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Mesozoic spreading kinematics: consequences for Cenozoic Central and Western Mediterranean subduction

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SUMMARY

The highly complex tectonics of Central and Western Mediterranean subduction are well documented, but there is significant debate about the responsible dynamics. The motions of the main plates, Africa including Adria, Europe and Iberia, imposed initial and boundary conditions on the evolution of subduction that are often not considered. To quantitatively evaluate these conditions, we make a set of reconstructions from Mesozoic opening through Cenozoic closing of the Alpine Tethys, using main-plate kinematic data from several authors. Geologic and tectonic information are only added to constrain the location of the break-up boundary and a single plate-margin rearrangement at the end of the opening phase. Otherwise, the plates remain undeformed. This rigid-plate approach illustrates the context in which surface deformation and subduction occurred and provides estimates (with uncertainties) of the amount of material that should be accounted for in orogens or documented seismically in the mantle. Full tectonic reconstructions should satisfy such constraints. Opening led to alternating domains of predominantly oceanic lithosphere formed by normal spreading and domains dominated by transform motion, floored mainly by extended continental lithosphere. The transform domain structures provide logical decoupling zones to allow Penninic, Ligurian and Pyrenean basins to start subducting independently. The complex buoyancy in the transform domain linking Ligurian and Penninic basins, and obliquity between directions of opening and closing may account for a number of the oceanic basins and continental slivers often invoked to explain Alpine geology. The significant proportion of continental lithosphere in the Alboran would have favoured delamination of mantle lithosphere over subduction. The almost completely subducted Penninic slab obstructed subduction of the Ligurian domain in the direction of Africa–Europe convergence, possibly forcing the rollback of the Appeninic/Calabrian trench.

Key words: lithosphere, Mediterranean, plate motions, subduction.

1 INTRODUCTION

Although often treated as 'normal', Cenozoic subduction in the Mediterranean was a lot more complex than Pacific-style subduction. Many small subduction zones with lengths of only a few 100 km to up to a 1000 km have been active below the Alboran, Pyrenees, North Africa, Calabria, Apennines, Alps, Dinarides, Carpathians and the Hellenic domain (Fig. 1). Some of these may be connected (e.g. the Calabrian and Apenninic slabs); others appear to be laterally segmented (Alpine: Lippitsch *et al.* 2003; Carpathian: Sperner *et al.* 2001; Apenninic: Amato *et al.* 1998), or have opposing dips (Alps and Apennines). Furthermore, several subduction zones developed along orientations oblique to perpendicular to the overall north–south convergence (Apenninic, Calabrian, Carpathian) and retreated during their evolution (all except Dinaride: Dercourt *et al.* 1986; Dewey *et al.* 1989; Gueguen *et al.* 1998; Jolivet & Faccenna 2000). In a number of the zones (Alps, Apennines, Alboran), substantial portions of continental lithosphere may have been subducted (Argnani 1990; Serri *et al.* 1993; Vissers *et al.* 1995; Schmid & Kissling 2000; Gattacceca & Speranza 2002).

Reconstructions proposed for the last 80–65 Ma agree on the general kinematic and tectonic evolution during Africa–Europe convergence. However, they differ in the interpretation of what dynamically drives the different subduction zones and what controls their retreat and fragmented nature. Various mechanisms have been proposed to play some role, for example, orogenic collapse (for the Apennines, Faccenna *et al.* 1997), delamination or convective removal of the lithosphere (for the Alboran and northern Tyrrhenian, Serri *et al.* 1993; Vissers *et al.* 1995; Seber *et al.* 1996;

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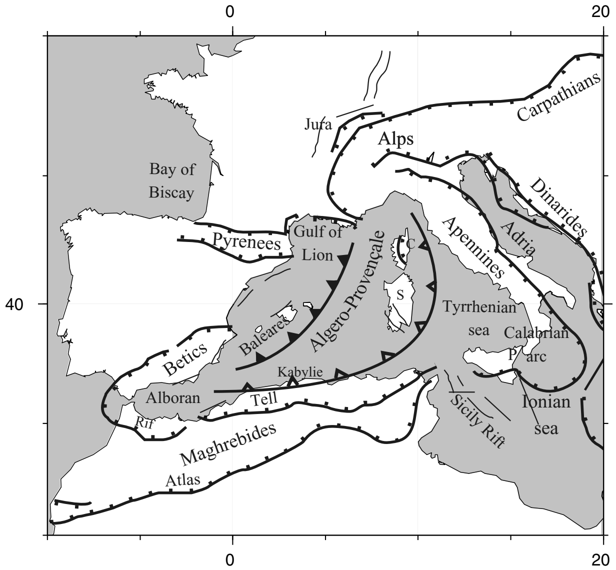

Figure 1. Tectonic map of Western and Central Mediterranean. Lines with small squares show current position of the Maghrebian-Apenninic-Alpine convergence front. Thin lines mark extensional faults. P—Peloritani block, S—Sardinia, C—Corsica. Cenozoic migration of the Maghrebian-Apenninic trench is illustrated by the lines with solid triangles (35 Ma) and open triangles (16 Ma). (based on reconstructions by Frizon de Lamotte *et al.* 2000; Gueguen *et al.* 1998; Faccenna *et al.* 2001).

Calvert *et al.* 2000; Faccenna *et al.* 2001), slab tearing (for the Alboran and African/Apenninic/Calabrian slabs, Faccenna *et al.* 2004; Goes *et al.* 2004), slab detachment and its lateral migration (for Alps, Apennines, northern Africa, Alboran, and Carpathians, von Blanckenburg & Davies 1995; Zeck 1996; Carminati *et al.* 1998; Wortel & Spakman 2000), and interaction of the slabs with the transition zone (for the Apenninic/Calabrian slab, Faccenna *et al.* 2001, 2004).

For a better understanding of the dynamics of the area, initial and boundary conditions need to be known. In particular, the shape, size and buoyancy structure of the now completely consumed basin control its tendency to subduct. These features are directly inherited from the opening phase that formed the Mesozoic Alpine Tethys (a name coined by Stampfli *et al.* 2001 for the whole set of West and Central Mediterranean basins opened during this time). Yet, previous studies of Mediterranean subduction dynamics have usually not considered the initial conditions imposed by the opening phase, and often neglect the boundary conditions that the motions of the major plates put on the overall evolution. The many kinematic and tectonic reconstructions for this area (e.g. Le Pichon *et al.* 1977; Dercourt *et al.* 1986; Savostin *et al.* 1986; Rowley & Lottes 1988; Dewey *et al.* 1989; Royer *et al.* 1992; Mazzoli & Helman 1994; Rosenbaum *et al.* 2002; Stampfli *et al.* 2002; Cavazza *et al.* 2004) basically contain this information; and a few also included considerations of possible dynamics (Schettino & Scotese 2002; Stampfli *et al.* 2002). However, they have not been used to quantify such initial and boundary conditions.

In this paper, we investigate the consequences of the motions of the large plates (Africa + Adria, Europe, and Iberia) for the geometry, size and character (oceanic or continental) of the Alpine Tethys from the main rifting in the Liassic (∼190 Ma) until present. Our emphasis is on the lithosphere. Plate motions are determined from present-to-past reconstructions. However, tracking lithospheric evolution forward in time illustrates the constraints under which crustal deformation (documented in surface geology) and mantle subduction (imaged by seismic tomography) occurred. Our aim is not to produce tectonic reconstructions, but rather to quantify the kinematic framework for them. Their level of detail is sufficient for an improved dynamic understanding. We will show that the character of the subducted Alpine Tethyan slab may have played an important role in the evolution of Central and Western Mediterranean subduction. Strong gradients in buoyancy would have facilitated slab deformation and possibly segmentation, and space problems that developed may have required slab retreat.

