

History of subduction and back-arc extension in the Central Mediterranean

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Accepted 2001 January 19. Received 2001 January 15; in original form 2000 June 16

SUMMARY

Geological and geophysical constraints to reconstruct the evolution of the Central Mediterranean subduction zone are presented. Geological observations such as upper plate stratigraphy, HP–LT metamorphic assemblages, foredeep/trench stratigraphy, arc volcanism and the back-arc extension process are used to define the infant stage of the subduction zone and its latest, back-arc phase. Based on this data set, the time dependence of the amount of subducted material in comparison with the tomographic images of the upper mantle along two cross-sections from the northern Apennines and from Calabria to the Gulf of Lyon can be derived. Further, the reconstruction is used to unravel the main evolutionary trends of the subduction process. Results of this analysis indicate that (1) subduction in the Central Mediterranean is as old as 80 Myr, (2) the slab descended slowly into the mantle during the first 20–30 Myr (subduction speeds were probably less than 1 cm year⁻¹), (3) subduction accelerated afterwards, producing arc volcanism and back-arc extension and (4) the slab reached the 660 km transition zone after 60–70 Myr. This time-dependent scenario, where a slow initiation is followed by a roughly exponential increase in the subduction speed, can be modelled by equating the viscous dissipation per unit length due to the bending of oceanic lithosphere to the rate of change of potential energy by slab pull. Finally, the third stage is controlled by the interaction between the slab and the 660 km transition zone. In the southern region, this results in an important re-shaping of the slab and intermittent pulses of back-arc extension. In the northern region, the decrease in the trench retreat can be explained by the entrance of light continental material at the trench.

Key words: back-arc extension, mantle convection, Mediterranean, subduction, tectonics, tomography.

1 INTRODUCTION

Most of our knowledge about the subduction process comes from present-day data, such as earthquake hypocentres which trace out snapshot images of slabs. The long-term evolution of subduction is far more uncertain, yet it can provide important insights and help us understand the interaction between tectonic activity and mantle convection. In order to unravel the history of subduction, it is necessary to integrate three different sets of data: tomography images of the mantle, which give a picture of the distribution of cold subducted material; plate tectonic reconstruction, which gives the rate of convergence at plate boundaries; and geological data, which constrain the age and

style of subduction. Particularly noteworthy geological signals of the subduction process are the development of dynamic topography at the trench (e.g. Stern & Holt 1994; Royden 1993), deep burial of crustal rocks and the formation of the arc magmatism (e.g. Jacob *et al.* 1976). In addition, we can also expect uplift of the upper plate margin during subduction initiation, followed by its rapid subsidence due to viscous coupling after the downwelling has picked up speed (e.g. Mitrovica *et al.* 1989; Gurnis 1992), and when the slab has reached the 660 km discontinuity, impedance to flow will be accompanied by a decrease in subduction velocity and trench migration rate (e.g. Christensen & Yuen 1984; Kincaid & Olson 1987; Zhong & Gurnis 1995). All of these dynamic effects will leave geological

fingerprints that can be interpreted and used for a reconstruction. Such features include large-scale unconformities and marine depositions on the upper plate accompanying the initiation of subduction (e.g., Cohen 1982; Mitrovica *et al.* 1989; Gurnis 1992), siliciclastic over hemipelagic deposits in the trench area, high pressure/low temperature metamorphism (HP/LT, Peacock 1996) and spatial, temporal and chemical patterns of the volcanic arc (Jacob *et al.* 1976; Keith 1978), to name a few.

Here, a complete reconstruction of the history of subduction in the Central Mediterranean is presented. This history uses a variety of geophysical and geological data and is tied into a consistent geodynamic framework explaining the evolution of subduction with a simple model of confined mantle convection (Fig. 1). The subduction zone examined, presently located along the Apennine belt from the Calabrian arc to the northern Apennines, was oriented roughly parallel to the direction of relative plate convergence during its whole lifetime (Dewey *et al.* 1989; Patacca *et al.* 1990). Therefore, the rate of convergence at the trench was very low and the consumption of the oceanic lithosphere was probably mostly driven by its negative buoyancy, resulting in trench rollback (Malinverno & Ryan 1986; Royden 1993; Giunchi *et al.* 1996; Faccenna *et al.* 1996). This absence of significant plate convergence and the availability of geophysical and geological data from decades of research make the Central Mediterranean a prime candidate to unravel the evolution of subduction zones.

