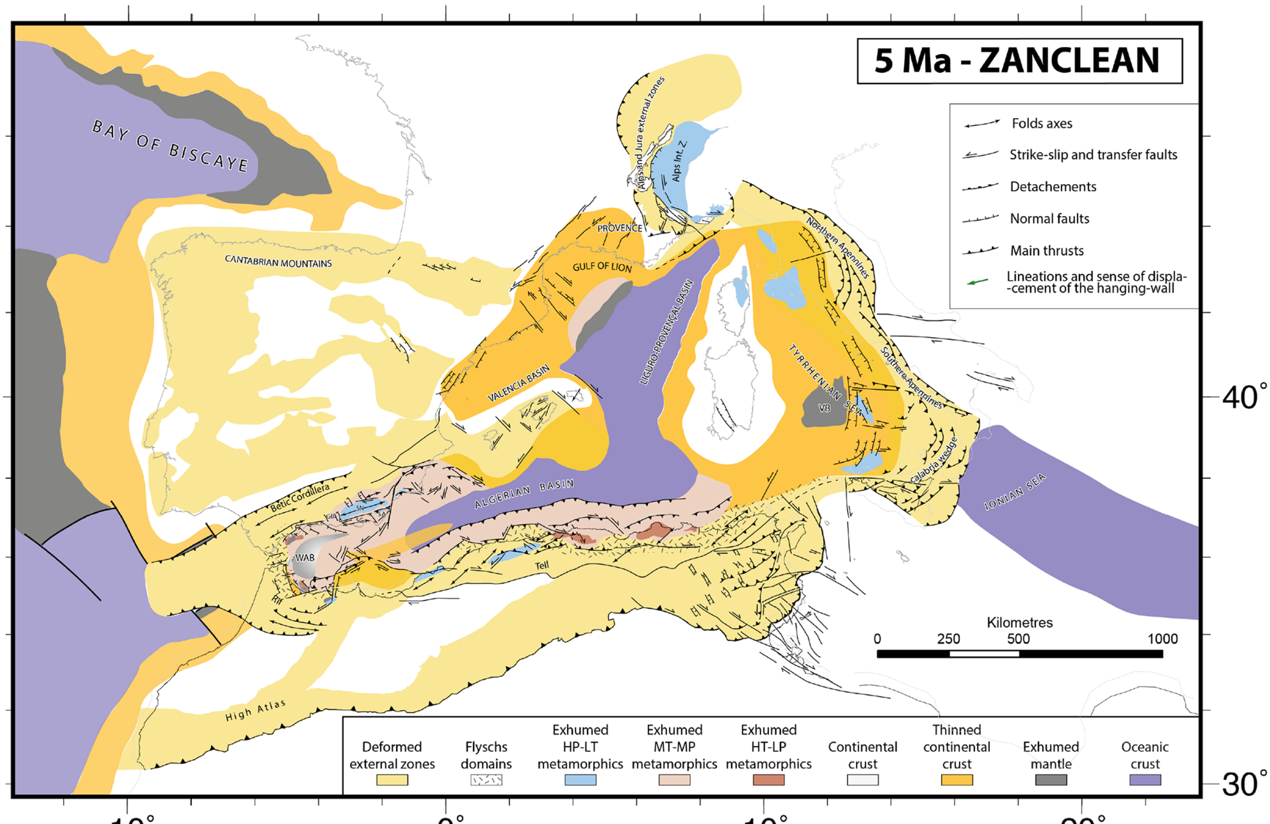

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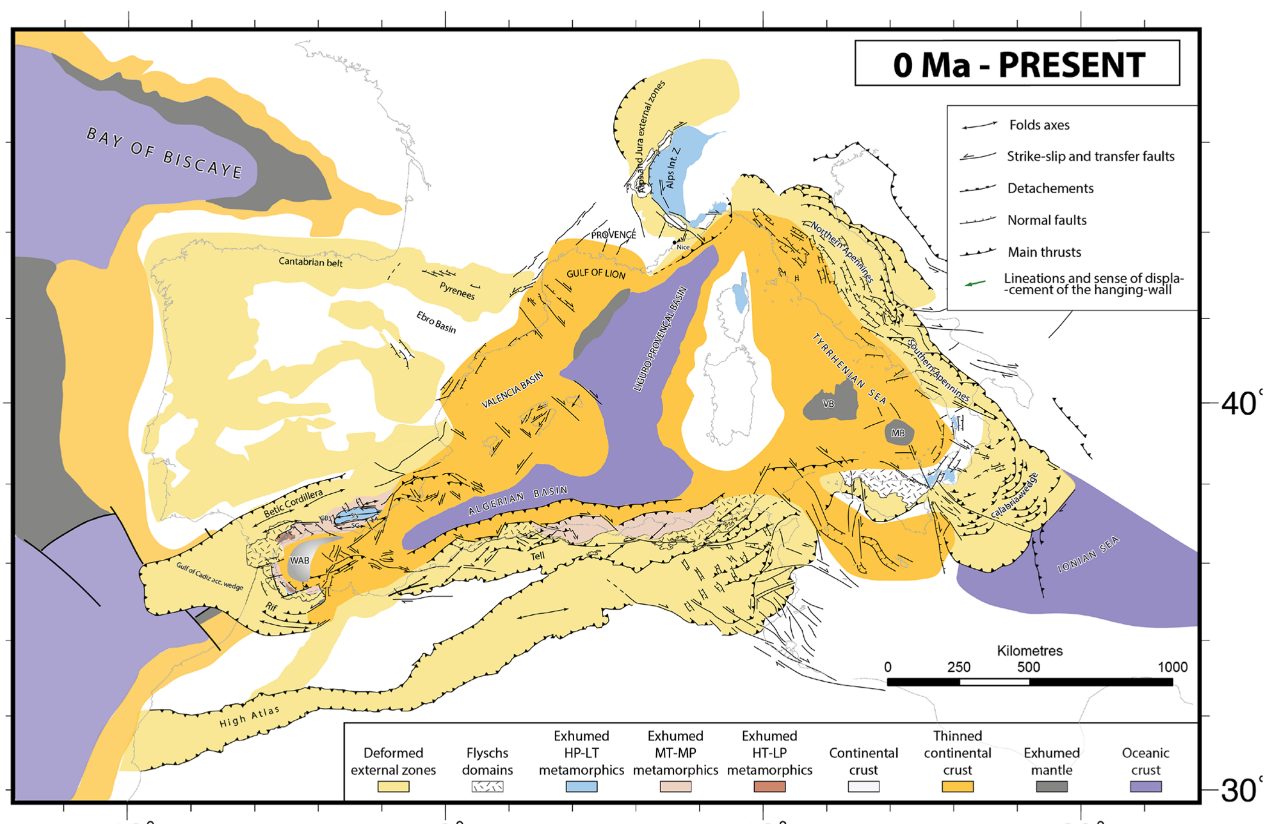


(h)

Fig. 5. Continued.

2010; Rangin et al., 2010; Espurt et al., 2012; Fournier et al., 2016). The H*P*-L*T* metamorphic rocks of Alpine Corsica are the southernmost extension of the Ligurian Schistes Lustrés Nappe of the Alps (Caron et al., 1981; Faure and Malavieille, 1981; Mattauer et al., 1981; Malavieille, 1983; Gibbons and Warburton, 1986; Warburton, 1986; Fournier et al., 1991; Malavieille et al., 1998; Molli et al., 2006; Molli and Malavieille, 2010). They record their last H*P*-L*T* overprint at ~34 Ma like in the Alps (Vitale Brovarone and Herwartz, 2013). Exhumation in Alpine Corsica is mostly accommodated by low-angle detachments that bring non-metamorphosed units directly on top of the H*P*-L*T* Schistes Lustrés Nappe (Jolivet et al., 1990; Fournier et al., 1991; Beaudoin et al., 2017). From 32 Ma onward, the thrust front migrates together with the retreating slab and the tectonic units once belonging to the accretionary wedge are then transferred to the back-arc region and exhumed mostly by extension there (Jolivet et al., 1991; Daniel et al., 1996; Brunet et al., 2000; Maggi et al., 2012; Beaudoin et al., 2017). This situation is typical of the Mediterranean setting and is also observed in the Aegean region, north of the Hellenic trench (Fig. 4) (Jolivet and Brun, 2010). In Figure 6A, we show only the 35 and 23 Ma stages and some shortening was active between 35 and 32 Ma, so that the retreat active between 32 and 23 is not expressed.

Farther south, the AlKaPeca block also records the end of H*P*-L*T* events in the Betic Cordillera (Bessière, 2019). H*P*-L*T* metamorphism in the Alpujárride and Nevado-Filábride complexes results from subduction of portions of the continental margin of Iberia and of the AlKaPeCa block. The width of the accretionary wedge is not known. We consider that the peak of pressure is attained around 40 Ma in the Nevado-Filábride complex and the Alpujárride complex (Platt et al., 2005; Li and Massonne, 2018; Bessière, 2019). In the Alpujárride complex available ages are dispersed with concentration around 20 Ma but the oldest ages are Eocene, around 40 Ma (Platt et al., 2005; Bessière, 2019). In the underlying Nevado-Filábride complex, the age of the peak pressure is debated. Augier et al. (2005a) used the $^{40}Ar/^{39}Ar$ method on white micas crystallized along the decompression path from the peak of pressure in the *P-T* field of the eclogite facies to the greenschist facies and they show a systematic decrease of ages from about 30 Ma to about 14 Ma. They concluded that the peak of pressure was attained in the Eocene. López Sánchez-Vizcaíno et al. (2001) and Platt et al. (2006) instead obtained younger middle Miocene ages using the U/Pb method on zircons and the Lu/Hf on garnets that they attributed to the maximum pressure. More recently, Li and Massonne (2018) dated the peak of pressure in the Bédar-Macael unit of oceanic or hyper-extension affinity, based on the U/Pb method on monazite, to ~40 Ma, which better fits the results obtained by Augier et al. (2005a). We thus consider the young ages obtained by López Sánchez-Vizcaíno et al. (2001) and Platt et al. (2006) as exhumation ages (see Bessière, 2019 for a more detailed discussion). Exhumation-related shearing in the Alpujárride complex is N-S or NE-SW with a top-NE sense of shear and the peak of pressure dates back to 38 Ma, according to Bessière (2019). Exhumation is accommodated by extension and detachments such as the Malaguide-Alpujárride Contact (Lonergan and Platt, 1995) (MAC in Fig. 5A), a large detachment marking the contact with the Malaguide complex above (Platt and Vissers, 1989; Augier et al., 2005a; Platt et al., 2013; Williams and Platt, 2018) and series of detachments inside the nappe stack of the Alpujárride Complex (Azañón et al., 1997; Azañon and Crespo-Blanc, 2000). The direction of extension during this period is N-S or NE-SW and most of the deformation coeval with exhumation is non-coaxial with a top-to the north asymmetry (Jolivet et al., 2008; Augier et al., 2013). The position of the AlKaPeCa block in the reconstructions at 35 Ma is the result of removing the extension that has occurred since that period to return to a normal thickness of 30 km. The finite displacement is thus a minimum and the initial position at 35 Ma could then be farther to the NE, a situation that would not be much different from that proposed in van Hinsbergen et al. (2014).

The AlKaPeCa block has not yet collided with the northern margin of Africa, but the space left in between is very narrow (Leprêtre et al., 2014). Compression is also active during this period in the Atlas that records a first period of shortening along its entire length from Morocco to Tunisia (Chalouan et al., 2008; Frizon de Lamotte et al., 2008, 2011; Leprêtre et al., 2018). The beginning of this period is thus characterized by a generalized compressional regime active from the Atlas all the way to the Pyrenean-Cantabrian belt (Jolivet et al., 2016). Then, from 32 Ma onward, the subduction regime changes totally and slab retreat starts (Jolivet and Faccenna, 2000). Shortening however continues in the Southern Pyrenees with the propagation of the southern thrust front until the early Miocene (Muñoz, 2002; Jolivet et al., 2007; Labaume et al., 2016; Muñoz et al., 2018) while the eastern Pyrenees record a large-scale episode of exhumation and uplift coeval with the inception of rifting in the Gulf of Lion around 30 Ma (Morris et al., 1998; Fitzgerald et al., 1999). This change of subduction regime is associated with the development of a discontinuous magmatic arc with calc-alkaline affinity including plutonic bodies with adakitic characteristics such as the Esterellite in Provence (Réhault et al., 2012) (Fig. 4A). The first subduction-related magmatism develops in Sardinia as soon as 38 Ma (Lustrino et al., 2009).