2 KINEMATIC RECONSTRUCTION METHOD

2.1 General approach

We focus on the Central and Western Mediterranean. Areas west of Iberia, and east of Adria, including the Carpathians, are only roughly represented. The Jurassic Alpine Tethys formed as an offshoot of the Central Atlantic (Le Pichon 1968; Pitman & Talwani 1972), in response to the sinistral transcurrent motion of the African plate with respect to the Ibero–European plate. By about 130 Ma (chron M17), motion steps into the North Atlantic, abandoning the Tethyan arm; Iberia starts to move approximately with Africa, opening the Pyrenean basin. By ∼84 Ma all opening has ceased and progressive consumption of the whole Alpine Tethys occurs, partially by active subduction and collision, partially by trench rollback (Fig. 1).

Thus, the reasonably well-constrained motions of the major plates controlled the opening and imposed the boundary conditions of its closure. We use a set of reconstructions of the motion of Africa and Iberia w.r.t. Europe to obtain estimates, with uncertainty bounds, of the Alpine-Tethys' evolution from the Jurassic until today. We do not include the motions of smaller blocks since (1) their geometry and history of formation are still under discussion, and (2) these blocks move in response to the large-scale motions of the main plates. Adria is treated as a rigid promontory of Africa throughout the evolution (see discussion in Section 2.4). Our reconstructions do not extend east of Adria, to the Balkan–Carpathian area, where block motions do play an important role.

To constrain the full size of the basins formed in the opening phase, our reconstructions start in Liassic (∼190 Ma), with the onset of major rifting. Earlier rifting contributed only minor amounts of extension (e.g. Dercourt *et al.* 1986; Froitzheim & Manatschal 1996). We reconstruct the original African, Adriatic, European and Iberian plate configuration at this time from a range of rotation poles (Fig. 2). A rifting boundary is then imposed, based on geologic and tectonic constraints. We find there is limited space for alternative boundary configurations. Subsequently, the plates are shifted according to several published sets of rotation poles. Newly formed oceanic domains are added to the major plates and then move along with them. One rearrangement of plate boundaries after the opening phase is included, as geological data show the Alpine Tethys dissected into (at least) three independent lithospheric domains: the Penninic moving with Europe, the Ligurian moving with Africa and the Pyrenean moving with Iberia. In all other phases of the evolution, the plates and boundaries are kept rigid and do not deform. This entails that boundaries are not adjusted for collisional shortening, nor are trench rollback and back-arc extension included. This rigid plate approach illustrates the context within which such boundary deformation occurred. The resulting plate overlaps at convergent

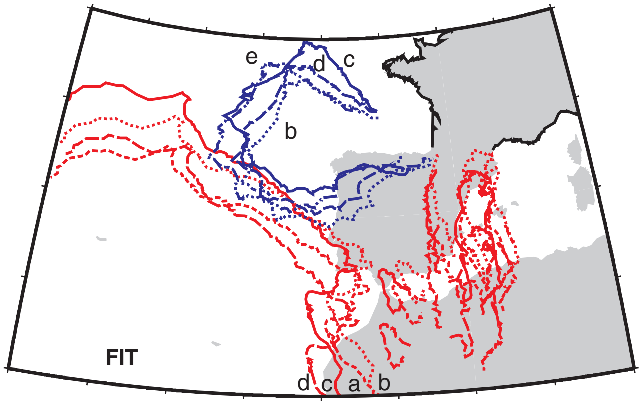

Figure 2. Plate configuration at the start of rifting. Finite rotations are relative to Europe according poles by (a) Le Pichon *et al.* (1977), (b) Savostin *et al.* (1986), (c) Rosenbaum *et al.* (2002), (d) Olivet (1996) and (e) Rowley & Lottes (1988). African domain in red, Iberian domain in blue, European domain in black (see text for discussion). Equidistant cylindrical projection, tickmarks every 5 degree. In grey shading present coastlines are shown for reference.

boundaries document how much motion must have been accommodated by either shortening, which should be consistent with the amount of material now found in an orogen, or by subduction, which should correspond to the volume of material that seismic tomography images in the mantle. Geologic constraints were only used to define the original plate boundary and its single rearrangement. By adding as few additional assumptions as possible, we evaluate the actual kinematic constraints and generate scenarios that can be verified against other geological and geophysical data.

2.2 Rotation poles

We present kinematic evolutions based on rotations poles compiled by Savostin *et al.* (1986), Royer *et al.* (1992) and Rosenbaum *et al.* (2002). In the remainder of the paper these evolutions are referred to as S86, R92 and R02. These and other kinematic reconstructions (Le Pichon *et al.* 1977; Biju-Duval *et al.* 1977; Rowley & Lottes 1988; Dewey *et al.* 1989; Ricou 1994; Olivet 1996; Hay *et al.* 1999; Stampfli *et al.* 2001, 2002; Schettino & Scotese 2002; Sibuet *et al.* 2004) are all based on the magnetic anomaly evaluations of Pitman & Talwani (1972), Srivastava & Tapscott (1986) and Klitgord & Schouten (1986). Only S86, R92 and R02 analyze the complete time span of interest for the three plates involved. They cover the range of interpretations quite well.

The oldest of the three, by Savostin *et al.* (1986), covers plate motion around the Tethys belt from Iberia to the Far East, and formed the basis of the comprehensive tectonic reconstruction by Dercourt *et al.* (1986). Royer *et al.* (1992) applied a global plate circuit method to obtain a global set of rotation poles for an updated compilation of oceanic magnetic anomalies. Recently, Rosenbaum *et al.* (2002), using a similar approach, reanalyzed the rotation poles of the plates that affected Mediterranean tectonic evolution.

The main differences between the three reconstructions, for the evolution of our study area, are the motions of Iberia. Its motion is independent in S86 and R02, faster than African motion in the first, slower in the second, whereas it is attached to either Africa or Europe, in R92. The differences in African motion from the various rotation pole sets do not affect the main characteristics of Central and Western Mediterranean plate kinematics.

In our models, Iberia and Africa wander with respect to a fixed Eurasia. Configurations are evaluated at the following Atlantic magnetic anomalies, which coincide with the main tectonic phases in the studied area (with approximate ages from Klitgord & Schouten 1986): Black Spur Anomaly (BSA) (170 Ma), M17 (131 Ma, M10N in Klitgord & Schouten 1986), M0 (118 Ma), 34 (84 Ma), 30 (67 Ma), 13 (35 Ma) and 6 (20 Ma). These magnetic stages are common to all rotation pole sets. Uncertainties in the age of the magnetic stages slightly change plate velocities, but do not affect the main characteristics of the reconstructions.

2.3 Initial plate configuration

Magnetic anomalies cannot give constraints on the pre-rifting geometry prior to the BSA. Various pre-rifting configurations have been proposed on the basis of the continental fit in Atlantic (Fig. 2). To obtain the original position of Africa, Savostin *et al.* (1986) follow Le Pichon *et al.* (1977), who estimated the stretching of the continental lithosphere prior to opening of the Central Atlantic to be about 100 km on each side. Royer *et al.* (1992) and Rosenbaum *et al.* (2002) use rifting poles from Klitgord & Schouten (1986) and Srivastava *et al.* (1990). Pre-rifting positions of the northwestern African margin from different plate motions differ by no more than 200 km (Fig. 2, and Fig. 3 FIT).

Iberia's pre-rifting position is subject to more debate. Rosenbaum *et al.* (2002) and Savostin *et al.* (1986) chose relatively tight and loose closures of the Bay of Biscay, respectively (Fig. 2). Royer *et al.* (1992) do not treat Iberia as an independent block, but move it with Europe, or between chron M0 and 13 (118–35 Ma), with Africa (following Roest & Srivastava 1991). This is a reasonable first approximation (although Iberia moved somewhat independently between M17 and 13), but it does not provide a pre-rifting position of Iberia. For R92, we put Iberia before 131 Ma at a position (dashed in Fig. 3b) approximately 100 km further west than its pre-drifting location, where it does not overlap with pre-rifting Adria, and is situated within the range of Iberia positions derived from other reconstructions. The chosen initial position for Iberia will affect its subsequent kinematic evolution.