The kinematics of the Central Mediterranean subduction zone will be reconstructed from its initial stage (Upper Cretaceous, ~90–80 Ma) to the present-day along two cross-sections. The first, southern section runs from Calabria via Sardinia to the Gulf of Lyon, while the second, northern section runs from the northern Apennines via Corsica to the Gulf of Lyon (Fig. 1). The constraints we use are present-day images of the slab as inferred from recent seismic tomography and geological data,

both from the literature and from this study. The model presented here divides the evolution of subduction in the Central Mediterranean into three stages: (1) slow initiation, characterized by very low subduction speeds, (2) slab development with a roughly exponential increase of subduction speed and opening of a first back-arc basin; and (3) slow-down due to interaction of the slab with the 660 km transition zone.

On the southern section, which is dominated by subduction of oceanic lithosphere, the last phase is characterized by a transient slow-down of subduction, followed by an increase in subduction speed and a second phase of back-arc spreading. On the northern section, there is a permanent decrease of subduction speed that is most likely to have been caused not only by a stagnant slab but also by continental material entering the trench from 30 Ma onwards.

2 TOMOGRAPHIC IMAGES OF THE UPPER MANTLE

Present-day subduction in the Central Mediterranean shows up in the Southern Tyrrhenian Sea as a continuous, NW-dipping Wadati–Benioff zone (Isacks & Molnar 1971; Giardini & Velonà 1988). Seismicity is distributed in a continuous 200 km wide 40–50 km thick volume, plunging down to a depth of ~450 km with a dip of 70° (Selvaggi & Chiarabba 1995). Focal mechanisms show that the slab is, at least in its middle portion (165–370 km), subjected to in-plane compression, whereas the stress state is heterogeneous in its upper portion (Frepoli *et al.* 1996). Active volcanic arcs related to the subducting slab are localized in the Eolian and Neapolitan volcanic district.

Fig. 2 shows images from a recent tomographic model of the upper mantle beneath the Italian Peninsula by Lucente *et al.* (1999). Their inversion was performed using a high-quality

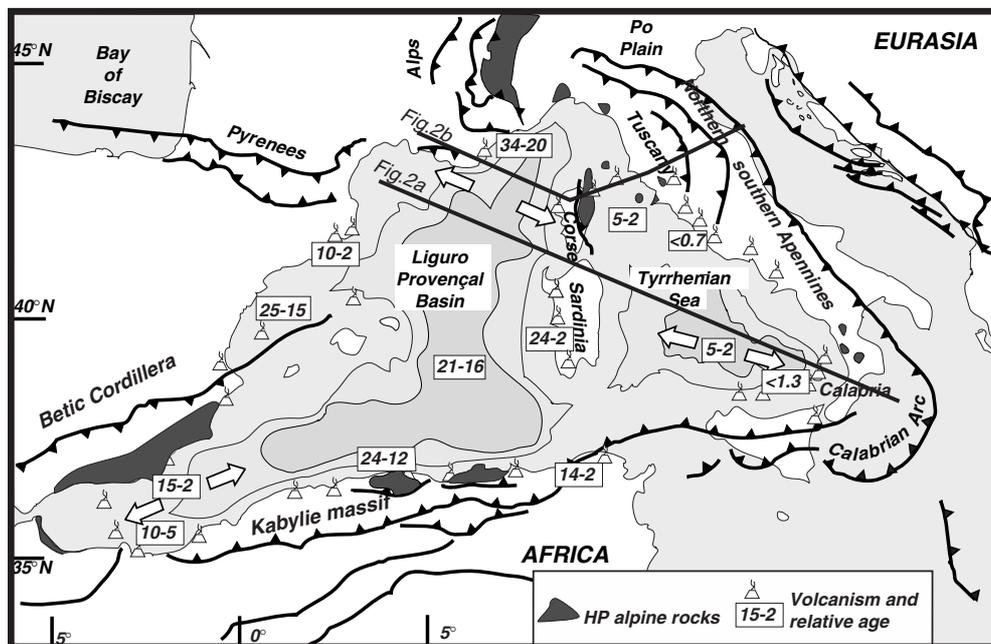


Figure 1. Simplified tectonic map of the central Mediterranean. Shown are the distribution of HP alpine metamorphism, the average ages and the distribution of volcanism (from Lonergan & White 1997). Also indicated are profiles A and B, along which we reconstruct the extensional geological record (*cf.* Fig. 2).