5.2 28–23 Ma, Oligocene, Late Rupelian-Early Chattian to Oligocene-Miocene boundary (Figs. 5B and 5C)

The early Oligocene corresponds to the first stage of back-arc extension in the whole Mediterranean around 30–32 Ma (Jolivet and Faccenna, 2000), which is active everywhere above the SW-NE striking subduction zone consuming the remaining Ionian oceanic lithosphere. Rifting starts to the northwest of Corsica and Sardinia with low-angle detachments shaping the Gulf of Lion passive margin (Jolivet et al., 2015a, 2019). Extension with southeast-dipping low-angle detachments is also active in Alpine Corsica, exhuming the H*P*-L*T* metamorphic units of the Schistes Lustrés (East Tenda Shear Zone) and Tenda Massif (Jolivet et al., 1990, 1991; Fournier et al., 1991; Beaudoin et al., 2017). During the course of slab retreat, the accretionary wedge in Corsica retreats eastward, followed by the back-arc extensional domain that is instead widening. The distance between the trench and the magmatic arc decreases with time, an indication of the increasing steepness of the slab explaining the narrowness of the wedge (Figs. 6A and 6B) (Jolivet et al., 1998; Brunet et al., 2000). Extension is also active in Sardinia with the formation of tilted blocks controlled by northwest-dipping normal faults (Cherchi and Montadert, 1982; Casula *et al.*, 2001). N-S extension exhumes H*P*-L*T* metamorphic units of the Alpujárride in the future Betic-Rif orogen (Azañón *et al.*, 1997; Martinez-Martinez and Azañon, 1997; Azañon and Crespo-Blanc, 2000). Coevally, a southward migration of the southern thrust front of the Pyrenees is recorded (Jolivet *et al.*, 2007; Labaume *et al.*, 2016). A calc-alkaline magmatic arc develops from Sardinia (Lustrino *et al.*, 2009) to the northeastern Valencia Basin (Maillard *et al.*, 2020) (Fig. 4A). Evidence of magmatic product then shift toward the south across the transfer zone between Sardinia and the Baleares Islands.

5.3 23–18 Ma, Oligocene-Miocene boundary to Burdigalian (Figs. 5C and 5D)

At 23 Ma, a large oceanic domain is still to be consumed by subduction south of the AlKaPeca terranes. Slab retreat proceeds mostly southward at the longitude of Algeria and Tunisia. Farther west in the future Betic-Rif domain, recorded extension is still mostly N-S in the first part of this period with detachments accommodating the exhumation of the H*P*-L*T* Alpujárride complex in the future eastern Betics and part of the Rif (Crespo-Blanc *et al.*, 1994; Negro *et al.*, 2005; Jolivet *et al.*, 2008).

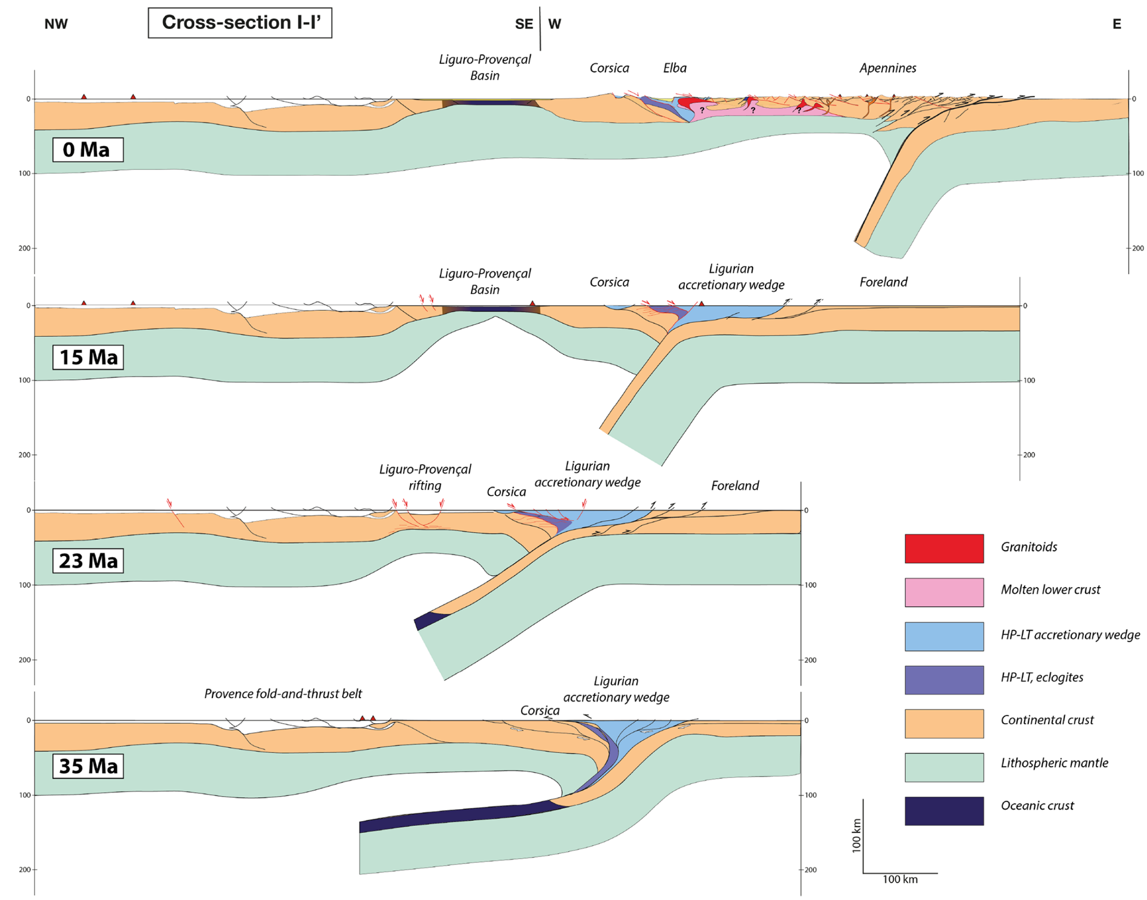
In the west of the Betics, H*P*-L*T* metamorphic rocks are exhumed below the MAC detachment and above the west-directed basal thrust of the Ronda peridotite and Dorsale Calcaire (Platt *et al.*, 1998, 2003b; Mazzoli and Martin-Algarra, 2011; Mazzoli *et al.*, 2013; Précigout *et al.*, 2013; Frasca *et al.*, 2016; Gueydan *et al.*, 2019). The nature of the domain subducting below the Alpujárride units is partly unknown. It could be partly oceanic and partly a highly thinned continental crust.

At this period, a high-temperature metamorphic event and a fast exhumation episode are recorded in the Western Betics and the Rif, attested by a clustering of radiometric ages between 23 and 19 Ma with a variety of geochronological systems (Loomis, 1975; Priem *et al.*, 1979; Zeck *et al.*, 1989, 1992; Monié *et al.*, 1994; Sosson *et al.*, 1998; Platt and Whitehouse, 1999; Sànchez-Rodríguez and Gebauer, 2000; Zeck and Williams, 2001; Platt *et al.*, 2003a, 2003b, 2005; Whitehouse and Platt, 2003; Esteban *et al.*, 2004, 2011; Michard *et al.*, 2006; Rossetti *et al.*, 2010; Frasca *et al.*, 2017; Homonnay *et al.*, 2018; Li and Massonne, 2018; Bessière, 2019; Gómez-Pugnaire *et al.*, 2019). It is recorded in all individual nappes of the Alpujárride, from east to west (Esteban *et al.*, 2004; Platt *et al.*, 2005; Li and Massonne, 2018; Bessière, 2019). The significance of this clustering of ages around 20 Ma is still under discussion. It is classically interpreted as a resetting of all chronometers due to a thermal pulse related to an episode of slab detachment or slab tearing coeval with the overthrusting of the Alpujárride complex, including the Ronda peridotite, onto the continental margin of Iberia (Mazzoli and Martin-Algarra, 2011; Mazzoli *et al.*, 2013; Gueydan *et al.*, 2019). It could indeed be caused by the HT-LP event in the western Betics, but this event is not strong enough to fully overprint the former HP-LT parageneses in the eastern Betics where they are quite well preserved (Bessière 2019). Alternatively, it could correspond to an exhumation event synchronously recorded in the whole Betic domain (Bessière 2019). The coeval fast exhumation of all units still buried make them cross the closure temperature of all geochronological systems.

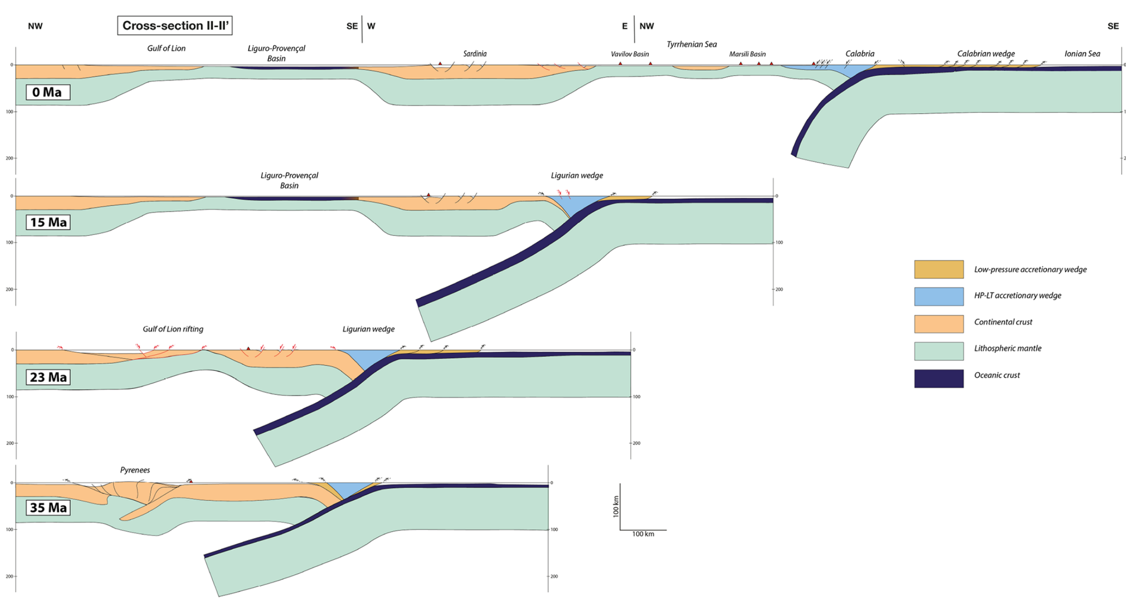
In the Gulf of Lion, this period is that of rifting and exhumation with formation of a MCC made of lower crust, below northwest-dipping detachments east of the Catalan transfer zone (Jolivet *et al.*, 2015a; Canva *et al.*, 2020; Maillard *et al.*, 2020) (Figs. 5C and 6B). Fast rotation of Corsica and Sardinia follows, coeval with the emplacement of oceanic crust or exhumed mantle in the Liguro-Provençal basin (Vigliotti and Kent, 1990; Gattacceca, 2000; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007; Maffione *et al.*, 2008; Dannowski *et al.*, 2020). The previously formed Pyrenean chain is rapidly collapsing in this region and extensional structures rework the Provence fold-and-thrust belt (Gorini *et al.*, 1993; Séranne, 1999). In the southern Pyrenees, the last south-vergent thrusts propagate in the Ebro basin until about 20 Ma (Muñoz, 2002; Vergés *et al.*, 2002; Jolivet *et al.*, 2007; Bosch *et al.*, 2016; Labaume *et al.*, 2016). Slow extension is active in the Valencia Basin associated with intense magmatism, especially in the transition zone with the Gulf of Lion through a series of NW-SE striking dextral transfer faults (Maillard and Mauffret, 1999; Maillard *et al.*, 2020). Magmatism is still active in the Sardinia volcanic arc and northward from offshore Corsica to the Ligurian Sea (Lustrino *et al.*, 2009; Réhault *et al.*, 2012).