A small amount of overlap between Africa and Iberia occurs in the Betic area (Fig. 2), but these units only formed later as a result of opening of the Alboran domain and consequent collision. Potential space problems between the Adriatic promontory and northern Iberia are discussed below.

2.4 The Adria block

The position of the Adria block plays an important role in the opening phase, affecting both the shape and size of the Mesozoic basins. It has been debated whether Adria moved as an independent block (Dercourt *et al.* 1986; Catalano *et al.* 1991; Márton & Nardi 1994; Catalano *et al.* 2001) (or two, Stampfli *et al.* 2001), or was essentially attached to Africa during most of its Mesozoic and Cenozoic evolution (Muttoni *et al.* 2001; Wortmann *et al.* 2001; Bosellini 2002; Rosenbaum *et al.* 2002). Most arguments point to the latter interpretation.

Paleomagnetic data (reviewed by Channel 1986; Lowrie 1986; Westphal *et al.* 1986; Meloni *et al.* 1997; Muttoni *et al.* 2001) does not resolve any independent rotation since Permian. Ionian spreading (which would have rotated Adria) likely occurred before (Stampfli & Mosar 1999; Catalano *et al.* 2001; Muttoni *et al.* 2001; Stampfli *et al.* 2001a; Rosenbaum *et al.* 2004).

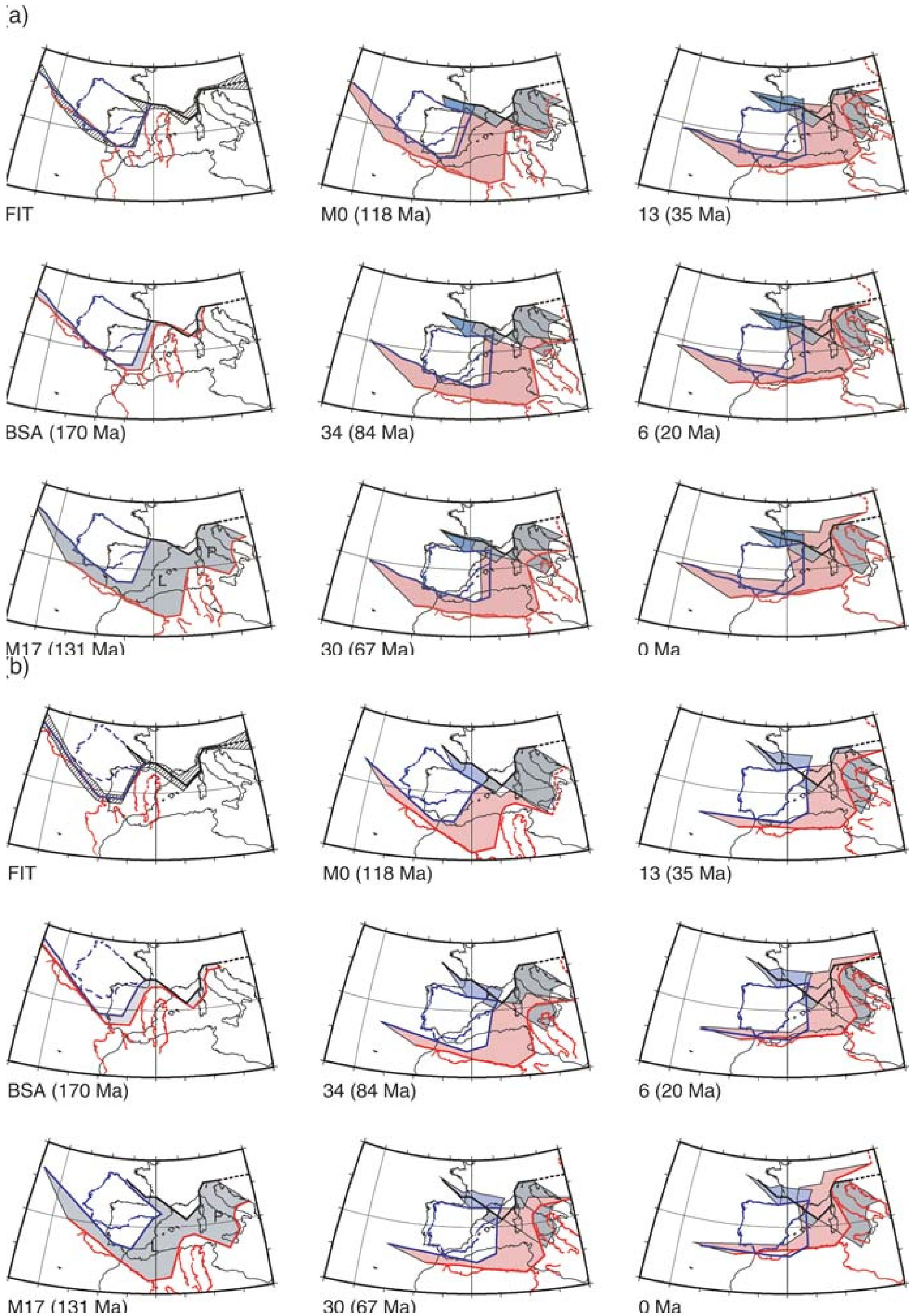

Figure 3. Plate kinematic reconstructions based on rotation poles of Africa and Iberia relative to a fixed Eurasia by (a) Savostin *et al.* (1986), (b) Royer *et al.* (2002) and (c) Rosenbaum *et al.* (2002), shown using an equidistant cylindrical projection. African domains are in red, Iberian ones in blue, and Europe in black. The dashed FIT position of Iberia is our interpretation (see text). For reference, the present-day coastlines are shown in thin black lines, with a frame with tick marks every 5 degree, and gridlines every 20 degree. Internal deformation, slab rollback and motions of smaller blocks (e.g. Corsica-Sardinia) are not included. Uncertainties in the location of the opening boundary are shown on the FIT panels (hatched). After the opening phase (FIT to 131 Ma), the created domains (shaded) are attached to one of the major plates (see text for discussion). Overlaps resulting in the convergence phase (M0—present) give an indication of the volume of material subducted and/or incorporated into orogens.