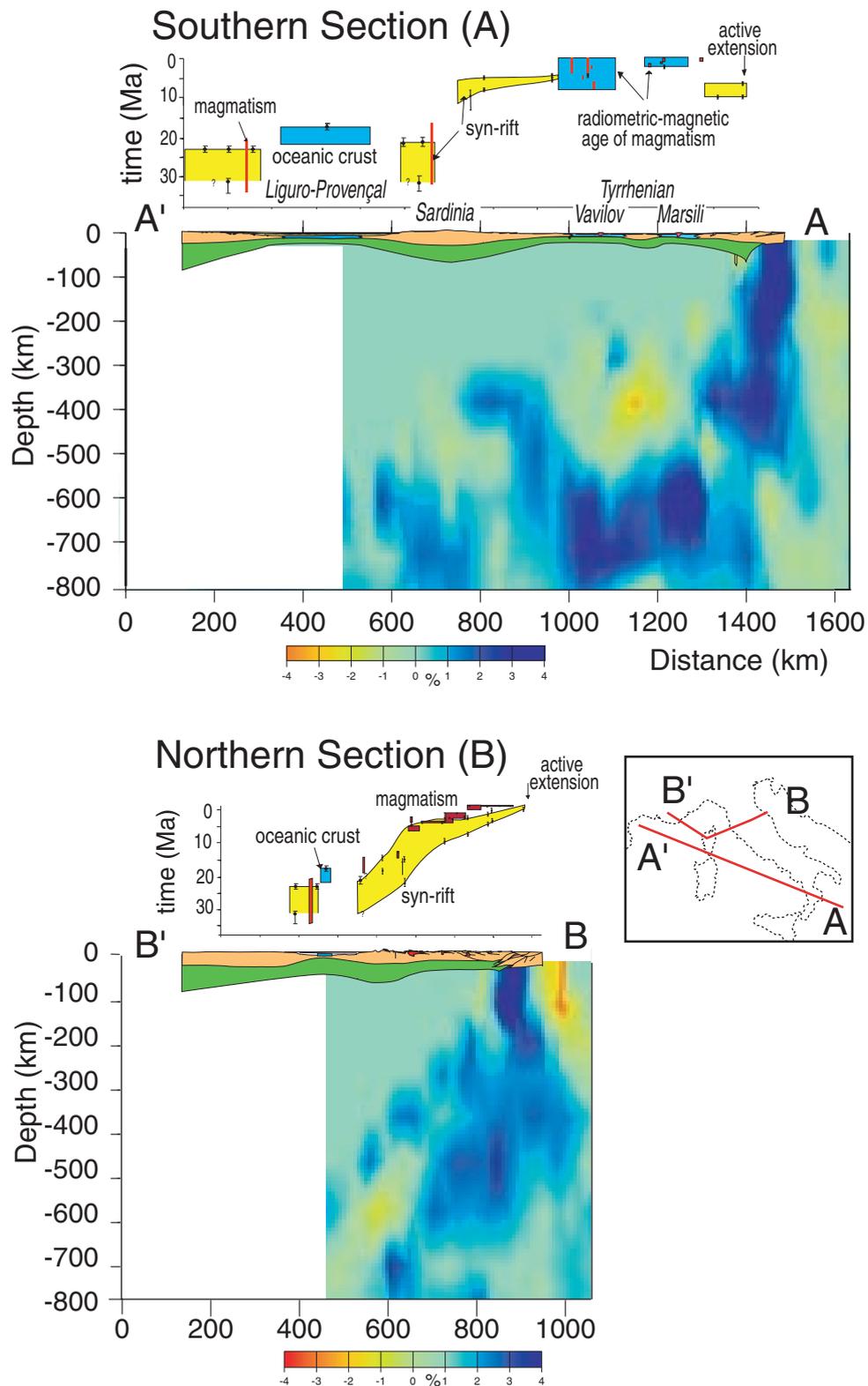


Figure 2. Tectonic cross-sections from (a) Calabria to the Gulf of Lyon and (b) Northern Apennines to the Gulf of Lyon (for location see Fig. 1); the tomographic model is from Lucente *et al.* (1999). Lithospheric cross-sections have been drawn using lithospheric thickness from Suhadolc & Panza (1989), crustal thickness from Nicolich (1989) and Chamot-Rook *et al.* (1999) and Barchi *et al.* (1998). Age of back-arc extension from Faccenna *et al.* (1997).