5.4 18–15 Ma, Miocene, Burdigalian to Langhian (Figs. 5D and 5E)

In the Burdigalian, an oceanic space is still open between the southward migrating AlKaPeCa block and the northern African margin, but an accretionary wedge is developing at the expense of the Flysch units that will then overthrust the margin (Leprêtre *et al.*, 2014). The same geometry can be found farther west, west of the retreating Gibraltar arc (Crespo-Blanc and Frizon de Lamotte, 2006). In the overriding domain, back-arc extension is active with the continuing opening of the Liguro-Provençal basin and the Algerian basin. The oceanic crust of the Algerian basin progresses westward, following the migration of the Alboran domain, which width increases because of back-arc extension. The opening is associated with NW-SE trending transfer faults that accommodate the differential rotation of the Corsica-Sardinia block on the one hand and the AlKaPeca block on the other hand (Mauffret *et al.*, 1995; Maillard and Mauffret, 1999; Maillard *et al.*, 2020). Emplacement of oceanic crust follows hyper-extension of the Gulf of Lion margin and exhumation of mantle and lower crust of the Gulf of Lion MCC below northwest-dipping detachments (Jolivet *et al.*, 2015a; Moulin *et al.*, 2015). In the Liguro-Provençal and Algerian basins, oceanic crust is emplaced but no distinct ridge, nor clear magnetic anomaly pattern can be recognized (Rollet *et al.*, 2002). Recent surveys suggest that the crust once attributed to oceanic accretion is in fact exhumed mantle with localized magmatism (Dannowski *et al.*, 2020). Extension is active in Alpine Corsica with east-dipping detachments, completing the exhumation of the HP-LT metamorphic units of the Schistes Lustrés (Jolivet *et al.*, 1991; Daniel *et al.*, 1996). One of these detachments is the East Tenda Shear Zone, active from ~ 30 Ma to 21 Ma in the ductile

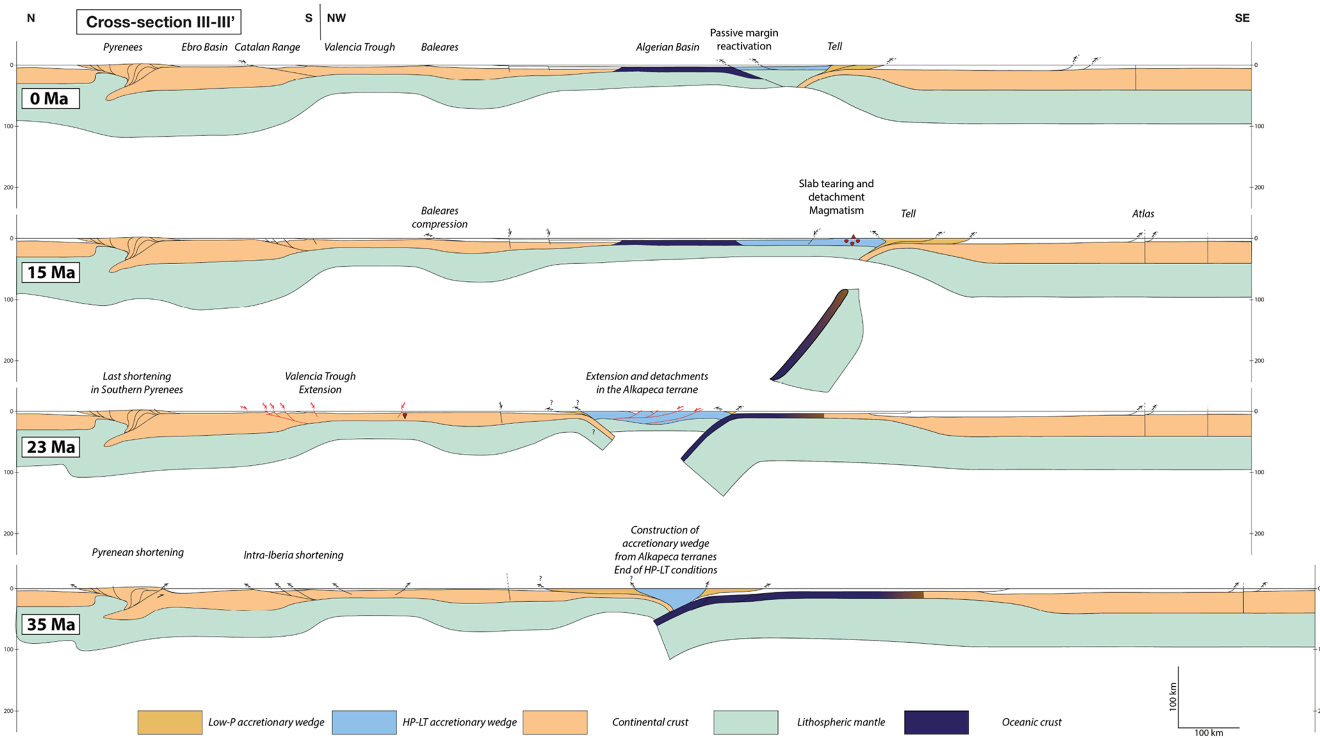


(a)

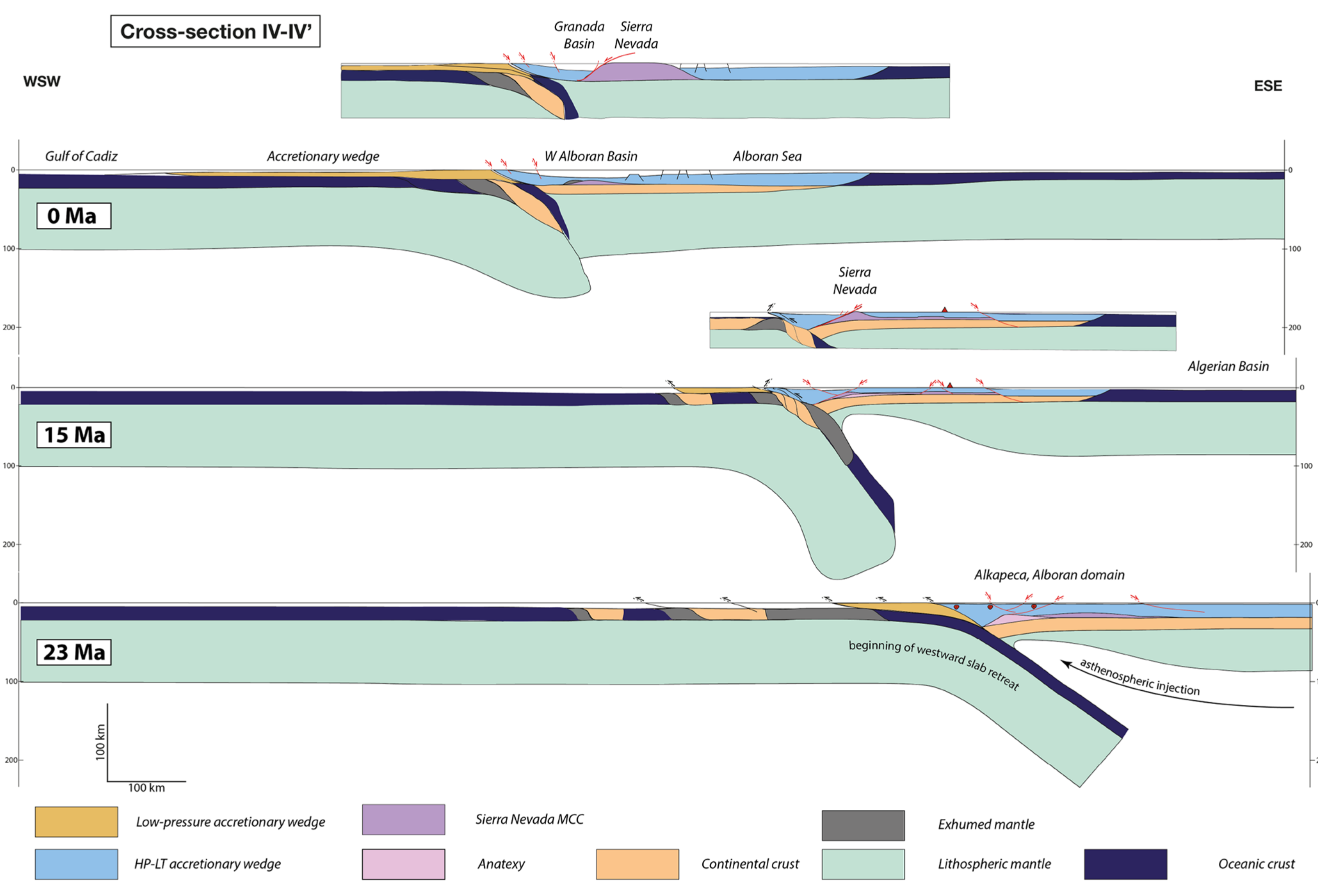


(b)

Fig. 6. Reconstructed cross-sections at selected periods. See cross-sections location on Figure 1. The structure of the Pyrenees is adapted from Chevrot *et al.* (2018). (A) Section across the Provence fold-and-thrust belt, the Liguro-Provençal basin, Corsica, the Tyrrhenian Sea and the Tuscan Archipelago and, finally the Apennines. (B) Section across the Gulf of Lion passive margin, the Liguro-Provençal Basin, Sardinia, The Southern Tyrrhenian Sea, Calabria and the Calabrian subduction zone. 6c: section across the Pyrenees, the Catalan Range, the Valencia Trough, the Balearic Islands, the Algerian Basin, the north African margin of Algeria and the Tell. 6d: section across the Gulf of cadiz accretionary wedge, the Gibraltar Arc, the Alboran Sea. Shorter sections show a parallel E-W onland line across the Betic Cordillera.



(c)



(d)

Fig. 6. Continued.

field and in brittle conditions afterward (Daniel *et al.*, 1996; Rossetti *et al.*, 2015; Beaudoin *et al.*, 2017), controlling the deposition of sediments into the asymmetric Saint Florent basin from the Burdigalian onward (Ferrandini *et al.*, 1996; Cavazza *et al.*, 2007). In the Betics and Rif, back-arc extension is also active behind the future Gibraltar arc thrust front. The exhumation of the Nevado-Filábride Units in the Sierra Nevada-Sierra de los Filabres MCC has started below the Filabres detachment with a top-to-the west kinematics, mostly accommodated by ductile deformation (Jabaloy *et al.*, 1993; Augier *et al.*, 2005c, 2005d). As mentioned above, we consider here that the middle Miocene ages obtained in the Nevado-Filábride complex and interpreted as dating the HP-LT metamorphism in this complex (López Sánchez-Vizcaíno *et al.*, 2001; Platt *et al.*, 2006) as exhumation ages (see a discussion in Bessière *et al.*, this volume).

The 18–20 Ma period corresponds to the age of a change in the direction of extension and exhumation that switches from N-S to E-W, following the westward retreat of the slab subducting below the Betic-Rif orogen (Jolivet *et al.*, 2006, 2008). At about 15 Ma, an almost continuous magmatic arc is active from the Ligurian Sea to Corsica, then Sardinia and along the northern African margin with potassic calc-alkaline affinity (Fig. 4A).

5.5 15–9 Ma, Miocene, Langhian to Tortonian (Figs. 5E and 5F)

15 Ma is a key-date in the evolution of the Western Mediterranean region. It immediately follows the completion of the collision of the AlKaPeCa block, carried by the overriding plate during the southward retreat of the subduction zone, with the northern margin of Africa and complete resorption of the Tethys remnants there, after the end of opening of the Liguro-Provençal basin (Chamot-Rooke *et al.*, 1999; Leprêtre *et al.*, 2014). This collision has been diachronous as shown on the reconstructions but the oceanic space north of the present Tell definitely closes at about 15 Ma. No southward retreat is then possible after this final docking of AlKaPeCa on Africa and the slab only retreats eastward along a transfer fault north of Sicily (Rosenbaum and Lister, 2004) or westward along the northern margin of Africa. Magmatism, essentially granitoids, develops in northern Algeria and Tunisia with potassic calco-alkaline affinities and then migrates westward along the margin to reach the southern margin of the Alboran Sea with transitional affinity at 9 Ma and alkaline affinity afterward (Fig. 4B). This migration has been interpreted as a witness of the progressive tearing of the slab from east to west (Maury *et al.*, 2000; Coulon *et al.*, 2002) supported by tomographic imaging showing a detached slab (Fichtner and Villaseñor, 2015) or no slab (Spakman and Wortel, 2004) below the Algerian margin nowadays. Near the front of subduction, the Betic-Rif accretionary wedge continues to form above the retreating slab at the expense of the passive margins. In the Internal Zones of the Betics, the exhumation of the Sierra Nevada MCC below the Filabres detachment is still active with top-to-the west shear sense across the ductile-to-brittle transition and the last synkinematic metamorphic minerals form (Jabaloy *et al.*, 1993; Augier *et al.*, 2005a). Last increments of motion occur in the brittle field still with a dominant top-to-the-west sense of shear until 12 Ma. At that time, most of the extensional intramountain basins present in the eastern Betics form above extensional detachments with different kinematics and infill dynamics (Meijninger and Vissers, 2006; Augier *et al.*, 2013; Do Couto *et al.*, 2014). The opening of the Algerian Basin proceeds with a westward migration following the westward retreat of the slab below the Gibraltar arc (Mauffret *et al.*, 2004; Driussi *et al.*, 2015). Further east, the retreat of the Apennines slab is stalling for a few Myr after the opening of the Liguro-Provençal basin and before the opening of the Tyrrhenian Sea (Faccenna *et al.*, 2001a, 2001b, 2003).