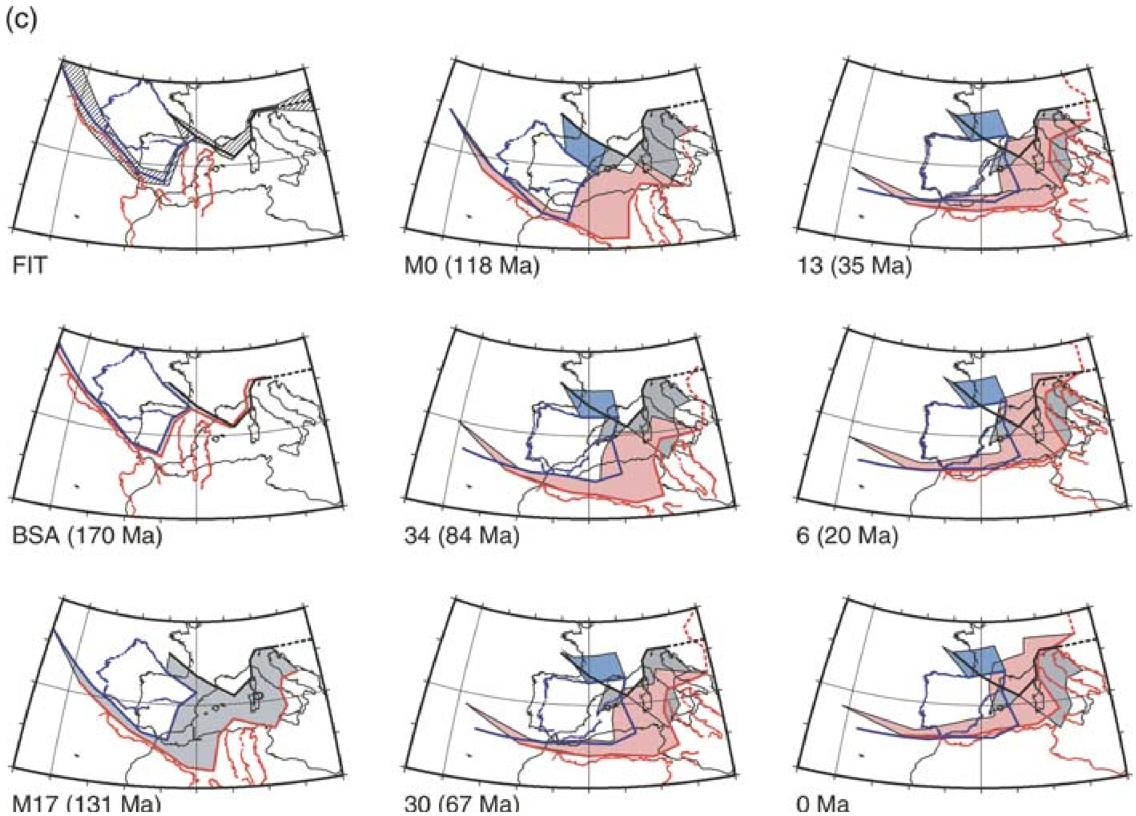

Figure 3. (*Continued*.)

An often-used argument for independent Mesozoic motion of Adria is that when reconstructed to the start of rifting, an Adriatic promontory rigidly attached to Africa would overlap with Iberia (Savostin *et al.* 1986; Stampfli & Mosar 1999; Wortmann *et al.* 2001). Whether a reconstruction yields overlap depends on several factors. (1) The assumed original size of Adria, which requires correction for Mesozoic stretching and Cenozoic shortening. The thin-skinned tectonics of the Apennines involve Triassic to recent units. Faults cutting the pre-Mesozoic basement of the Apennines have only small net offsets. This indicates that the Mesozoic stretching that formed these faults is approximately compensated by the Cenozoic shortening, which reactivated them (Calamita *et al.* 1994; Lemoine & Trümpy 1987; Catalano *et al.* 2003; Boccaletti *et al.* 1990). Assuming this was true along Adria's whole northwestern boundary, its original size is approximately the current one. This results in only minor to no overlaps with Iberia in the initial configuration (Fig. 2). (2) Rifting between Iberia and Adria probably occurred in a more distributed fashion. This would move Adria somewhat less far west than shown in the FIT reconstructions, where it is moved rigidly with Africa, further reducing potential overlap. So, although geologic data cannot preclude non-rotational motions of Adria independent of Africa since the rifting phase, no data require it and we choose the simplest option and fix Adria to Africa in our reconstructions.

3 THE OPENING PHASE

3.1 Initial pre-rifting plate boundary

The initial plate boundary geometry we use is the simplest one that satisfies the reconstructed initial positions of the continents (Figs 2 and 3 FIT) as well as the main geological and tectonic constraints.

The northern Africa boundary was a strike-slip zone from Paleozoic to Mesozoic time, approximately paralleling the present-day African margin (Dercourt *et al.* 1986; Vai 1991). The distance between undeformed Iberia and Africa, leaves only a narrow zone for the African–Iberian plate boundary (Figs 2 and 3). In the Late Cretaceous prior to Paleogene collision, reconstructions have positioned it between 100 km north of the current Tellian front and 150–200 km south of the current Betic front (Dercourt *et al.* 1986; Frizon de Lamotte *et al.* 2000; Stampfli & Borel 2002).

The Adria–Iberia boundary is constrained by paleogeographic reconstructions of the domain distribution between Adria and Iberia (Gealey 1988; Meulenkamp & Sissingh 2003; Philip 2003). On the Iberian side, the plate includes the Kabilian blocks, the Baleares, the Calabro-Peloritani block and part of Sardinia, as well as the continental lithosphere that has now stretched to form the southern Tyrrhenian (Bayer *et al.* 1973; Burrus 1984; Olivet 1996; Gueguen *et al.* 1998). West of the southern Apennines, sufficient area has to be available for the 300–400-km-wide Lagonegro and Campanian carbonate platforms, which are now stacked in the Apennines (Malinverno & Ryan 1986). Sardinia and the Calabro-Peloritani block first started colliding with such Adriatic carbonate platforms 15 Ma. (Mattei *et al.* 2002; Speranza *et al.* 2002), indicating that a western Adria boundary cannot be located much east of the current position of Sardinia.

The margin between Europe and Iberia is drawn approximately along the North Pyrenean Fault (Roure *et al.* 1989; Choukroune 1992; Verges *et al.* 2002). Adria's northern margin is constrained by restoration of the southern Alpine domain (Lemoine & Trümpy 1987) and on the European side, by the stable part of the Eurasian plate. Tectonic reconstructions accounting for crustal volume locate the European margin around 400 km southwest from the current Jura front (Lemoine & Trümpy 1987; Trümpy 1988; Ménard *et al.* 1991).

South of the French Provençal coast, our European margin includes space for most of Corsica and part of Sardinia (Bayer *et al.* 1973; Burrus 1984; Olivet 1996; Faccenna *et al.* 2002). This interpretation is consistent with a structural analysis (Trümpy 1988) that constrains the European margin east of Corsica to have a north–south trend then turning tightly westward around Corsica and Sardinia and entering the Pyrenean area. The initial configuration contains no space for these blocks on the Iberian side. An Iberian origin and late Mesozoic docking of Corsica and Sardinia to Europe (Stampfli *et al.* 2002) would require that either the space for these blocks formed during Mesozoic stretching (in the 'Piemont' domain, discussed below), or Adria was not rigid (Stampfli & Mosar 1999). In this case a straighter Provençal boundary is possible (Fig. 3 uncertainties in initial boundary).

On the southern side of the Alpine European margin lie the Austroalpine, Dinaride and some Pannonian and Carpathian domains (see Dercourt *et al.* 1986; Olivet 1996; Wortmann *et al.* 2001; Stampfli *et al.* 2002). These domains roughly travelled with Adria, although small-block interactions and intra-plate deformation complicate their motion (Balla 1982; Csontos & Vörös 2004). We treat them as a rigid part of Adria until 130 Ma, to define the eastern boundary of the Penninic. After 130–110 Ma, they start to behave more independently, where the Carpathian domains are extruded east over a significant distance, and we do not attempt to reconstruct their evolution.

There is little space to draw a significantly different initial rift boundary (±100–200 km in most places, Fig. 3 FIT configurations). In detail, the margin was probably more complex, but this does not affect the main features of our reconstruction. Therefore, we show uncertainties in the initial boundary in Fig. 3, but do not propagate them through our analysis. We now examine the kinematic consequences of this initial boundary configuration.

3.2 Opening phase: FIT to M17-34

3.2.1 Basin shape and size

During this phase, the African domain shifts southeastward with respect to the Ibero–Eurasian plate. Early motion results in continental rifting between the eastern Iberian margin and western Adriatic domain, mainly transform motion in the Alboran and along the French part of the European margin (the Provençal) and a small amount of extension in the future western Alpine domain. Subsequent fast extension between the BSA (170 Ma) and M17 (130 Ma) led to the opening of the Alpine Tethys in which two main extensional basins are formed: the Penninic to the northeast and the Ligurian to the southwest, offset by the Adriatic promontory. These basins alternate with narrower, transtensional, domains, the Alboran realm between Iberia and northern Africa, a domain connecting the Ligurian and Penninic basins, which we will call Piemont, and the more recently opened Pyrenean Trough.