data set of ~6000 teleseismic *P*- and *PKP*-wave travel times. The southern cross-section from Calabria to the Gulf of Lyon (Fig. 2a) shows an almost continuous high-velocity body extending from the surface below Calabria dipping to the NW

at 70°–80° and then turning horizontally in the transition zone until reaching Sardinia. At a depth of ~500 km the anomaly seems to thin significantly and may be disrupted, whereas it appears to thicken between 600 and 700 km depth.

The estimated total length of the high-velocity zone, interpreted as subducted lithosphere (Lucente *et al.* 1999), is ~ 1200 km measured from the base of the lithosphere, including its lowermost flat portion. This interpretation implies that subducted material has piled up at 660 km without slab penetration into the lower mantle.

Fig. 2(b) shows the Lucente *et al.* (1999) model for our northern cross-section from the Apennines to the Gulf of Lyon. The images show an almost continuous velocity anomaly (up to 3–4 per cent) that dips toward the WSW at $\sim 70^\circ$ – 80° at shallower depths and seems to bend slightly in the transition zone. The slab anomaly corresponds well with earthquake locations, whereas seismicity is limited to shallow depths (~ 90 km) in this region (Selvaggi & Amato 1992).

Using different data sets, Spakman *et al.* (1993), Piromallo & Morelli (1997) and Piromallo *et al.* (2000) have obtained velocity models for the Mediterranean that are similar in both geometry and total length to the Lucente *et al.* (1999) model. Lithospheric structures on the surface along the same cross-sections reveal two thinned regions locally floored by oceanic crust, separated by the 80 km thick Sardinia–Corsica continental block (Suhadolc & Panza 1989).

3 GEOLOGICAL CONSTRAINTS

Geologically, we can discern two different phases for the evolution of the Central Mediterranean subduction zone: (1) formation of the first instability and development of the slab and its progression into the upper mantle, and (2) the episodic opening of back-arc basins. In the following section, the geological data (summarized in Fig. 3) that we considered as remarkable are combined to define the age and style of the subduction process in the Apennines.

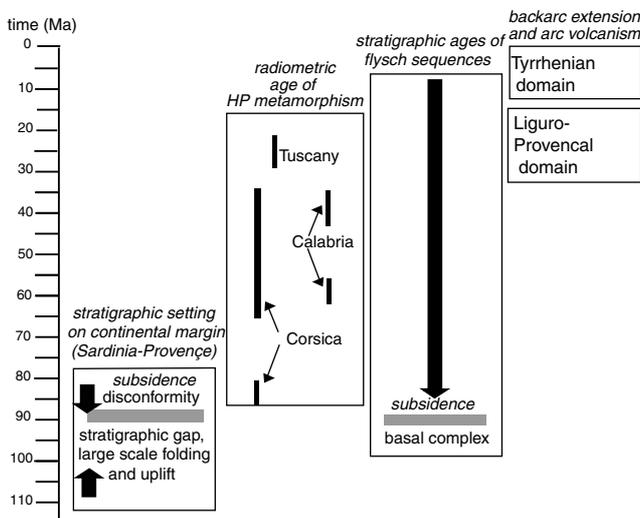


Figure 3. Geological data used to constrain the early stages of subduction in the Central Mediterranean area. Stratigraphic age of Sardinia–Provençal unconformity from D’Argenio & Mindszenty (1991). Radiometric ages of HP metamorphism from Borsi & Dubois (1968), Shenk (1980) and Rossetti *et al.* (2001) for Calabria, Monié *et al.* (1996), Lahondère & Guerrot (1997), Jolivet *et al.* (1998) and Brunet *et al.* (2000) for Northern Apennines and Corsica. Stratigraphic ages of flysch from Principi & Treves (1984), Marroni *et al.* (1992) and Monaco & Tortorici (1995).