One puzzling observation remains difficult to explain, the middle Miocene thick-skinned shortening event recorded in Ibiza and Mallorca, in the Balearic islands (Fig. 5E). This short event is well documented in the field and offshore (Sàbat *et al.*, 1997; Roca, 2001; Etheve *et al.*, 2016). It comes after an episode of Oligocene and early Miocene rifting and before a second episode of extension at the end of the middle Miocene with a component of dextral motion along the Emile Baudot escarpment (Driussi *et al.*, 2015; Etheve *et al.*, 2016). The Oligocene-early Miocene extension is coeval with back-arc opening in the Liguro-Provençal basin and the transtensional episode with the opening of the southern Algerian basin and fast migration of the Alboran block toward the west. So, this short compression is recorded in an overall extensional period. Etheve *et al.* (2016) note that it happens at the time of docking of the AlKaPeca block on the north African margin. Compressional stresses may have then been transferred across the oceanic Algerian basin from the Kabylies to the Balearic islands.

5.6 9–5 Ma, Miocene, Tortonian to Zanclean (Figs. 5F and 5G)

The N-S compression recorded in the Pliocene in the Betics was first established in the Late Tortonian, some 8 Ma ago. At 9 Ma, the Sierra Nevada MCC completes its exhumation as recorded by low-*T* thermochronology (Johnson *et al.*, 1997) still controlled by tectonic denudation and extension due to westward slab retreat (Galindo-Zaldívar *et al.*, 1989; Augier *et al.*, 2005a). From ~8 Ma, a drastic change is well recorded in the Eastern Betics where part of intramountain basins (Sorbas, Tabernas, Huercal-Overa) are inverted, and the MCCs of the Sierra de Los-Filabres-Sierra Nevada, Sierra Alhamilla, Sierra de Gador are uplifted with respect to the intervening basins by a series of long wavelength folds (Comas *et al.*, 1999; Meijninger and Vissers, 2006; Iribarren *et al.*, 2009; Augier *et al.*, 2013; Janowski *et al.*, 2017). At 9 Ma, the tight Gibraltar arc seen today was not yet formed and the reconstructions show a N-S arrangement of the various tectonic units seen today around the sharp bend (Fig. 5F) and during this period, the Gibraltar arc progressively reached its present-day curvature (Crespo-Blanc *et al.*, 2016). This is associated with the propagation of the oceanic crust of the Algerian Basin toward the west and continuing E-W extension in the Alboran region and the initiation of the strike-slip regime (Mauffret *et al.*, 2004; Driussi *et al.*, 2015). The Southern Tyrrhenian Sea opens during this period east of Sardinia with the formation of the Vavilov Basin, mostly by mantle exhumation and some localized volcanism (Prada *et al.*, 2014, 2016, 2018), while the thrust front of the Apennines propagates eastward. Extension is also active east of Corsica in the northern Tyrrhenian Sea, with east-dipping detachments exhuming granitoids and MCCs, such as in the islands of Elba, Monte Cristo and Giglio (Keller and Pialli, 1990; Jolivet *et al.*, 1998; Collettini and Holdsworth, 2004). During this period, the Messinian salinity crisis represents a short dessication event just before the return of seawater in the Zanclean (Hsü *et al.*, 1973; 1978; Clauzon *et al.*, 1996; Krijgsman *et al.*, 1999; Jolivet *et al.*, 2006; Lofi *et al.*, 2010; Bache *et al.*, 2011; Garcia-Castellanos and Villaseñor, 2011) that occurred during the rifting of the western Tyrrhenian basin (Lymer *et al.*, 2016). Alkaline magmatism develops in the French Massif Central (Fig. 4B) (Nehlig *et al.*, 2003).

The main change during this period is thus the return to N-S shortening in the Eastern Betics, while E-W extension is still active in the Granada basin and the Gibraltar Arc still retreating westward.

5.7 5–0 Ma, Pliocene, Zanclean to Present (Figs. 5G and 5H)

The main changes between the Early Pliocene and the Present is the fast opening of the Southern Tyrrhenian Sea, namely the Marsili Basin, as well as the onset of shortening along the northern margin of Africa, Sicily and southern margin of France, offshore between the Gulf of Lion and Gulf of Genova (Mauffret *et al.*, 2004; Billi *et al.*, 2011; Maillard and Mauffret, 2013). N-S compression is recorded also during the Pliocene in the High Atlas (Frizon de Lamotte *et al.*, 2011; Lanari *et al.*, 2020a, 2020b), as well as in the Betics where large wavelength folds amplify the domes of the Sierra Nevada, Sierra Alhamilla, Sierra de Gador (Janowski *et al.*, 2017). In this overall N-S compressional context, E-W extension is still active, especially in the Betic Cordillera, west of the Sierra Nevada metamorphic dome, with active normal faults along the boundary between the metamorphic dome and the Granada Basin (Galindo-Zaldivar *et al.*, 1999; Booth-Rea, 2001). This period also sees 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). N-S compression and E-W extension is also associated with active conjugate sinistral and dextral strike-slip faults and compressional ridges in the eastern and central part of the Alboran Sea (Booth-Rea *et al.*, 2003, 2004; Martinez-Garcia *et al.*, 2013; Estrada *et al.*, 2017; Lafosse *et al.*, 2018; d'Acremont *et al.*, 2020). This N-S shortening was progressively established over the whole Western and Central Mediterranean triggering today the compressional earthquakes and active faults mapped offshore (Deverchère *et al.*, 2003; Billi *et al.*, 2011). N-S or NNW-SSE shortening is recorded in the High Atlas and the rate of uplift increases after 6 Ma (Frizon de Lamotte *et al.*, 2000; Benaouali-Mebarek *et al.*, 2006; Lanari *et al.*, 2020a, 2020b) and becomes more NW-SE in the north (France) (Cornet and Burlet, 1992; Dèzes *et al.*, 2004; Baize *et al.*, 2013). Zitellini *et al.* (2019) recently brought observations in favor of a compressional reactivation of the southern Tyrrhenian Sea during the Pliocene. Coevally, the Calabria and Gulf of Cadiz accretionary wedges continuously form until the Quaternary (Gutscher *et al.*, 2002; Polonia *et al.*, 2011; Gutscher *et al.*, 2012, 2017). Magmatism is widely distributed within the whole studied region during this time interval with principally alkaline chemistry, except in the east along the Italian peninsula and the nearby offshore domain with the development of calc-alkaline and tholeiitic volcanism above the Apennines subduction zone (Fig. 4B). The recent period also shows the development of alkaline magmatism in the westernmost Mediterranean, best explained by the northward migration of a finger of hot asthenospheric mantle coming from the Canary hot spot (Duggen *et al.*, 2004, 2009).

6 Comparison with the Eastern Mediterranean

The reconstructions at the scale of the Mediterranean region (Figs. 4A and 4B) show the coeval evolution of the Western Mediterranean basins and Aegean domain (see Menant *et al.*, 2016a, for details on the Aegean domain). The Aegean Sea was formed during the same period as the Western Mediterranean Basins. Fast slab retreat and related extension starts at ~ 32–35 Ma and migrates southward during the southward retreat of the Hellenic subduction. The main difference between these two regions is that the Aegean domain never reached the stage of oceanic crust emplacement despite more than 500 km of southward retreat since the end of the Eocene (Jolivet and Brun, 2010). A series of metamorphic core complexes form during the Oligocene and the Miocene starting from the Rhodope in the north and migrating southward toward the Cyclades and Crete. Extension had in fact started earlier in the Rhodope massif with the exhumation of a large MCC (Kounov *et al.*, 2020) but without clear evidence for significant slab retreat such as the migration of a volcanic arc (Jolivet and Brun, 2010). One of the salient features of the Aegean region is the formation of slab tears at around 15 Ma leading to a fast clockwise rotation of the Hellenides and part of the Cyclades until 8 Ma (Kissel and Laj, 1988; de Boorder *et al.*, 1998; Kissel *et al.*, 2002; van Hinsbergen *et al.*, 2005; 2006; Dilek and Altunkaynak, 2009; Jolivet *et al.*, 2015a, 2015b). This same period (15–8 Ma) is also characterized by slab tearing and fast rotations/displacement in the Western Mediterranean, as the western part of the docked AlKaPeCa block starts its westward migration along the northern margin of Africa at around 15–16 Ma. This tectonic evolution is accompanied by a migration of magmatism from east to west along the north African margin (Maury *et al.*, 2000; Do Couto *et al.*, 2016), similarly to the magmatic activity in the western Anatolia-Aegean domain, also as a response to slab tearing (Menant *et al.*, 2016a, 2016b). This is particularly the case of the Miocene Aegean granitoids (Jolivet *et al.*, 2015b; Menant *et al.*, 2016b) which show a westward migration in the middle-late Miocene, superimposing on the global N-S migration of the magmatic activity in the whole Aegean domain in response to slab retreat since 30–35 Ma (Pe-Piper and Piper, 2006; Jolivet and Brun, 2010). The coeval evolution of these two distant domains probably signs a common, global triggering cause for slab tearing in the Mediterranean region.

A second common feature to the Aegean and Western Mediterranean domain is the progressive generalization of alkaline volcanism from 9 Ma to the Present, see also Wilson and Bianchini (1999), Lustrino *et al.* (2011) and Melchiorre *et al.* (2017). Through time one sees a progressive evolution from calc-alkaline to highly-potassic calc-alkaline and then to alkaline in most regions, which could be the result of the progressive slab retreat aided by slab tearing, leading to the enlargement of the back-arc domains, drastic thinning of the overriding crust and underlying asthenospheric upwelling that triggered partial melting of a depleted mantle (Agostini *et al.*, 2007; Menant *et al.*, 2016a). This coeval magmatic evolution across the whole Mediterranean domain reinforces the similar timing noted above in terms of deformation and calls for a similar evolution in the mantle below the Mediterranean. It is not clear whether the progressive formation of back-arc domains alone can explain this evolution in the nature of magmas or an additional influence of mantle plumes is required (Duggen *et al.*, 2009; Faccenna and Becker, 2010). As proposed by Faccenna *et al.* (2013b) the magmatism of Turkey after 10 Ma is strongly influenced by material of the Afar plume that migrated from south to north after the Africa-Eurasia collision at ~30 Ma. Similarly Duggen *et al.* (2009) proposed that a sub-lithospheric corridor has conveyed the hot material of the Canary plume northward until the Western Mediterranean. Both the intrinsic evolution of the subduction zone and the influx of plume material in the Mediterranean realm may have contributed to the observed evolution.

7 Discussion

We now critically discuss the implications of these reconstructions. We start with their reliability and put the emphasis on observations that are not explained by this kinematic model. We then discuss the consequences of the proposed kinematics in terms of driving forces before discussing the rheological implications in the upper and the lower plates. The initial configuration of the subduction zones 35 Ma ago and their polarity is then discussed and compared to other published situations. The transition from the formation of mountain belts to back-arc extension and the end of the HP-LT metamorphism are then summarized and discussed. Finally, the geometry and kinematic evolution of transfer zones in the Western Mediterranean is studied.

7.1 Reliability of reconstructions

Our reconstructions are purely based upon kinematic parameters derived from independent geological data sets and independent from any pre-conceived model (Fig. 3; Supplementary Material #1). Rates of shortening in mountain belts, rates of extension in rifted areas, the timing of shortening or extension, paleomagnetic rotations, ages of magmatic or metamorphic events are indeed all independent from each other. They were taken from the vast corpus of literature available on this region in papers that were not dealing with large-scale geodynamic models. The resulting model is thus independent from our prejudice on the important role played by slab retreat in the dynamics of this fast deforming region. Then, the data gathered in the literature come with some errors that are difficult to estimate. Managing several hundred features in GPlates is complex and entering error bars on each data would have rendered the task impossible. The general picture is however not so different to the first order from other less detailed reconstructions such as those proposed by Rosenbaum *et al.* (2002) for instance, with however more details in the configuration of subduction zones. The novelty of the model also lies in the unprecedented details of the tectonic history decorating the reconstructions and of the motion paths that brings new insights on the kinematic evolution of the region since the late Eocene. The integration of the Western Mediterranean within the whole Mediterranean realm is also an originality of the model and provides an alternative vision to van Hinsbergen *et al.* (2019), reconciling the tectonic history of this region with its magmatic and metamorphic evolution. Different choices in the vast geological data base of this region would probably lead to slightly different reconstructions, but we are confident that the first-order picture would remain the same. Compared to the reconstructions proposed by Schettino and Turco (2006), our starting hypotheses are entirely different as we do not use any information from magnetic anomalies in the back-arc basins and our blocks are deformable, not rigid. Using magnetic anomalies is the most reliable way of reconstructing the past plate kinematics, but in the Western Mediterranean these anomalies are not clearly identified and some of the postulated oceanic crust was shown to be exhumed mantle instead; the justification of this approach is thus questionable.