The characteristics of these basins are similar in all three reconstructions. The Alboran domain is a 100–200-km-wide corridor extending all along the northern African margin and Iberian southern margin; the Ligurian and the Penninic measure 800–1100 km and 650–800 km in the direction perpendicular to spreading and 1200–1550 km and 850 km lengthwise, respectively, while the width of the intervening Piemont is 200–300 km, measured perpendicular to the transform motion. The width of the Piemont domain that is formed depends on the detailed evolution between 170 and 110 Ma, but it is never wider than about 300 km. Sinistral transform motion in this time interval amounts to 900–1000 km in the Alboran domain and 800–1000 km in the region connecting Ligurian and Penninic. This offset between the Ligurian and Penninic is controlled by the Adriatic promontory, and would be changed if Adria moved independently of Africa (Dercourt *et al.* 1986; Stampfli *et al.* 2002). In the Pyrenean Trough, the kinematics yield a 160 to 380-km-wide basin with an 80–400 km sinistral transform component.

With the exception of the Pyrenean Trough, the Alpine Tethys' basins reach their maximum size by about 130 Ma (M17), with only minor adjustments of up to chron 30. In S86, the Pyrenean domain does not open until after 130 Ma (M17), whereas in R92 and R02 opening starts during or shortly after rifting (Fig. 3). In all reconstructions, the trough attains its maximum area by anomaly 34. These differences in Iberia motion result in different basin sizes. The absence of pre-M17 Iberia motion in S86 leads to a 30 per cent larger Ligurian and factor-two larger Alboran basin than in the other two reconstructions (Fig. 4). Instead, R92 and R02 create a 20–40 per cent larger Pyrenean basin and assign a larger portion of the spreading to the Atlantic. Alternative plate boundaries do not significantly change the area or the main configuration, but may somewhat change basin shape, for example, the S-shape of the Penninic is straightened if the European margin contains less of a corner to accommodate Corsica-Sardinia blocks. A somewhat different Africa–Europe motion leads to an almost 50 per cent larger Penninic domain in the case of R92. The differences between the reconstructions yield reasonable estimates of the uncertainty in basin size.

3.2.2 Nature of the basins

Age of the oceanic lithosphere and the area and amount of stretching of the continental lithosphere determine the buoyancy of the basins, and thus their tendency to facilitate or resist subduction. Based on their orientation w.r.t. the direction of spreading, the Ligurian and Penninic basins were probably mostly underlain by continuous oceanic lithosphere, centred around a spreading ridge with short transforms. Age of this oceanic lithosphere ranged from 0 to 40 Myr over the half width of the basin. The spreading rates were around 2 cm yr^{-1}. By contrast, the intervening Alboran and Piemont domains had an orientation almost parallel to the direction of spreading. In these basins, short spreading ridges (or pull-apart basins) would have been offset by long transform faults. These domains were likely characterized by large buoyancy gradients, as the transforms would juxtapose 40 Myr old against newly formed oceanic lithosphere or (extended) continental lithosphere.

Using geologic constraints, we added estimates of the basin area generated by continental rifting to Fig. 4. The inner Alboran, extending from Ligurian to Gibraltar, was probably almost completely underlain by thinned continental lithosphere. When restored to their original size, the Subbetic and the Rifan–Tellian continental extensional platforms have been estimated to occupy a region 200–250 km wide (Hay *et al.* 1992; Frizon de Lamotte *et al.* 2000; Platt *et al.* 2003). This is about the width of the Alboran domain in the reconstructions, implying that most of this extension was achieved by continental rifting. It is possible that some of the continental lithosphere was very strongly extended (by factors exceeding 1.5– 2), reducing crustal thicknesses to 10–20 km. The Atlantic part of this domain was probably mostly oceanic.

We estimate the amount of stretched continent along the borders of the Ligurian and Penninic from the area generated before the

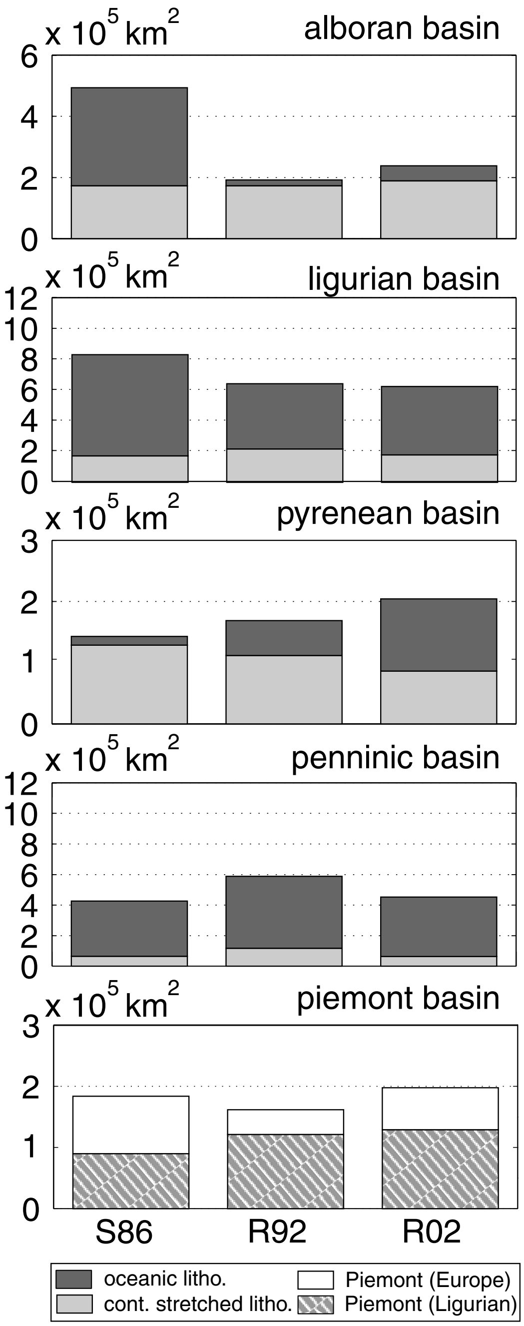

Figure 4. Maximum Alpine Tethys subbasin areas after the opening phase. Estimates on the proportion of stretched continent for the Alboran and Pyrenean basins are based on Mesozoic basin reconstructions by Choukroune (1992), Verges *et al.* (2002), Frizon de Lamotte *et al.* (2000) and Dercourt *et al.* (1986). Continental estimates in the Ligurian, and Penninic basins are constrained by the area produced in our reconstruction between the start of major rifting and the start of break-up at the BSA. The Piemont domain was probably mostly continental. In detail, complex rifting may have somewhat altered the basins' shape, but the comparison between the three reconstructions gives an idea about the uncertainties in basin size and proportions of continental and oceanic lithosphere.

BSA (Fig. 3), that is, before oceanization started. This constitutes between 15 and 25 per cent of the total basin area (Fig. 4). Extension factors probably ranged between 1.1 and 1.5 (Wooler *et al.* 1992). Note that our estimates are for the total area generated by continental extension. Stretching may well have been distributed over several parallel zones, leaving a complex geologic record of alternating continental slivers and marine basins.

The intervening Piemont domain probably was largely underlain by (very stretched, Froitzheim & Eberli 1990; Ménard *et al.* 1991) continental lithosphere, like the Alboran. Depending on how strain in this region localized, it could have contained various smaller basins and higher standing blocks.