3.1 Formation and development of the slab

The whole Mediterranean region was subjected to a large-scale rapid pulse of in-plane compressional stress during the Late Cretaceous (~ 95 – 85 Ma). This event is marked by large scale folding mainly localized along the southern, North African (Bosworth *et al.* 1999) and northern, Iberian–European (Guieu & Roussel 1990; D’Argenio & Mindszenty 1991), Tethyan passive margins of the Ligurian–Piedmont ocean (Dercourt *et al.* 1986) and subordinately along the Apulia microplate (D’Argenio & Mindszenty 1991). Along the northern margin, in particular in Sardinia and in the Provençal area, this event is followed by a rapid subsidence as marked by the presence of a large-scale angular unconformity (~ 90 – 85 Ma, Fig. 3; Guieu & Roussel 1990; D’Argenio & Mindszenty 1991). Soon after this widespread compressional episode, the first evidence of the formation of a trench and the subduction process along the northern, Iberian–European margin occur. This is marked by the onset of siliciclastic deposition, presently crop out in scattered sites along the Apennine chain (~ 85 – 80 Ma, Fig. 3; Principi & Treves 1984; Marroni *et al.* 1992; Monaco & Tortorici 1995). In particular, in the northern and central Apennines, the older foredeep system locally shows continuous sedimentation for approximately 15–20 Myr in the same basins from Campanian (80–85 Ma) to the early Palaeocene (65 Ma) (Marroni *et al.* 1992). Afterwards, during the Neogene, the life of each flysch basin decreased to a few million years and moved rapidly towards the east to the site of present-day deposition in the Adriatic foredeep (e.g. Patacca *et al.* 1990; Cipollari & Consentino 1996). This suggests that during the first stage of slab evolution (Late Cretaceous–Palaeogene) subduction may have progressed slowly (less than 1 cm yr^{-1} ; Marroni *et al.* 1992) and then accelerated.

At the same time crustal material was dragged down into the wedge undergoing HP/LT metamorphism. In the Northern Tyrrhenian region, the HP/LT facies crop out in Alpine Corsica and in the inner Northern Apennines (Figs 1 and 3) and are organized in a double-vergent wedge (Carmignani & Kligfield 1990; Jolivet *et al.* 1998). A maximum pressure metamorphic peak of ~ 16 – 20 kbar is recorded in the Corsica metapelites and pressure progressively decreases to ~ 15 kbar in the Tuscan archipelago and to ~ 10 kbar onshore Tuscany (Jolivet *et al.* 1998). A maximum pressure metamorphic peak of ~ 16 – 20 kbar is recorded in the Corsica metapelites, and pressure progressively decreases to ~ 15 kbar in the Tuscan archipelago, and to ~ 10 kbar onshore Tuscany (Jolivet *et al.* 1998). Ages of blueschist units are from 60 to 35 Myr in Corsica decreasing to ~ 25 Myr in the inner Apennine (Fig. 3; Monié *et al.* 1996; Jolivet *et al.* 1998; Brunet *et al.* 2000). The oldest reliable age for eclogites units in Corsica is 80–85 Myr (Lahondère & Guerrot 1997). In Calabria, metamorphic units show radiometric ages of HP metamorphism ranging from Palaeocene (~ 65 – 58 Ma; Borsi & Dubois 1968; ~ 40 Ma; Shenk 1980) to the Early Oligocene (~ 35 Ma; Monié *et al.* 1996; Rossetti *et al.* 2001). These data indicate that the formation of HP metamorphic facies and their exhumation at the surface can be traced as a continuous process, starting at least during the Palaeocene up to the Miocene (Jolivet *et al.* 1998) and younging towards the east.