Despite the errors inherent to the nature of the geological data we used, we think our reconstruction model is solid, but some special consideration should however be focused on the tectonic evolution of the Balearic islands that is not easily explained. If the middle Miocene compressional event recorded on Ibiza and Mallorca can be understood as a consequence of the docking of the AlKaPeCa block with the north African margin (Etheve *et al.*, 2016), neither northwestward-migrating shortening observed on Mallorca from the late Oligocene to the Langhian (Sàbat *et al.*, 2011) nor its different timing in Ibiza and Majorca are explained. Why should the Balearic islands be under compression while the surrounding basins are extending is a difficult and debated question that should be addressed in future works and this is a limitation of our model.

The reconstructions of the Alboran domain are also likely to be debated because of the lack of information on the finite amount of retreat of the subduction below Gibraltar as already mentioned above. In our reconstructions we move the trench eastward back in time – so that the Alboran crust returns to the thickness of an orogen of about 40 km but an alternative choice of a thicker crust would make the retreat larger. This is another weak point of our reconstructions. A similar problem arises for the pre-rift reconstruction of the Gulf of Lion margin. We assume that the crust had a normal thickness of about 30 km before rifting started using the restoration of Jolivet *et al.* (2015a, 2015b). One could instead assume a larger thickness and thus obtain a tighter fit of the Corsica-Sardinia block along the Provence margin. At the scale of the whole Western Mediterranean it would not change much the general picture but it would modify the pre-rift geometry at the scale of the Pyrenees-Gulf of Lion transition. Another debated point is the age of the H*P*-L*T* parageneses in the Internal Zones of the Betics, and more specifically the Nevado-Filábride Complex. We have deliberately chosen to set the peak of pressure in the eclogite-facies in the Eocene, following the results obtained by Augier *et al.* (2005a) and Li and Massonne (2018), rather than the middle Miocene ages obtained by López Sánchez-Vizcaíno *et al.* (2001) and Platt *et al.* (2006), as discussed earlier in the paper. This choice has implications on the geometry and kinematics of the Alboran domain accretionary wedge.

7.2 Kinematics in various frameworks and driving forces

In order to discuss the various possible drivers of the observed kinematics and deformation, in the western and central Mediterranean region, we plot motion paths from our reconstruction model in three different kinematic frameworks, using a facility of GPlates (Fig. 7). The three frameworks are with (1) Eurasia fixed, (2) Africa fixed and (3) the hot spots framework (Seton *et al.*, 2012; Nirrengarten, 2016; Nirrengarten *et al.*, 2018).

The finite amount of slab retreat in our reconstructions comes with large uncertainties, especially for the western margin of the AlKaPeCa block. The amount of retreat of Gibraltar subduction zone is obtained using the amount of shortening within the Betic-Rif orogen (Crespo-Blanc *et al.*, 2016), the amount of extension in large MCCs such as the Sierra Nevada-Sierra de los Filabres, the assumption of a normal initial crustal thickness (~ 30 km) in the future Alboran Sea and the width of the oceanic crust in the Algerian Basin. Using a thicker crust before back-arc rifting would end up with a larger finite displacement of the Alboran block. Our reconstructions thus predict a minimum westward displacement of about 400 km of the Gibraltar arc. The present-day position of the subduction front is further to the west but it corresponds to the front of the sedimentary wedge built at the expense of sediments deposited on the Atlantic oceanic crust, not to the front of the Alboran domain and is thus not indicative of the amount of retreat. A recent study of the East Algerian basin (Driussi *et al.*, 2015), south of the Balearic islands, suggests that the oceanic crust in the basin was emplaced in two steps: a first period with NW-SE extension and SE-ward slab retreat and then E-W extension and westward slab retreat. Depending upon the respective widths of the oceanic domains opened during these two stages, the finite westward displacement of the subduction zone below the western margin of the Alboran block will be different. Maximizing the westward displacement would place the initial position of the trench farther to the northeast, an option chosen for instance by van Hinsbergen *et al.* (2014). Our reconstruction shows a minimum amount of retreat instead, but larger values cannot be ruled out.

Note that some motion paths do not span the entire time frame of 35 Ma because they denote the displacements of features that did not exist in the earliest stages. Because the different features may belong to blocks that have totally different kinematic histories, some motion paths may cross each other, which may look odd. For instance, features belonging to the Apulian block since the earliest stages first follow the motion of Africa before they are caught in the eastward motion due to back-arc extension in the Tyrrhenian Sea. The motion path of such features can then be crossed by the path followed by more internal blocks belonging to the AlKaPeCa block and moving faster during the opening of the southern Tyrrhenian Sea and they are thus not at the same place at the same period. A more detailed explanation is provided in Supplementary Material #3.

These motion paths representing the kinematics within the crust are then compared to the fabrics of the underlying asthenosphere through SKS-wave anisotropy taken from Faccenna *et al.* (2014) and references therein, including Díaz *et al.* (2015) and Salimbeni *et al.* (2013). A more recent compilation for the Alps and North Adriatic region can be found in Salimbeni *et al.* (2018). We here assume that this anisotropy represents the flux in the asthenosphere beneath this region, or the shearing direction between the flowing asthenosphere and the rigid lithosphere (Barruol *et al.*, 2004; Lucente *et al.*, 2006; Buontempo *et al.*, 2008; Jolivet *et al.*, 2009). This observation is also in line with the conclusion of a shear-wave tomography model suggesting an eastward flow of asthenospheric mantle underneath the western Mediterranean back-arc basins (Panza *et al.*, 2007).

As shown in Jolivet *et al.* (2009), the inferred mantle flow below the back-arc domain fits the trend of stretching lineations in exhumed metamorphic core complexes where the deformation in the middle and lower crust can be observed. A swing in the fast direction of the anisotropy is observed from the northern Tyrrhenian Sea (E-W) to the Apennines (N-S), with SKS anisotropy perpendicular to the strike of the arc in the back-arc region and parallel to the arc closer to the trench, a situation that is observed in young orogens (Meissner *et al.*, 2002). Here we see that the motion paths in the Eurasia-fixed framework also show a good fit with the fast direction of the anisotropy in the mantle in back-arc regions, for instance in the northern Tyrrhenian Sea, in Corsica and also in the Internal Betics. The direction of the motion paths is the same in the back-arc region and in the mountain belts and it is parallel to both back-arc extension and shortening in the belt. The motion path are thus perpendicular to the seismic anisotropy that is parallel to the Apennines or the Gibraltar arc (Jolivet *et al.*, 2009).

From the comparison of these two different proxies (mantle fabrics and crustal flow) two main conclusions can be reached. (1) the best fit is obtained with the Eurasia-fixed framework and (2) the crustal flow due to slab retreat toward the east or the west dominates the regional kinematics. It is dominant over the absolute motion of Africa and Eurasia in the hot spots framework. It is best expressed by the eastward motion of Calabria (~ 800 km) (Figs. 7 and 8) that is almost similar in all three frameworks, also on the Gibraltar side where the absolute motion paths are clearly deviated by back-arc opening of the Alboran Sea (~ 400 km, which is a minimum estimate). In the southern Tyrrhenian Sea the earliest part of the motion path (before 15 Ma) are not aligned with mantle fabric. This is because the southern Tyrrhenian back-arc opening only started at 15–10 Ma. As discussed by Jolivet *et al.* (2009) the mantle fabric can be totally reset in less than 10 Ma and only the recent part of the motion path should be considered for comparison.

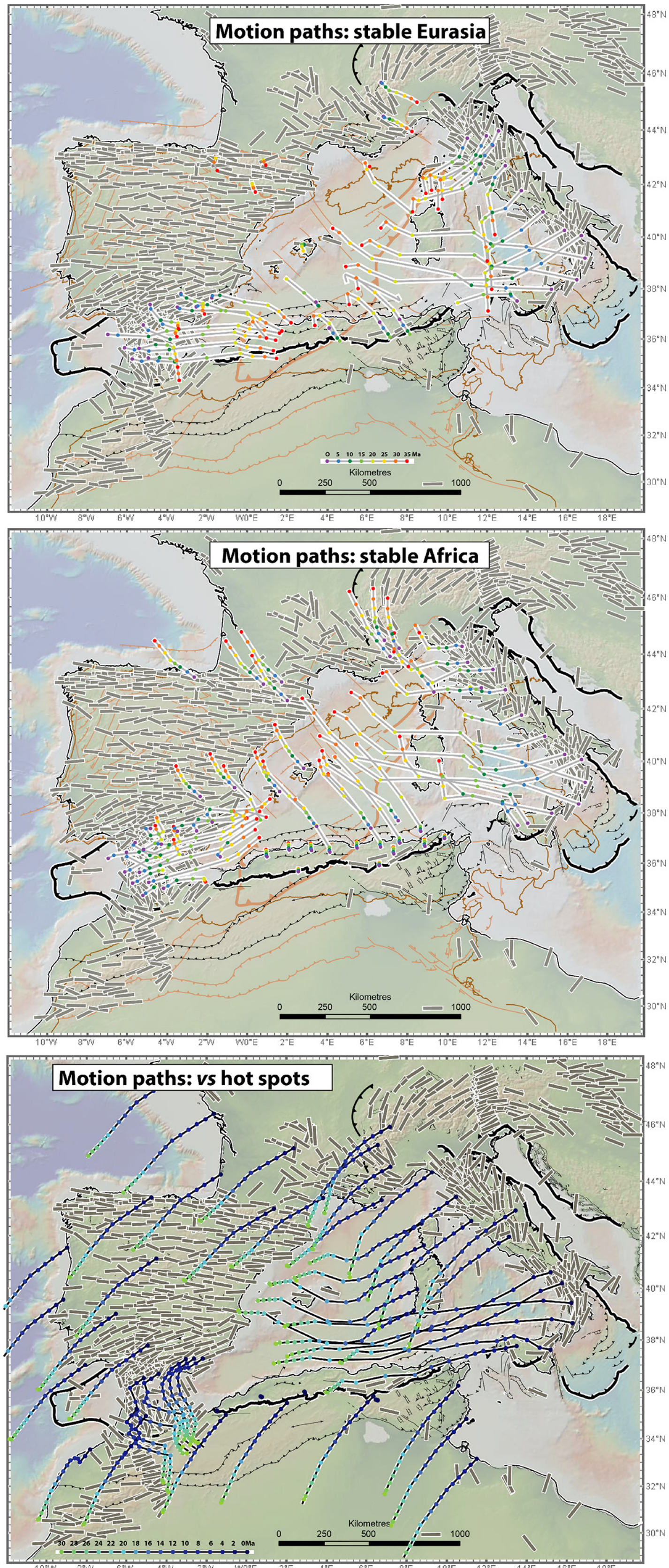


Fig. 7. Motion paths calculated from the reconstructions in three different kinematic frameworks. Upper: stable Eurasia; Middle: stable Africa, Lower: absolute motion (Atlantic-Indian hot-spots). Grey bars: fast polarization directions of SKS-waves (Faccenna *et al.*, 2014; Díaz *et al.*, 2015).