Reconstructions of the continental part of the Mesozoic Pyrenean Trough show it to have been less than 160–170-km wide (Choukroune 1992; Verges *et al.* 2002). This leaves no room for oceanic lithosphere in S86 (except a small oceanic part in the Bay of Biscay), whereas the other reconstructions require significant portions of ocean floor.

4 THE CLOSING PHASE

4.1 New plate boundaries

By the end of chron M17, the opening motion dies out in the Alpine Tethys and steps into the North Atlantic. The main plate boundary is transferred from the Alboran to the Pyrenean domain where the Gulf of Biscay and Pyrenean Trough open, accompanied by significant translation. The southern Alpine Tethys (Ligurian/Alboran) then only accommodates the minor Iberia–Africa motion. The Alpine Tethys north of Iberia (Penninic) starts to subduct below the north end of the Adriatic promontory (Dercourt *et al.* 1986), while simultaneously some last east-west extension occurs.

These changes require new plate boundaries. What controls where subduction initiates is subject to debate, but existing zones of weakness, e.g. transform faults or continent-ocean boundaries, are probable candidates (Cloetingh *et al.* 1982; Hall *et al.* 2003). We choose a relatively simple option. The new plate boundary is imposed in two time steps (Fig. 3). (1) At the end of M17 (131 Ma), the Alpine Tethys is dissected, the Penninic attached to Europe and the Ligurian/Alboran to Africa. (2) At the end of M0 (118 Ma), the Pyrenean domain is attached to Iberia. The location of this boundary is motivated by the subduction vergence documented in the various mountain belts: northward dipping below the Pyrenees, northwestward dipping below the Magrebian–Apenninic belts and southeastward dipping the western and central Alps (Dercourt *et al.* 1986; Choukroune 1992; Gueguen *et al.* 1998; Verges *et al.* 2002; Stampfli *et al.* 2002; Lippitsch *et al.* 2003).

Most of the new plate boundaries are positioned at ocean-continent borders. The Piemont domain forms a natural decoupling zone for the Penninic and Ligurian basins, with zones of weakness (transform faults and pull-apart basins) of a favourable orientation. Within this zone, various plate boundaries can be drawn. We attach the westernmost part to the European continental margin, which it adjoins, and the eastern part to the north of nearby Adria. This solution provides extra continental area (not volume) that may account for some proposed terranes, e.g. Briançonnais or western Alpine terranes, and possibly part of Corsica and Sardinia, if their later docking to Europe is preferred (Lacombe & Jolivet 2005; Stampfli *et al.* 2002). The Pyrenean domain is separated from the European part of the Piemont along an extension of the Iberian margin.

More detailed study may reveal that further adaptation of the boundary is necessary to satisfy all data, but is beyond the scope of this paper. For example, small holes developing between some of the plates after M17 and M0 (Fig. 3) might be avoided with more spatio-temporal complexity of the boundary.

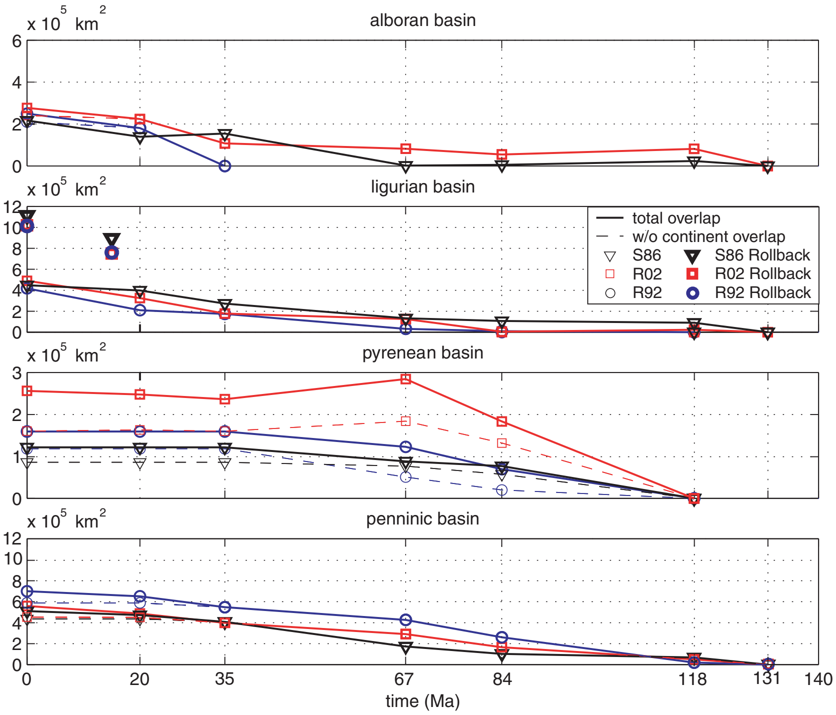

Figure 5. The evolution of overlap of each of the Alpine Tethys' subbasins with Europe (including part of the Piemont), Iberia, and other basins. All overlap is assigned to the presumed lower plate. Overlap is accommodated by either subduction or orogenic shortening. Ligurian and Penninic evolutions are characterized by relatively constant active consumption rates of 0.2–1·10^4 km^2 Myr^{-1} The bold symbols in the Ligurian panel indicate the extra area overridden by trench rollback. The reconstructions show more differences in timing of Alboran and Pyrenean overlap evolution, but agree on most of the Pyrenean closure to have occurred before 67 Ma, whereas the main closure of the Alboran occurs after 67 Ma.

4.2 Closing reconstruction, M17–0 (131–0 Ma)

Between M17 and chron 34, opening slowly dies out and subduction starts. During the relatively rapid (up to 1.5 cm yr^{-1}) Europe–Africa convergence between chron 34 and 13, Penninic and Pyrenean lithosphere are subducted, leading to collision by 35 Ma. From 35 Ma onwards, only the southern remnants of the Alpine Tethys, the Ligurian and Alboran domains, are left. Their subduction in response to the slow Africa–European plate motion is restricted by Alpine and Pyrenean collision. Ligurian subduction is finally achieved by fast trench rollback in two phases (Fig. 1), opening the Algero-Provençal basin between 35 and 20 Ma, and after stalling for about 10 Myr, the Tyrrhenian basin (8–0 Ma) (Burrus 1984; Rehault *et al.* 1984; Faccenna *et al.* 1997; Gueguen *et al.* 1998; Sowerbutts & Underhill 1998; Jolivet & Faccenna 2000). In the time frame 25–0 Ma, Alboran crustal units are thrust westward over the Iberian and African margins, while its lithosphere founders, opening the westernmost end of the current Algerian basin in their wake (Lonergan & White 1997; Gueguen *et al.* 1998; Frizon de Lamotte *et al.* 2000; Faccenna *et al.* 2004).

Our reconstructions illustrate how much of the basins disappeared through active convergence. We estimate the rate of basin consumption from the progressing overlap between each of the basins and the surrounding continents (Figs 5 and 6). All overlap is assigned to the lower plate, e.g. Ligurian–Penninic overlap is assigned to the Penninic, which is the first to start subducting, but overlap between the Ligurian and the Provençal part of the Piemont is assigned to the Ligurian. Note that this estimate of basin consumption does not imply anything about the mechanism by which shortening was achieved.

4.2.1 Alboran

As early as 131–118 Ma, the Alboran responds to the several tens of kilometrs of northeast–southwest convergence in the reconstructions where Iberia moves independently. However, most of the basin's shortening is achieved around (S86) or after 35 Ma (R92 and R02) (Fig. 5), with peak rates of up to 8·10^3 km^2 Myr^{-1}. The total amount of convergence is similar in all reconstructions (Fig. 6), but in S86, where the largest Alboran domain was created, only the eastern (largely continental) part of the basin disappears. In R02 and R92, the basins were completely consumed, and since 20 Ma, a small area of pre-Mesozoic basement was involved (Fig. 6). Although this African overlap can be somewhat reduced by modifying the initial boundary, it is difficult to completely avoid. A comparison with continent/ocean proportions after opening (Fig. 4) shows that 80–95 per cent of the overlapped basin was continental.