In the Apennines, from the Oligocene onwards (30 Ma), the passive continental margin of Apulia entered the trench (Fig. 4), leading to subduction of continental lithosphere (Dercourt *et al.* 1986) as attested by the incorporation of continental

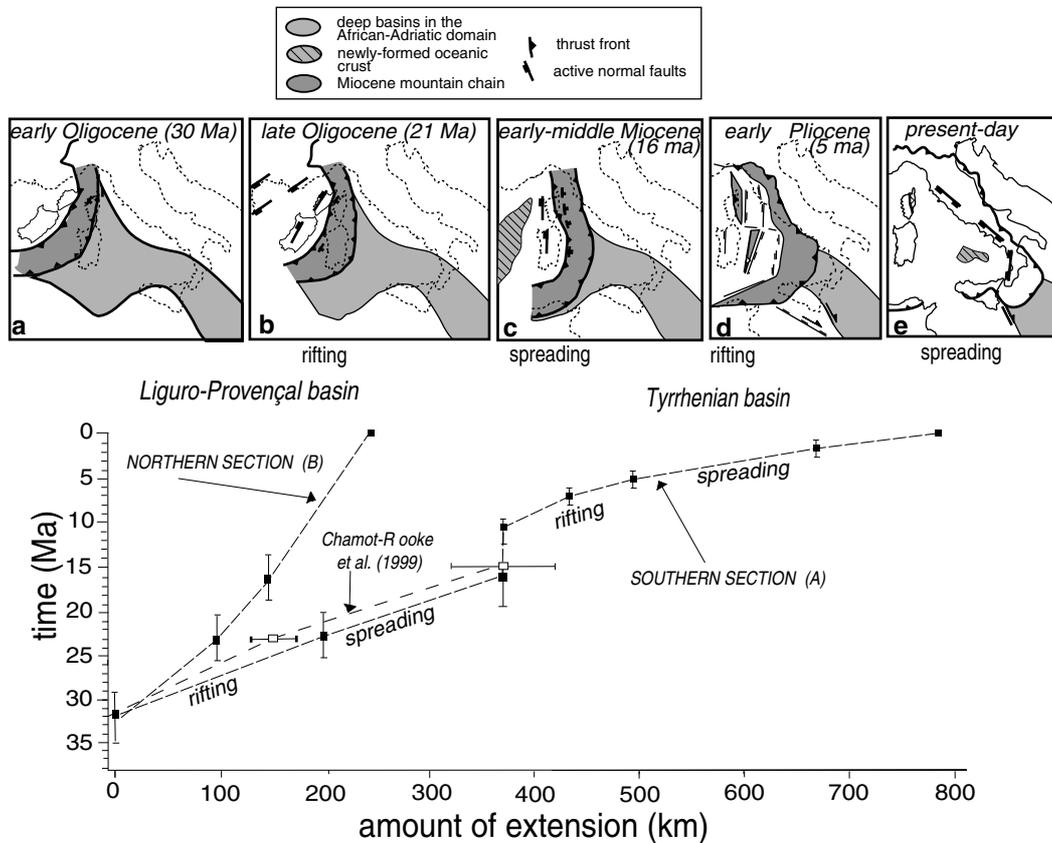


Figure 4. Palaeogeographic reconstruction of back-arc extension in the Central Mediterranean (modified from Faccenna *et al.* 1997) and estimation of the amount of extension in time along the cross-sections of Fig. 2.

passive margin rocks into the Apennine accretionary wedge (e.g. Boccaletti *et al.* 1971; Carmignani & Kligfield 1990). The geometry of the Apulian–Adria plate is unknown. However, palaeogeographic reconstruction (Dercourt *et al.* 1986) suggests that in the southern section the width of the continental plate was reduced to a few hundred kilometres, being confined to the east by the Ionian basin (Fig. 4).

3.2 The opening of the back-arc basins

From ~35–30 Ma, during the formation of HP/LT metamorphic facies rocks and the deposition of siliciclastic deposits, arc volcanism and extension started behind the accretionary wedge (Beccaluva *et al.* 1989). This indicates that the slab must have reached a minimum depth of 100–150 km to produce melting and arc volcanism at the surface (Jacob *et al.* 1976).

In order to estimate the amount and the rate of back-arc extension we plot the age of the syn-rift deposits, the age of oceanic crust and volcanism along a cross-section running perpendicular to the strike of the basins (Fig. 2). Arc volcanism appeared first in Sardinia (~34–32 Ma) and in the Provence, followed by the formation of the first extensional basins (~30–23 Ma Cherchi & Montandert 1982; Gorini *et al.* 1994; Beccaluva *et al.* 1989). From the late Aquitanian, post-rift deposits unconformably covered the syn-rift sequence (Gorini *et al.* 1994; Seranne 1999) during oceanic spreading (Burrus 1984) and the ~25°–30° counterclockwise rotation of the Corsica–Sardinia block, which most probably occurred between 21 and 16 Ma (Van der Voo 1993). After a few million years, extension

shifted to the Southern Tyrrhenian basin. Syn-rift deposition started at ~10–12 Ma in both Sardinia and Calabria (Kasten & Mascle 1990; Sartori 1990), followed by the formation of localized spreading centres (4–5 Ma, Vavilov basin, 2 Ma, Marsili basin) with drifting and arching of the Calabria block (Sartori 1990).