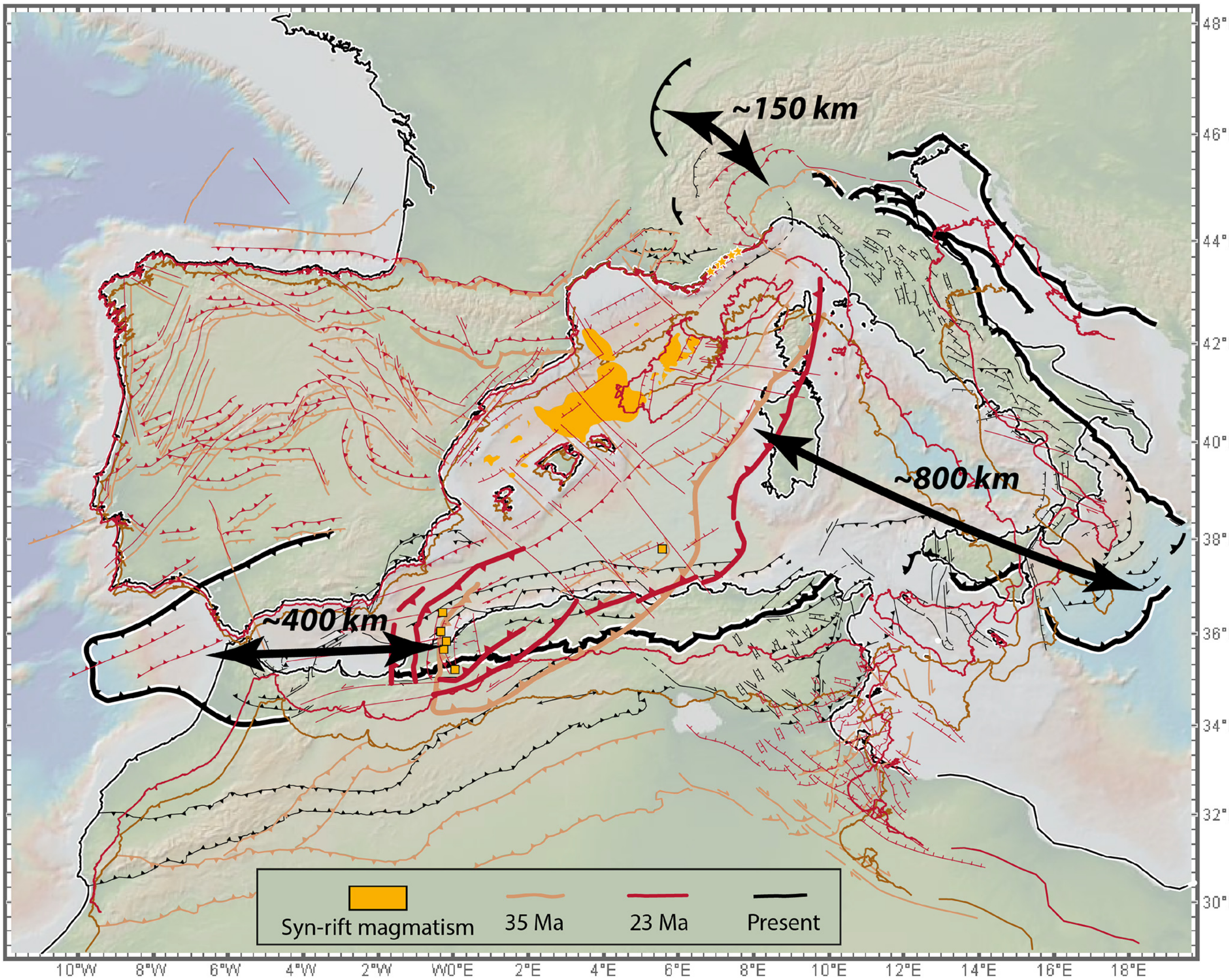


Fig. 8. Synthetic résumé of the displacements since 35 Ma. Red lines show the reconstructions at 23 Ma and light-brown lines the reconstructions at 35 Ma. The orange domain is the volcanic province recognized in the Valencia basin until the transition with the Gulf of Lion and in Sardinia. Orange stars represent the adakitic Oligocene volcanism of Provence. Orange squares stand for the Late Eocene-Oligocene magmatism of Malaga region.

The best fit of mantle fabric and crustal kinematics obtained within the Eurasia-fixed framework suggests that mantle flow due to slab retreat, modulated by slab tearing and slab detachment, is the dominant driver of the deformation in Mediterranean back-arc regions, as commonly proposed (Wortel and Spakman, 1992; Carminati *et al.*, 1998a, 1998b; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a, 2001b, 2004; Jolivet *et al.*, 2009; Faccenna and Becker, 2010; van Hinsbergen *et al.*, 2014). The asthenospheric flow induced by slab retreat underneath the overriding plate moreover creates a component of shear at the base of the lithosphere that further controls the deformation in the back-arc region (Jolivet *et al.*, 2009, 2015a, 2015b, 2019; Sternai *et al.*, 2014). The poorer correlation of the motion paths in the hot-spots reference frame with the seismic anisotropy shows that absolute motion of the African and Eurasian plates is not the dominant process creating the mantle fabric here. Slab retreat is instead the first-order driver. This interpretation is well in line with recent numerical modelling of the interaction between surface kinematics and mantle convection showing that plate attached to a subducting slab are driven by slab pull and plates not attached to a subducting slab are mostly driven by the flow of mantle underneath (Coltice *et al.*, 2019). In this school of thought, slab dynamics is the main driver of plate deformation in back-arc regions, but other models have been proposed. In a recent paper, Mantovani *et al.* (2020), after Mantovani *et al.* (2009), discuss the geological features of the Western Mediterranean and favor an alternative mechanism involving mostly extrusion driven by the African-Eurasia convergence. This model is partly similar to early models proposed by McKenzie (1972, 1978) and Tapponnier (1977). It was also proposed to explain the deformation of south-east Asia as a consequence of the northward indentation of India into Asia (Tapponnier *et al.*, 1982). Extrusion models however involve large-scale strike-slip fault systems, such as the Red-River Shear Zone or the North Anatolian Fault, linking the region of crustal thickening in the collision zone and the back-arc domain outside the collision zone. No such fault is observed in the Western Mediterranean region and one does not find either the equivalent of the Tibetan plateau or the Anatolian plateau there. A second objection to this type of model is that it can hardly explain the observed extension at the leading edge of the extruded system at variance with models involving slab retreat such as Sternai *et al.* (2014) and Capitanio (2014). Further numerical studies should focus on the respective merits of both classes of models. A third objection is the observed velocities in back-arc regions on the Western Mediterranean which are much higher than the velocity of Africa-Eurasia convergence. We thus think that the domination of eastward and westward kinematics over the absolute kinematics is a consequence of the high velocity of slab retreat that amounts to more than 10 cm/yr, up to 19 cm/yr according to Nicolosi *et al.* (2012), especially in the South Tyrrhenian region, in comparison to the slower absolute motion of Africa and Eurasia about 1 cm/yr (Macchiavelli *et al.*, 2017).

7.3 Nature of the subducting lithosphere

In our reconstructions, we show a partly oceanic connection through en-échelon basins controlled by left-lateral transform faults in the narrow space between African and Iberia, never wider than 200–300 km, between the Mediterranean domain and the Atlantic. The nature of crust flooring the basin is in fact not precisely known. All recent reconstructions in the Mesozoic indeed consider a segmented domain along NW-SE trending transform faults accommodating the oblique Africa-Eurasia motion. Schettino and Turco (2009) show a fully oceanized domain while Fernàndez *et al.* (2019) show a more detailed picture including exhumed mantle along the ocean-continent transitions in between the transform faults. Leprêtre *et al.* (2018) and Gimeno-Vives *et al.* (2019) show a partly oceanized domain with exhumed mantle. We have followed these recent interpretations in the reconstructed maps.

There was for sure a marine domain where platform sediments were deposited in the Mesozoic before the beginning of shortening in the Late Cretaceous, but no evidence of true ophiolites that would testify for a large open ocean has ever been found. The only tectonic unit that could call for an oceanic environment is the Bédar-Macael Unit at the top of the Nevado-Filábride complex where metabasites and ultramafic rocks show evidence of HP-LT metamorphism in the eclogite facies during the Eocene (Puga *et al.*, 2000, 2002, 2017; Augier *et al.*, 2005a, 2005b, 2005c, 2005d; Li and Massonne, 2018). The so-called Cobdar Ocean (Cb. O. in Fig. 5A) could have been a small ocean or a hyper-extended continent-ocean transition that could also include farther west the Ronda and Beni Bousera peridotite as proposed by van Hinsbergen *et al.* (2014) and Bessière (2019). Its exact width and position before the formation of the nappe stack is, however, difficult to assess. Our best guess is that it was located to the north of the AlKaPeCa terranes and was integrated in the accretionary wedge by a southward subduction during the Eocene, like in Chalouan and Michard (2004). Otherwise, all tectonic units involved in the Betic-Rif accretionary wedge are of continental origin, probably parts of a thinned continental domain located between Africa and Iberia that was progressively incorporated in the Betic-Rif accretionary wedge during the westward migration of the arc. It could be indeed argued instead that the true oceanic domain has been totally subducted and that no remains can be found within the orogen and we cannot totally exclude this possibility (see Vergés and Fernàndez, 2012).

Further to the east and northeast, in the Apennine-Calabria subduction zone, the subducting lithosphere is more likely to be truly oceanic for two main reasons. First, a calc-alkaline magmatic arc has developed above the subduction zone in Sardinia and Provence from 32 Ma, suggesting the classic subduction of an oceanic domain. The second reason is that the slab now seen in tomographic models below the Tyrrhenian Sea was dense enough to induce a fast retreat, which is less likely with a continental lithosphere.

7.4 thick, hot and weak crust in the upper plate

Post-dating Late Cretaceous-Eocene mountain building, from Corsica to the Betic-Rif orogen, rifting in back-arc regions of the Western Mediterranean, like in the Eastern Mediterranean, leads to the formation of MCCs, topped with low-angle normal faults and ductile shear zones, and in some cases intrusions of granitoids involving a significant component of crustal melt (Daniel and Jolivet, 1995; Westerman *et al.*, 2004; Farina *et al.*, 2010). Thickened in the preceding stage, the lower continental crust was thus warm and weak when extension started. In the Tuscan archipelago, the Monte Capanne and Porto Azurro plutons of Elba Island (Figs. 4B and 5F) were exhumed below east-dipping low-angle detachments (Keller and Pialli, 1990; Daniel and Jolivet, 1995; Jolivet *et al.*, 1998; Collettini and Holdsworth, 2004; Westerman *et al.*, 2004). The Betic Cordillera provides the example of the Sierra Nevada-Sierra de Los Filabres and Sierra Alhamilla MCCs characterized by HP-LT followed by HT-LP (amphibolite-facies) metamorphism exhumed below the top-to-the west Filabres low-angle detachment (Jabaloy *et al.*, 1993; Augier *et al.*, 2005c). In the Gulf of Lion, the distal part of the margin has been interpreted as a MCC made of exhumed lower crust at the time of rifting below a north-dipping detachment (Jolivet *et al.*, 2015a, 2019). A conspicuous sub-aerial erosion surface at the top of this MCC shows that this rifting stage was accommodated with a shallow basin or even emerged crust (Jolivet *et al.*, 2015a), while the transition zone toward the Valencia Basin shows a boudinage of the lower crust and low-angle ductile shear zones (Maillard *et al.*, 2020). This region is also characterized by intense volcanism and magmatic underplating during rifting both on land (Sardinia) (Casula *et al.*, 2001; Lustrino *et al.*, 2009) and offshore (Eastern Valencia Basin) (Maillard and Mauffret, 1999; Maillard *et al.*, 2020). While most of the crustal thinning observed in the southwestern part of the Valencia basin dates back to pre-Albian times (Etheve *et al.*, 2018), the northeastern part of the basin instead shows evidence for Oligocene rifting (Maillard *et al.*, 2020). All these observations show that the crust was weak and hot in the overriding plate over the eastern Pyrenees, the future Gulf of Lion and eastern Valencia basin, which appears crucial for transmitting shear stress from the flowing asthenosphere to the lower crust and for explaining the similar mantle and deformation pattern in these warm regions (Fig. 7;

see also discussion in Sections 7.2 and 7.6) (Jolivet *et al.*, 2018, 2019). Three reasons explain this weakness: previous crustal thickening, intense magmatism and the presence of the asthenosphere at low depth during slab retreat and tearing.

7.5 Subduction polarity

In opposition to most classical models, Vergés and Fernàndez (2012) (Fig. 1B) proposed that the slab now observed below the Gibraltar arc and southern Iberia evolved from the southeast-dipping subduction below the Alboran domain, separated from the main northwest-dipping subduction of the Ionian Sea further east by a NW-SE transform fault. Other models favor a single northwestward subduction that has acquired the observed strong curvature during retreat (Lonergan and White, 1997; Lacombe and Jolivet, 2005; van Hinsbergen *et al.*, 2014) or two facing subductions and much less westward retreat (Michard *et al.*, 2002; Chalouan and Michard, 2004; Leprêtre *et al.*, 2018). Our reconstructions do not show a continuous simple northward subduction and, instead, favor a situation partly similar to Michard *et al.* (2002) on the one hand, and Vergés and Fernàndez (2012) on the other hand. Reconstructing the position of the trench without any *a priori* model leads to a situation at ~35 Ma where the future Gibraltar subduction is oriented N-S and extends eastward within a southeastward dipping trench where the so-called Cobdar Ocean is consumed. A significant difference with Vergés and Fernàndez (2012) is however that the upper plate of the south-dipping subduction is the AlKaPeCa block and not Africa. This situation is more similar to the proposition of Chalouan and Michard (2004) and Leprêtre *et al.* (2018) than van Hinsbergen *et al.* (2014) and it does not involve the transform fault postulated by Vergés and Fernàndez (2012) to accommodate the opposite dips of the two subduction zones, a structure that is not expressed in the local geology. Subduction below the two sides of the AlKaPeCa blocks solves this problem.