4.2.2 Ligurian

By moving east beyond its present-day position by 84 Ma, Iberia overrides the Ligurian oceanic lithosphere (S86 and R02), over a distance of about 150–200 km. By chron 13 (35 Ma), the northward underthrusting amounts to 100–200 km and 20–30 per cent of the Ligurian basin is overridden by Europe (including the Provençal part of the Piemont) and Iberia (Fig. 5). Subsequently, the trench rolls back, overriding the remainder of the Ligurian basin in the first stage, and the area east of it in the second. The total basin overlap in Fig. 5 illustrates the amount of active convergence, which is achieved with an almost constant rate of 8·10^3 km^2 Myr^{-1}.

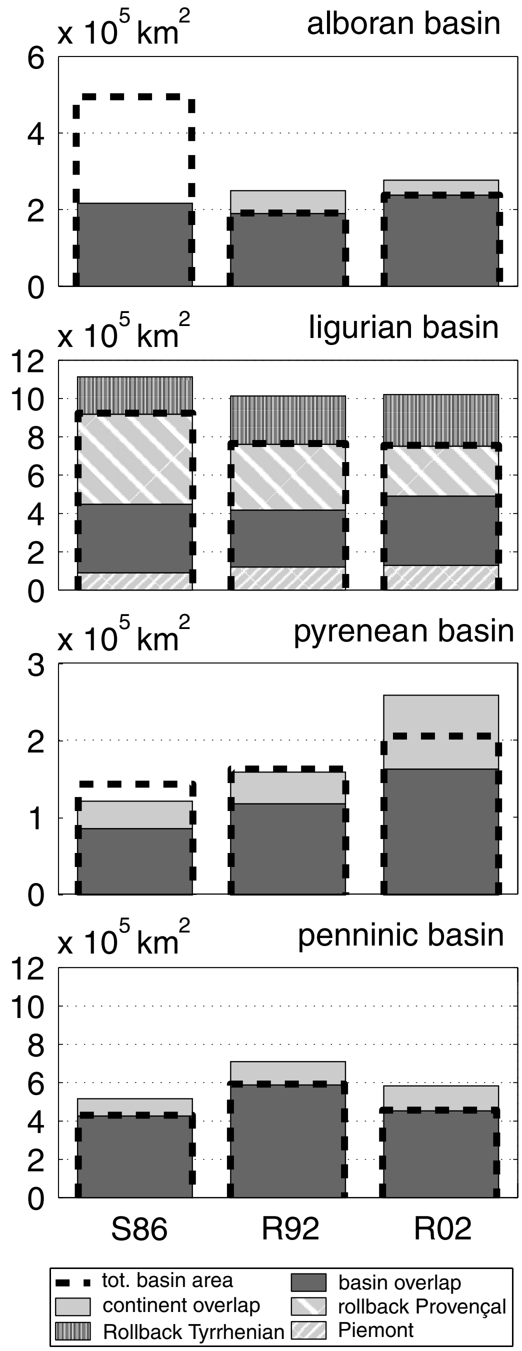

Figure 6. Total amount of active overlap of the Alpine Tethys subbasins' at 0 Ma. This includes only overlap due to main-plate convergence, in each case assigned to the lower plate. In most of the reconstructions, all the basins (comprising oceanic and extra continental lithosphere produced in the opening phase) have been completely consumed. The continental overlap consists of additional shortening that affected the bordering continents. Only 20–30 per cent of the lithosphere that disappeared below the Apennines could have been consumed by Africa–Europe convergence; the rest disappeared during two stages of trench rollback. In S86, the Atlantic part of the Alboran remains open.

Rollback more than doubles these consumption rates. By 0 Ma, 40–60 per cent of the original Ligurian has disappeared by active convergence. The eastern half of the Piemont basin, although moving with the Ligurian, was incorporated in the Alps. Its relative buoyancy caused it to override Europe, rather than underthrusting it.

4.2.3 *Pyrenees*

By the time of maximum opening of the Pyrenean Trough (118–84 Ma), closure at constant rates between 3 and $5{\cdot}10^3$ km^2 Myr^{-1} is already ongoing (Fig. 5). All evolutions show a drastic reduction of convergence after 67 Ma. In our reconstructions, the Pyrenean basin underplates Europe as well as the attached Provençal part of the Piemont domain (Fig. 3). The European–Provençal margin also overlaps the northeastern corner of unstretched Iberia (Fig. 6). Even with alternative initial plate boundaries (e.g. complexer within the Piemont domain) such an overlap cannot be avoided. The eastward widening of the overlap is consistent with the geometry of the Pyrenean orogen (Choukroune 1992; Verges *et al.* 2002). The involvement of the Provençal domain agrees with the observation that Pyrenean deformation propagated into Sardinia, at that time part of the European–Provençal margin (Faccenna *et al.* 2002), and Gulf of Lion (Lacombe & Jolivet 2005). Between 15 and 30 per cent of our reconstructed area of the Pyrenean basin (which only covers a small part of the Bay of Biscay) remains unsubducted.

4.2.4 *Penninic*

Initial convergence in the Alpine domain propagates from east to west, reaching the Western Alps by ~84 Ma. By this time, 300–400 km of Penninic lithosphere has gone down below Adria's northern margin. While the eastern part of the basin subducts below the continental margin, the western trench section is intraoceanic.

Post-M17 east–west extension and north-south convergence are compatible with the timing of extrusion and subsequent onset of collision in the Carpathian blocks (Balla 1982; Csontos & Vörös 2004). This consistency illustrates the importance of the boundary conditions imposed on the area by Africa–European kinematics. However, because further treating the eastern section of Adria may be incorrect, we dash its outline in Fig. 3 and do not add lithosphere that may have formed since M17 to the Penninic.

During the Paleocene to Eocene (65–35 Ma), the Penninic oceanic basin is almost entirely subducted under the Adriatic promontory. Active consumption rates are similar in all three reconstructions, with convergence velocities ~10 mm yr^{-1} with peaks around 14 mm yr^{-1}. Convergence turns into eastward progressing collision, when the Piemont domain starts to overlap the stretched and then the undeformed European margin around 35 Ma. From this moment onward, consumption rates lower from $2–11{\cdot}10^3$ km^2 Myr^{-1} to $2–6{\cdot}10^3$ km^2 Myr^{-1}. Today, convergence has consumed an area 20–25 per cent larger than that of the original Penninic basin (Fig. 6).

5 DISCUSSION

The structure of the Alpine Tethys, comprising two mainly oceanic basins flanked by narrow transtensional domains, has interesting implications for the subsequent subduction dynamics. Although such a structure has been drawn in other reconstructions, its consequences have not been fully recognized, because previous reconstructions did not focus on the lithosphere. Below we discuss some of the possible implications of the geometry and buoyancy structure of the Mesozoic Alpine Tethys for the Cenozoic subduction.

5.1 Effect of basin geometry

The offset between the two main basins forms a natural decoupling zone between the Alpine and Apenninic systems, and an obvious location for the reversal of subduction polarity (southeastwards in the Penninic vs northwestwards in the Ligurian). The offset in the southern Ligurian favours a decoupling of the Alboran and Ligurian (Apenninic) subduction systems. Further offsets at the eastern end of the Penninic could have allowed separate evolutions of the west–central and eastern Alpine systems as indicated by the different orientations of the seismically imaged slabs (Lippitsch *et al.* 2003). The subduction of two main basins is consistent with the two concentrations of high seismic-velocity material, one in the transition zone under the Tyrrhenian and another one south of the Alps (Piromallo & Morelli 1997; Bijwaard *et al.* 1998).