In the Northern Tyrrhenian sea, extension is marked by a system of spaced crustal shear zones, sedimentary basins and magmatic activity which gets younger towards the east from Corsica to the Apennine chain (Fig. 2b; Serri *et al.* 1993; Bartole 1995; Jolivet *et al.* 1998). There are syn-rift deposits from at least 20 Ma in the Corsica basin (Mauffret & Contrucci 1999) and 2 Ma along the Apennine watershed (for reviews, see Bartole 1995 and Faccenna *et al.* 1997), where normal faults are still active. Magmatic activity accompanied and post-dated the end of the syn-rift growth of the basin. Both the extensional basin and the magmatic centre shifted eastwards, away from the rift axis, at an average velocity of 1.5–2 cm yr⁻¹ (Fig. 2b).

This data set allows us to reconstruct the stretching and spreading events in time. The amount of back-arc extension is estimated by reconstructing the main tectonic events at the scale of the crust in different steps. First, the oceanic-floored area of each basin is removed from the sections of Fig. 2, and then, using an area balancing technique, the crust is restored to the thickness of the shoulders (~30 km, see Finetti & Del Ben 1986; Chamot-Rooke *et al.* 1999), assuming that the locus of extension at the surface corresponds to the locus of maximum crustal thinning. The Northern Tyrrhenian section is restored to an average minimum thickness of 35 km, taking into account the previous thickening history (Jolivet *et al.* 1998). Fig. 4 shows

the amount of extension attained by the Southern Tyrrhenian (3, 5, 7 and 10 Ma), northern Tyrrhenian (16 and 23 Ma) and the Liguro-Provençal basins (16, 23 and 31 Ma).

Commenting on the southern section, it is noted that

(i) the total amount of extension (~ 780 km) is partitioned roughly equally between the two basins, with alternating episodes of rifting (~ 7 Myr) and oceanic spreading (~ 5 Myr);

(ii) the rate of extension is reduced at the end of spreading of the Liguro-Provençal basin and rifting in the Tyrrhenian initiated after a small pause of few million years and progressively accelerated.

With reference to the northern section, it is noted that

(i) The total amount of extension (~ 240 km) is divided between the Liguro-Provençal basin (~ 140 km) and the Northern Tyrrhenian basin (~ 100 km) and the oceanic spreading was confined to a narrow strip (~ 30 km) in the Ligurian basin (Mauffret *et al.* 1999);

(ii) After the initial stretching event in the Liguro-Provençal basin, the rate of extension progressively decreased.

We also find from both sections that the volcanic arc-trench gap is wide (~ 400 km) in comparison to the present day. This suggests, in agreement with the palaeoreconstruction of Beccaluva *et al.* (1989) and Seranne (1999), that the slab attained a shallow dip prior to and after the opening of the Liguro-Provençal basin. In addition, slab steepening has been proposed by Jolivet *et al.* (1999) and Brunet *et al.* (2000) along the northern section of the Tyrrhenian sea to account for the decreasing time gap between the HP compressional event and the onset of extension and magmatism.

The amount and velocity of extension estimated here is in good agreement with previous evaluations. Along the southern section of the Liguro-Provençal basin, the amount of rifting was estimated to be 150 ± 20 (Chamot-Rooke *et al.* 1999) while the amount of spreading was estimated to be 150 km (Mauffret *et al.* 1995), 230 (Burrus 1984) and 230 ± 20 km (Chamot-Rooke *et al.* 1999; Fig. 4). For the Tyrrhenian sea, the total extension evaluated by Malinverno & Ryan (1986) is of the order of 330–350 km. In the southern section of the Tyrrhenian sea similar values have been estimated by Patacca *et al.* (1990) and Spadini *et al.* (1995). Furthermore, the total amount of extension along the whole southern section agrees well with previous studies (Gueguen *et al.* 1998).