The motion paths (Fig. 7) show that the displacements with respect to Africa are almost parallel to the North African margin with very little convergence, and slightly oblique to the south Iberian margin with a minor amount of convergence. Most of the displacement is accommodated across the N-S trending subduction zone to the west. This situation is due to the option we have chosen for closing back the Algerian basin and the Alboran Basin. Had we instead chosen the van Hinsbergen *et al.*'s (2014) solution we would have had to consume most of the subducted lithosphere across the north-dipping subduction, progressively retreating toward the southwest. The solution of van Hinsbergen *et al.* (2014) is mainly inspired by a numerical model of a retreating slab leading to a final geometry similar to the observed one. Chertova *et al.* (2014) indeed test different models of evolution of the Western Mediterranean with 3-D numerical experiments. They conclude that the best fit of the present-day geometry of the imaged slab is obtained with an initial short northwestward subduction south of the Balearic Islands compared to models with a continuous subduction from Gibraltar to the Baleares or initial subduction underneath the African margin. A strong rotation of the slab, more than 180°, is observed with this model. The model also fits several time constraints such as the timing of docking of the AlKaPeCa blocks with the northern margin of Africa. In this study, we worked totally independently of any *a priori* model and our reconstruction does not lead to more than 400 km of slab retreat. The van Hinsbergen *et al.*'s solution remains possible because some of the shortening might have been accommodated in now subducted units. With the option we chose for reconstructing the Alboran domain, which minimizes its displacement, we end up with a situation intermediate between that of Vergés and Fernàndez (2012) and Chalouan and Michard (2004).

7.6 From shortening to extension in the overriding plate

The reconstructions highlight a 3-D complexity at the junction of the Pyrenees and the Gulf of Lion. The present-day Pyrenean Belt ends abruptly to the east where they give way to the rifted margin of the Gulf of Lion (Réhault *et al.*, 1984; Mauffret *et al.*, 1995; Séranne, 1999; Jolivet *et al.*, 2015a, 2015b, 2019; Chevrot *et al.*, 2018). Rifting starts at about 32 Ma, which is coeval with an uplift and exhumation episode in the Axial Zone of the Eastern Pyrenees (Gorini *et al.*, 1993; Gorini *et al.*, 1994; Fitzgerald *et al.*, 1999; Jolivet *et al.*, 2015a), as expected on the shoulder of a rift. However, shortening continues in the western part of the Pyrenees where south-vergent thrust faults are still active until the Early Miocene (Muñoz, 2002; Jolivet *et al.*, 2007; Labaume *et al.*, 2016). This 3-D pattern is complex and the geometry of the stress field not easy to understand. Why does compression continue during ~10 Ma in the west, while extension is well underway in the nearby Gulf of Lion and in the Valencia Basin?

Nevertheless, apart from limited on-going southward thrusting in the west, the main shortening across the Pyrenees ends approximately when the subduction regime changes in the whole Mediterranean, from compressional to extensional, in the late Eocene to Oligocene. Whether compressional stresses are still transmitted from Africa to Europe through the Pyrenees west of the extending domain is an open question. The Pyrenees end along the Mediterranean coastline and seismic data show the progressive crustal thinning from the mountain belt to the Gulf of Lion passive margin (Chevrot *et al.*, 2018; Díaz *et al.*, 2018). The first extensional features seen in the eastern Pyrenees consist of minor high-angle normal faults with a strike-slip component and the motion along these faults controlling the Cerdanya half-graben dates back to the Middle Miocene, thus posterior to rifting in the Gulf of Lion (Pous *et al.*, 1986; Cabrera *et al.*, 1988; Gabas *et al.*, 2016). The Catalan-Baleares Transfer fault can kinematically accommodate the transition from shortening to extension, but the distribution of stresses is complex. Nevertheless, it remains that the main Pyrenean shortening ceased progressively when slab retreat started, between 32 and 20 Ma and that the Pyrenees were fast brought down to sea level and then submerged in the east to be replaced by the Gulf of Lion passive margin (Jolivet *et al.*, 2020). 20 Ma is also the period of the last shortening event along the southern Pyrenean thrust front. One reason could be the change from southeastward to westward slab retreat in the Alboran region that deviated the mantle flow underneath in an E-W direction, precluding the transmission of N-S compressional stresses.

In the meantime, Africa-Eurasia convergence was active all along and its effects were felt again only once the Gibraltar slab had retreated westward far enough in the Alboran region (Jolivet *et al.*, 2006; Spakman *et al.*, 2018). Some compression was felt within the African plate during the Oligocene and Miocene south of the Atlas (Tesón *et al.*, 2010), but the main stages of construction of the Atlas are recorded in the Eocene and the Pliocene (Frizon de Lamotte *et al.*, 2000; Lanari *et al.*, 2020a, 2020b). One simple explanation is that the fast retreat was diverting the mantle flow toward the west, as is also suggested nowadays from SKS-wave anisotropy below Iberia (Fig. 7) further suggesting that mantle flow underneath plays a key-role in driving crustal deformation, at least in regions where the crust is warm and weak (Jolivet *et al.*, 2009, 2018; Sternai *et al.*, 2014; Menant *et al.*, 2016b). The interaction of the dipping slab with the mantle flow driving the absolute northeastward motion of Africa and Eurasia may play an active role in the present-day stress regime (Spakman *et al.*, 2018) but the Pliocene resumption of compression in the High-Atlas (Frizon de Lamotte *et al.*, 2011) and the recent NNW-SSE compression recorded all across France (Bergerat, 1987; Cornet and Burlet, 1992; Baize *et al.*, 2013) as well as the NW-SE trending σ_{hmax} within Iberia in a extension or strike-slip regime (de Vicente *et al.*, 2008) suggest that a large-scale engine should be looked for involving the relative motion of Africa, Adria and Europe as well as the mantle flow underneath.

7.7 Coeval end of HP-LT metamorphic conditions

Our reconstructions go back to the Late Eocene, which corresponds to the end of HP-LT metamorphic conditions in the studied region. Ages of the HP-LT metamorphism in the Betics (Alpujárrides and Nevado-Filábrides) are debated but Eocene ages are likely (Augier *et al.*, 2005a; Platt *et al.*, 2005; Li and Massonne, 2018). In addition, a recent dating campaign in the Alpujárride units in the eastern and central Betics (Bessière, 2019) shows that the peak of metamorphism dates back to ~38 Ma. We thus consider here that the middle Miocene ages obtained in the Nevado-Filábride complex (López Sánchez-Vizcaíno *et al.*, 2001; Platt *et al.*, 2006) are representative of the exhumation of this complex, not of the peak of pressure; for a detailed discussion of published radiochronological data, see Bessière (2019). Alpine Corsica, the Alps or Calabria show coeval peak metamorphism ages within errors, very close to the major change from shortening to extension in the overriding plate. The case is very clear in Alpine Corsica where the 34 Ma age is within error similar to the age of the first extension along the East Tenda Shear Zone (Brunet *et al.*, 2000). In the Dora Maira massif of the Western Alps, the youngest published ages of the UHP metamorphism are also around 33–35 Ma (Duchêne *et al.*, 1997; Rubatto *et al.*, 1997; Rubatto and Hermann, 2001; Schertl and Hammerschmidt, 2016), which is almost the age of the flysch-to-molasse transition in the western Alps and the westward propagation of the thrust front (Vignaroli *et al.*, 2008, 2009). In the Edough Massif, NE Algeria, the 32 Ma age of rutile is thought to be close to the age of peak pressure conditions (Bruguier *et al.*, 2017). Without evidence for UHP metamorphism, the alpine metamorphic units of Calabria also yield peak metamorphism ages between 33 and 38 Ma before overprinting with extensional retrograde deformation (Rossetti *et al.*, 2004). It thus appears that everywhere along the future Apennines subduction zone as well as in the Alps, available ages of the last HP-LT metamorphism are close to the age of the first back-arc rifting nearby.

We interpret this situation by considering that these ages represent the last moment when exhumed tectonic units reached the maximum depth within the subduction zone and started their exhumation because the subduction regime had changed. Because of slab retreat, the subduction channel opened and accreted units were soon exhumed. Slab retreat facilitates exhumation of HP-LT tectonic units, as shown by the case of the Cycladic Blueschists or the Phyllite-Quartzite Nappe in the Aegean region (Jolivet *et al.*, 1994, 2003; Brun and Faccenna, 2008; Jolivet and Brun, 2010). In the case of the Western Alps, the coeval fast exhumation of the UHP units and westward propagation of the thrust front could also be a consequence of the inception of slab retreat in the Apennines subduction zone and induced opening of the subduction channel below the Alps, as initially proposed by Vignaroli *et al.* (2008). A significant difference however appears between the Western and Eastern Mediterranean. In the west, there is almost no HP-LT metamorphic units yielding ages younger than 30 Ma, the island of Gorgona in the Northern Tyrrhenian Sea and the MCCs of Tuscany (Brunet *et al.*, 2000) with ages around 25 Ma being exceptions. In the Eastern Mediterranean instead, blueschist exhumation lasted until about 16–15 Ma in Crete and the Peloponnese (Jolivet *et al.*, 1996, 2010), despite active back-arc extension since about 30 Ma. This difference might be due to the presence of a wide oceanic domain not yet subducted south of Crete in the Miocene, whereas the Calabrian arc is much narrower. This would allow slab retreat to continue in the Eastern Mediterranean region associated with protracted basal accretion and exhumation of HP-LT metamorphic units until more recent periods.

7.8 Slab tearing, transfer zones and back-arc kinematics

We now focus on the pre-8 Ma period, before the change toward the resumption of compression, progressively generalized to the whole Western and Central Mediterranean. The total displacements of the Calabria and Gibraltar arc since 35 Ma are respectively 800 and 400 km, the latter value being a minimum (Fig. 8). These large displacements of the retreating slabs imply similar displacement of the mantle below the overriding plate, which implies long-distance effects of the retreat, which is well shown by seismic anisotropy (Barruol and Granet, 2002; Lucente *et al.*, 2006; Jolivet *et al.*, 2009; Salimbeni *et al.*, 2018). The eastward mantle flow related to this slab retreat is indeed felt as far as the Alps and the Pyrenees (Fig. 7). The fast direction of seismic anisotropy showing the toroidal asthenospheric flow around the Alpine arc gets parallel to the fast direction under Provence and the Pyrenees, which is in turn

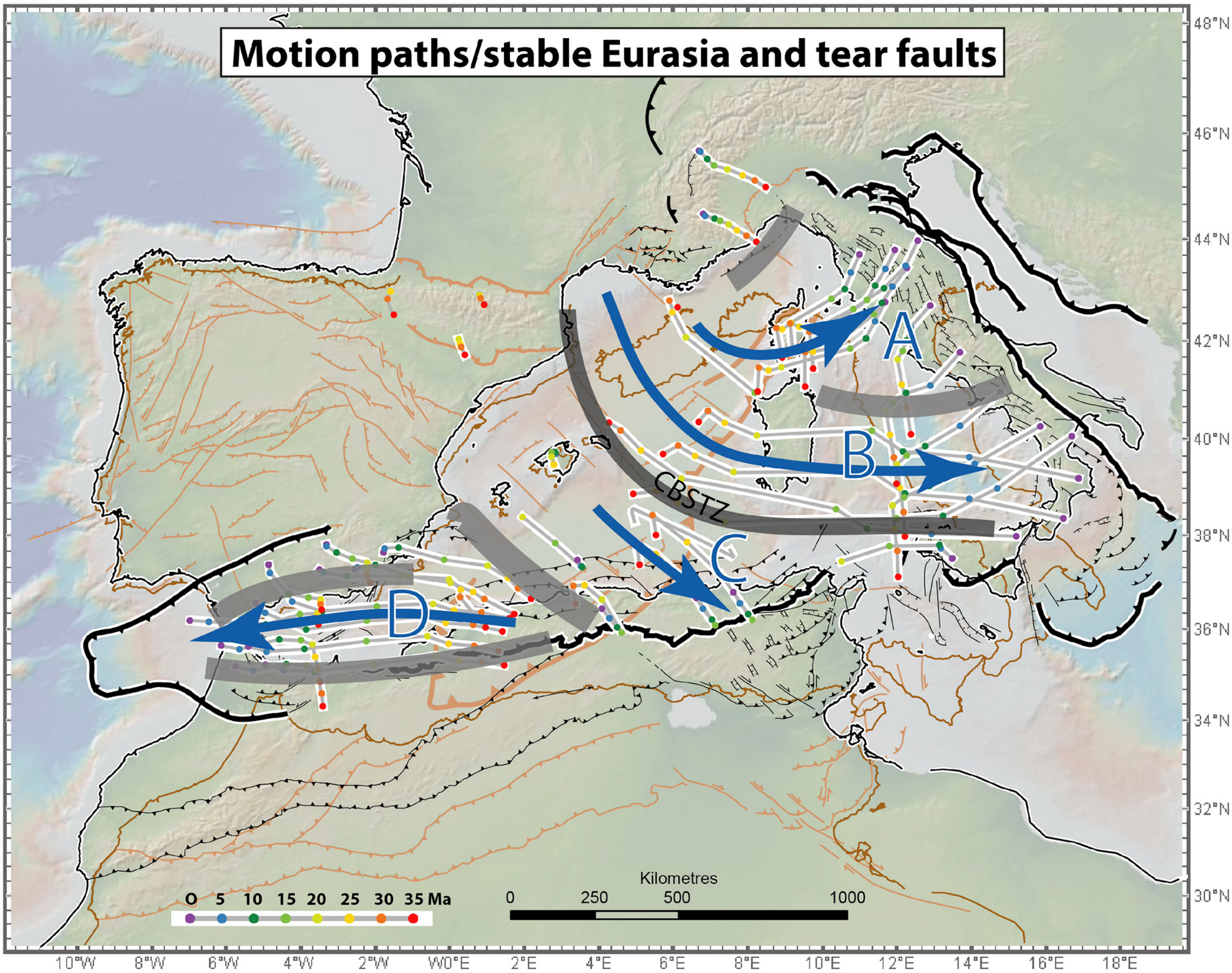


Fig. 9. Different kinematic domains and transfer zones (thick grey lines) discussed in the text. CBSTZ: Catalan-Baleares-Sicily Transfer Zone. Large blue arrows show the average direction of displacement between the main transfer zones.

parallel to the main direction of slab retreat on the map showing motion paths in the Eurasia-fixed framework.