Some time around 65 Ma (somewhat variable between reconstructions), a space problem develops, as the Ligurian slab attempts to subduct below the European margin to which the Penninic slab is attached. This overlap, together with starting continental collision, may have set the stage for detachment of the Penninic slab. Mantle melt with non-subduction characteristics, and an H*T* overprint on the exhumation path of the H*P* bodies have been associated with slab detachment starting around 45 Ma in the western Alps and propagating northwestward (von Blanckenburg & Davies 1995; Davies & von Blanckenburg 1995; Wortel & Spakman 2000; Brouwer *et al.* 2004), approximately tracking the progression of interaction with the Ligurian slab. The short continental Alpine slabs imaged by Lippitsch *et al.* (2003) accumulate in the phase after detachment of the main basin. Mantle flow around the interacting Ligurian and Penninic slabs could have caused the spiralling geometry of the Alps–Apennines intersection, and would have added to the complex evolution of the western Alps.

5.2 Effect of basin buoyancy

5.2.1 Alboran

Alboran deformation has been attributed to slab rollback (Royden 1993; Lonergan & White 1997), slab detachment or lateral slab segmentation (Blanco & Spakman 1993; Carminati *et al.* 1998; Faccenna *et al.* 2004), to orogenic collapse due to convective removal of the mantle lithosphere (Platt & Vissers 1989; Vissers *et al.* 1995; Platt *et al.* 2003) or to asymmetric delamination (Docherty & Banda 1995; Seber *et al.* 1996; Comas *et al.* 1999; Calvert *et al.* 2000). We inferred that the Mesozoic Alboran lithosphere was dominantly composed of thinned continent, which would have resisted subduction. Furthermore, it was a narrow basin, measuring only around 200 km in width, with subsequent convergence closing it in its narrowest direction. Dynamic models have shown that about 200 km of convergence is necessary before subduction is self-sustaining (e.g. Funiciello *et al.* 2003). These arguments make it unlikely that the Alboran contained any active subduction zone. The eastward dipping lithosphere imaged down to the transition zone (Calvert *et al.* 2000; Gutscher *et al.* 2002; Faccenna *et al.* 2004) speaks in favour of delamination. The increase in basin consumption that occurs in our reconstructions at around 30 Ma, might have been the trigger of the foundering of Alboran lithosphere. The reconstructions also show that collision started in the east, favouring an initiation of detachment there, consistent with the eastward dip, and westward propagation of thrusting in the Betic-Rif-Tell chain, with contemporaneous extension and volcanism in the internal domains (Vissers *et al.* 1995; Saadallah & Caby 1996; Faccenna *et al.* 2004).

5.2.2 Pyrenean

The character of the Pyrenean Trough varies strongly in the three different reconstructions. R92 and R02 produce a 180–210-km-long oceanic slab followed by 160–200 km of continental convergence, whereas S86 yields continental convergence only. Even the longest possible oceanic slab is barely long enough to result in self-sustained subduction, and the subsequent arrival of continental lithosphere at the trench would have further hampered active subduction. Forced underthrusting (and wrenching) is probably a better description of Pyrenean dynamics. Some oceanic lithosphere at the basin's leading edge might have facilitated developing an asymmetric subduction-like geometry (as imaged by Sibuet *et al.* 2004).

5.2.3 Piemont

The nature of the Piemont domain may provide an explanation for many of the ophiolitic complexes and continental slivers that are now incorporated in the Alps (e.g. the Briançonnais ribbon continent and the Valais ocean), Corsica, Calabria and the Apennines (e.g. Jolivet *et al.* 1991; Steffen *et al.* 1993; Malavieille *et al.* 1998; Rubatto *et al.* 1998; Padoa *et al.* 2001; Rossetti *et al.* 2001). Domain complexity and the obliquity between opening and closing directions may well alleviate most of the need to invoke separate oceanic basins and different generations of subduction zones. Note that our split of the Piemont domain joins rifted continental margin and adjacent and possibly intervening small ocean basins to both Europe and Adria. These characteristics may also facilitate ophiolite formation (Dewey 2003; Dilek 2003). Again, it should be borne in mind that the continental extension was not necessarily concentrated around the borders of our domains, but may have been distributed further inland, along preexisting weaknesses, potentially leaving regions closer to the margin relatively unstretched.

5.2.4 Ligurian

The initial conditions under which Ligurian subduction develops do not explain the initiation of rollback. By 35 Ma, a ~200 km long Ligurian slab had accumulated below the Provençal margin, just marginally long enough to start sinking under its own weight. Even less lithosphere had been thrust below Iberia. The preferred direction of Ligurian subduction would have been in the direction of Africa–Europe convergence (as shown in Fig. 3). We suggest that blocking of northward subduction of the Ligurian lithosphere by the already subducted Penninic basin left no other choice than trench retreat if subduction was to occur. Even detached, the Penninic slab requires several tens of millions of years to clear the way. Without trench rollback, subducted Ligurian material should have ended up at depth in a position close to where we draw it in the reconstruction (Fig. 3). Rollback deposited it where tomography images this material, south of the western Alpine front and east of Corsica and Sardinia (Piromallo & Morelli 1997; Carminati *et al.* 1998). For slab retreat to occur, the Adriatic part of the Piemont domain must have decoupled from the sinking Ligurian basin, as it had decoupled earlier on from the Penninic Ocean to the north. It was then incorporated into the Alps.

Once initiated, further trench rollback would have been driven by the lengthening slab as well as flow around its edges. Rollback stalled around 16 Ma when Sardinia and Corsica reach their current position (Faccenna *et al.* 2002; Speranza *et al.* 2002), which coincides with the eastern edge of our reconstructed Ligurian basin (see also, Dewey *et al.* 1989; Gattacceca & Speranza 2002; Stampfli *et al.* 2002). Eventually, the trench retreats further over extended Adriatic continental lithosphere and the western tip of the (oceanic) Ionian basin (Dewey *et al.* 1989; Gattacceca & Speranza 2002; Stampfli *et al.* 2002). The heterogeneous nature of the slab probably facilitated the formation of slab detachments and lateral tears (Carminati *et al.* 1998; Wortel & Spakman 2000; Faccenna *et al.* 2004; Goes *et al.* 2004).

6 CONCLUSIONS

We presented a set of kinematic reconstructions that document the basic constraints that the known large-scale plate motions pose on the tectonic evolution of the West and Central Mediterranean. Three different sets of rotation poles for the motions of Africa/Adria and Iberia relative to Europe are used to trace the evolution from the Mesozoic opening of the Alpine Tethys thru its Cenozoic closing. Tectonic and geologic data constrain the initial break-up boundary, and one plate-boundary reorganization at the start of closing. We kept the amount of non-kinematic information to a minimum to be able to isolate plate-kinematic effects and assess their uncertainties. By not allowing any plate deformation, the kinematic evolutions track all lithospheric material involved in collision and subduction.

Understanding these broad-scale constraints will help in unravelling the next level of complexity (i.e. crustal deformation patterns, motions of smaller blocks and slab rollback) and in producing complete tectonic reconstructions. These kinematic reconstructions are already sufficiently detailed though for various geophysical investigations. The obtained quantitative estimates of shortened area can be compared with orogenic volumes, as well as with the amount of subducted material imaged by seismic tomography (e.g. following de Jonge *et al.* 1993; Schmid *et al.* 2002, who included resolution assessments). The initial and boundary conditions emerging from the kinematic reconstructions will facilitate dynamic studies. We discussed how the size, shape and nature of the Alpine–Tethys' basins might have played a major role in the complex dynamic evolution of Mediterranean subduction.

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