The displacement field shows, however, discontinuities bounding four kinematically different domains (domains A, B, C and D in Fig. 9). Some are sharp like between domains B and C or north and south of domain D, other are more diffuse like between domains A and B. These discontinuities result from the independent behavior of torn slab pieces during retreat. The structures accommodating these differential movements in the overriding lithosphere (transfer zones) are variable from one place to another. The sharpest and largest of these transfer zones is between domains B and C. It runs from the western margin of the Gulf of Lion all the way to the northern margin of Sicily (Catalan-Balearic-Sicily Transfer Zone, CBSTZ in Fig. 9), with a strike changing following the small circles of the rotation of Corsica and Sardinia. Its northern part corresponds to a series of NNW-SSE-striking dextral transfer faults, the geometry of which is now well described based on the interpretation of seismic lines in the Gulf of Lion and Valencia Basin (Maillard *et al.*, 2020). The Catalan transfer Fault runs along the western margin of the Gulf of Lion and the eastern coastline of the Pyrenees. It was initially placed here for kinematic compatibility but was not clearly identified as such. A recent study (Canva *et al.*, 2020) shows that it is associated with a prominent NNW-SSE-striking magnetic anomaly interpreted as underplated gabbroic rocks. The North Balearic Fracture Zone is found further south. The fracture zone are well imaged in seismic profiles crossing the transition between the Valencia Basin and the Gulf of Lion and is associated with syn-rift basins and lower crustal domes (Maillard *et al.*, 2020). The relative motion across the CBSTZ is however not purely strike-slip. Motion paths in the Eurasia-fixed or Africa-fixed frameworks show a divergence southwest of Sardinia, implying a component of extension perpendicular to the transfer zone. This extension component could explain the fan-shaped pattern of oceanic magnetic anomalies in this region, *i. e.* the Hamilcar anomalies of Mauffret *et al.* (2004).

The transfer zone between Domains B and A is diffuse and does not correspond to any specific strike-slip fault. It is rather the limit between the northern and southern Tyrrhenian seas with a transition from upper crustal blocks in the north to highly thinned continental crust and exhumed mantle to the south (Prada *et al.*, 2014, 2018). The transfer zone between domain C and D is also diffuse, but Strzerzynski *et al.* (2010) have identified a large dextral discontinuity oriented N120°E limiting the Internal Kabylides to the west. It has been active between the Eocene and ~19 Ma, thus just before the main change in the direction of slab retreat from southward to westward. They also propose that similar structures do exist and have accommodated the opening of the Algerian Basin.

The northern and southern limits of domain D (the Alboran Sea) are more clearly identified. The southern limit corresponds to the northern steep Moroccan margin, reworked by compression in the recent period. The trajectories with fixed Africa show that the motion is oblique on the margin with a component of thrusting and a component of left-lateral shear with respect to the direction of the western north African margin (Fig. 7, 2nd panel). Several left-lateral faults (Nekor Fault, Jebha Fault) and the thrust contacts within the Rif have accommodated this differential movement (Lafosse *et al.*, 2018). The northern transfer zone corresponds to the Betics. Between 20 and 8 Ma the displacement has been mostly from east to west in the Betics with the exhumation of the Sierra Nevada-Sierra de los Filabres MCC, the Sierra Alhamilla MCC and some dextral strike-slip faults, plus a component of northward thrusting in the north (Platt and Vissers, 1989; Vissers *et al.*, 1995; Augier *et al.*, 2005c, 2013; Jolivet *et al.*, 2006, 2008; Martinez-Martinez *et al.*, 2006; Frasca *et al.*, 2015). The two MCC are elongated parallel to the direction of shearing, making them a-type domes in the sense of Jolivet *et al.* (2004). As shown by Le Pourhiet *et al.* (2012) such domes can form in an extensional context with a significant strike-slip component. Similar domes are found in the center and east of the Aegean Sea in the wide transfer zone above a major tear in the Hellenic slab (Jolivet *et al.*, 2015b).

The reaction of the crust in the overriding plate above slab tears is thus variable, with either localized strike-slip faults with some extensional component, deep intrusions and crustal domes, or with mainly a-type crustal domes oriented parallel to the tear fault, with a different expression in the brittle and lower crusts. Other examples show less localized structures that remain to be clearly identified.

The origin of the tear faults (in the subducting plate) is quite obvious in some cases, much less in others. The main tear below the CBSFZ can have originated along the ocean-continent transition south of Apulia (Figs. 5A and 9) but this is highly debatable as the geometry of the Apulian block is poorly known. The earliest evidence for a tear in this region is found in Provence where adakitic magmatism is found in the Late Eocene (Réhault *et al.*, 2012) (Fig. 4). The transition between the Pyrenees and the Gulf of Lion is approximately located south of the left-lateral bend between the northern thrust front of the Pyrenees and Provence. Although the strike of the thrust front along this bend is slightly different from the strike of the transfer zone, it is possible that this bend corresponds to a transfer zone active during Pyrenean shortening (Séranne, 1999), which would have been reactivated during subsequent back-arc rifting. The tear faults on either side of the Alboran Sea are easier to understand as they correspond to the limits of the oceanic or thinned continental domain between Iberia and Africa. The transfer zone between domain C and D could originally be one of the transform faults in the Tethyan oceanic domain that are thought to display a similar orientation as in Vergés and Fernàndez (2012) and Leprêtre *et al.* (2018).

In view of our reconstructions, it also appears that the spatial and temporal distribution of magmatism is in part controlled by slab retreat, tearing and transfer zones. Thus, the CBSTZ has localized magmatism since its initiation in the Oligocene and early Miocene (Canva *et al.*, 2020; Maillard *et al.*, 2020) until the recent development of offshore magmatism north of Sicily (Fig. 4) (Cocchi *et al.*, 2009). Figure 4 shows the distribution of magmatism in this region from the first adakites in the late Eocene to the Oligocene and early Miocene magmatism underplated below the Catalan transfer zone as well as the syn-rift magmatism of the northeastern Valencia basin along transfer zones. Transfer zones within the Valencia trough are indeed associated with magmatic venues (Maillard and Mauffret, 1999; Maillard *et al.*, 2020). The southeast extremity of the CBTSZ, north of Sicily, is spatially correlated with the recent emplacement of K- and Na-alkaline igneous rocks, above a slab tear and associated reorganization of mantle flow around the edges of the remaining piece of slab (Faccenna *et al.*, 2005). In addition, the westward and eastward migration of magmatism along the northern margin of Africa is related to the progressive pealing from underneath of the torn African lithospheric mantle during its retreat (Maury *et al.*, 2000). A similar timing, distribution and composition of magmatic products has also been reported further east in the Western Anatolia-Eastern Aegean domain during the middle Miocene (Fig. 4), related to a major slab tearing event occurring coevally underneath (Dilek and Altunkaynak, 2009; Menant *et al.*, 2016a, 2016b).

The change from N-S to E-W extension in the Betics occurred some 20 Ma ago (Crespo-Blanc *et al.*, 1994; Crespo-Blanc, 1995; Azañón *et al.*, 1997; Jolivet *et al.*, 2006). A similar change happened approximately at 15–20 Ma north of the CBSTZ as shown by the kinematic trajectories, at the approximate time when the AlKaPeCa units collided with the northern margin of Africa (Figs. 5D and 5E). In different reconstructions it happens as soon as 20–25 Ma (Leprêtre *et al.*, 2018) but the exact age of docking is difficult to assess precisely because the original outboard the Africa is now buried below the orogen. Considering an age of ~20 Ma for the docking would make it coeval with the change in the direction of extension and thus of the direction of slab retreat from N-S to E-W in the Betics. It can be proposed that it is precisely the docking on the African margin that led to tearing the subducting plate in two main pieces on either side of eastern Algeria where the oldest magmatic rocks are observed along the North African coast, one retreating eastward and one retreating westward. This scenario is compatible with the migration of magmatism from east to west along the northern margin of Africa between 20 and 9 Ma (Fig. 4) (Maury *et al.*, 2000; Coulon *et al.*, 2002). The transition through time from oceanic to continental subduction in the central part of the system at around 4°E would trigger a detachment of the slab there, which would then propagate eastward and westward.

8 Conclusions

Detailed reconstructions of the Western Mediterranean from 35 Ma to the Present are presented, based on geological constraints in mountain belts and extended domains, as well as paleomagnetic and P-T-t constraints. Once integrated within the entire Mediterranean realm, extracted paleo-tectonic maps and reconstructed cross-sections provide an unprecedented precise framework for discussing the interactions between retreating subducting slabs and the overriding plate in a confined environment.

One of the initial questions was the configuration of subduction zones before the inception of slab retreat. The backward step-by-step reconstruction leads to a model where two subduction zones are active. One subduction proceeds toward the NW below the Baleares, Corsica and Sardinia and one toward the east or southeast below the Alboran block. This configuration is intermediate between the options proposed by Michard *et al.* (2002) and Vergés and Fernàndez (2012).

The comparison of motion paths extracted from the reconstruction model shows a good fit with the fast polarization directions of SKS-waves in the Eurasia-fixed framework, suggesting that the main driver of crustal deformation between 35 and 8 Ma is the retreat of slabs eastward and westward and the associated asthenospheric flow, as the velocities of crustal blocks following the retreat are much larger than the motion of Africa with respect to Eurasia at the limits of the deforming area. The main direction of retreat was N-S or NW-SE until the AlKaPeCa blocks docked against the northern margin of Africa at around 20 Ma. The subsequent subduction of continental lithosphere then initiated a detachment of the slab that propagated as tears toward the west and east along the northern margin of Africa, inducing a change in the direction of extension, then mostly E-W. Other slab tears along subducting continental margins include the southern margin of Iberia and the south of Provence, leading to a faster rotation of the Corsica-Sardinia block in the late Oligocene-early Miocene.

Between 32 and 8 Ma the main driver was thus the retreat of slab pieces eastward and westward, and the related asthenospheric flow has prevented the transmission of compressional stresses across the Western Mediterranean. After 8 Ma, the resumption of N-S compression observed across the Western Mediterranean can be interpreted as the result of a recoupling of the Eurasia and Africa plate after the retreating slabs had migrated westward and eastward far enough.

As a consequence of the multiple segmentation of the two former subduction zones, several kinematic domains can be evidenced in the western Mediterranean region, which retreated in different directions with different velocities. The limits of these domains are transfer zones that form above tears in the slab. Several transfer zones are recognized, the largest one being the Catalan-Baleares-Sicily Transfer Zone. These transfer zones have different expressions, either transtensional strike-slip faults associated with syn-rift basins, crustal domes and magmatic underplating, or large-scale MCCs oriented parallel to the main shearing direction (a-type domes).

The association of a thick crust in former mountain belts and subduction-related magmatism induces a weak rheology in the overriding plate lithosphere, explaining the formation of low-angle normal faults and crustal domes, which are exhumed in back-arc basins. It probably also explains the distributed extension and the absence of simple localized strike-slip faults above slab tears, replaced by low-angle detachments and a-type domes.

Like in the eastern Mediterranean a clear link is observed between the distribution of magmatism and evolving kinematics in the back-arc region. Key-tectonic events such as the inception of back-arc extension or the formation of transfer zones, which relate at depth with slab retreat and slab tearing, are coeval with specific magmatic production like the Provençal adakites at 35–32 Ma or the migration of magmatism along the northern margin of Africa after ~ 20 Ma.

Supplementary Material

Supplementary Material #1. This file contains the following supplementary information: (1) Constraints used for reconstructions; (2) GPlates rotation file references list; (3) List of features (rigid polygons) used in the reconstructions.
Supplementary Material #2. This file contains the GPlates rotation file.
Supplementary Material #3. Figure S1: Detailed motion paths of selected points of the periphery of the Southern Tyrrhenian Sea showing crossings.

The Supplementary Material is available at http://www.bsgf.fr/10.1051/bsgf/2020040/olm.

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