4 AFRICAN MOTION AND TRENCH ORIENTATION

Several kinematic reconstructions were proposed for the Mediterranean region to define the way the African plate converges with Eurasia (e.g. Dewey *et al.* 1989; Dercourt *et al.* 1993). All of these models basically agree that Africa moved slowly relative to Eurasia with counterclockwise rotation, moving NE up to ~ 40 Ma, then N–S and finally NNW (Fig. 5). The relative velocity of a point located half way along the northern African coast was probably less than 3 cm yr^{-1} during the last 80 Myr, halving during the last 20–30 Myr (Jolivet & Faccenna 2000). This is in agreement with another recent estimate of absolute plate motion for Africa (Silver *et al.* 1998). Geodetic data indicate that Africa currently moves N20°W at a rate of 0.7 cm yr^{-1} in the Central Mediterranean (Ward 1994).

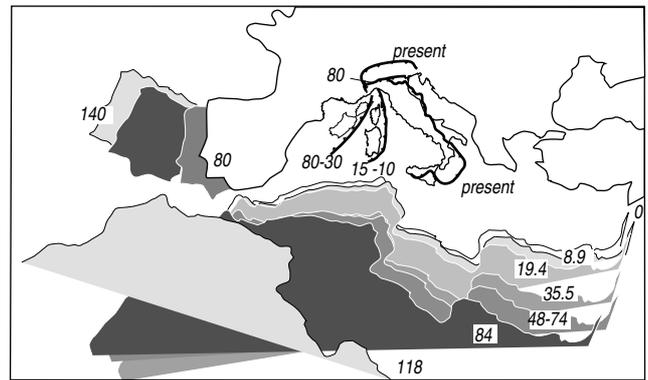


Figure 5. African motion with respect to fixed Eurasia as calculated by Dewey *et al.* (1989) and motion of Iberia (redrawn from Olivet 1996). The orientation of the Central Mediterranean trench is also shown, numbers are Myr.

In order to estimate the net convergence rate (the rate at which the plates moved perpendicularly to the trench), the orientation of the trench in time was reconstructed (Fig. 5). Palaeomagnetic data indicate that Iberia underwent a significant rotation ($22^\circ \pm 14^\circ$) with respect to Europe between 132 and 124 Ma (Van der Voo 1993; Moreau *et al.* 1997). After this episode, the rotation of the Iberian peninsula slowed down and was followed by translation (~ 120 Ma) during the initial phase of the opening of the Bay of Biscay (~ 85 Ma; Olivet 1996 and references therein). At that time, the trench was oriented NE–SW running parallel to the former Iberian passive margin (Fig. 5). Subsequently, the trench's position remained rather stable, turning to N–S only after the rotation of the Sardinia–Corsica block (~ 21 –16 Ma; Van der Voo 1993), and then turning again to its present-day position during the Southern Tyrrhenian spreading episodes (~ 5 –2 Ma). The trench was therefore oriented roughly parallel to the motion of Africa during most of the subduction process. With data from Dewey *et al.* (1989), it is possible to estimate that the total amount of net convergence produced by the motion of Africa in the past 80 Myr is ~ 240 km with an average rate of 3 mm yr^{-1} (Fig. 7a). The motion of the Adria microplate cannot contribute significantly to increase the convergence because its Tertiary motion can be assumed to be coherent with that of Africa (Chamot-Rooke 1986).

We therefore observe that the net convergence velocity on the Central Mediterranean trench appears to be very low when compared with other subduction zones worldwide (Jarrard 1986). As a consequence, the subduction process will be mostly driven by the negative buoyancy of the subducted slab, an effect that has been studied in the models of Faccenna *et al.* (1996) and Giunchi *et al.* (1996).

5 RECONSTRUCTING THE SUBDUCTION PROCESS

The reconstruction of the evolution process along the two cross-sections of Fig. 2 (Fig. 6) is described below. The present-day configuration of the slab was obtained from the 2–4 per cent upper mantle velocity anomaly as shown in Fig. 2, assuming a continuous slab. As convergence is considered negligible, the amount of back-arc extension shown in Fig. 4 is directly related to the amount of subduction. From the present-day

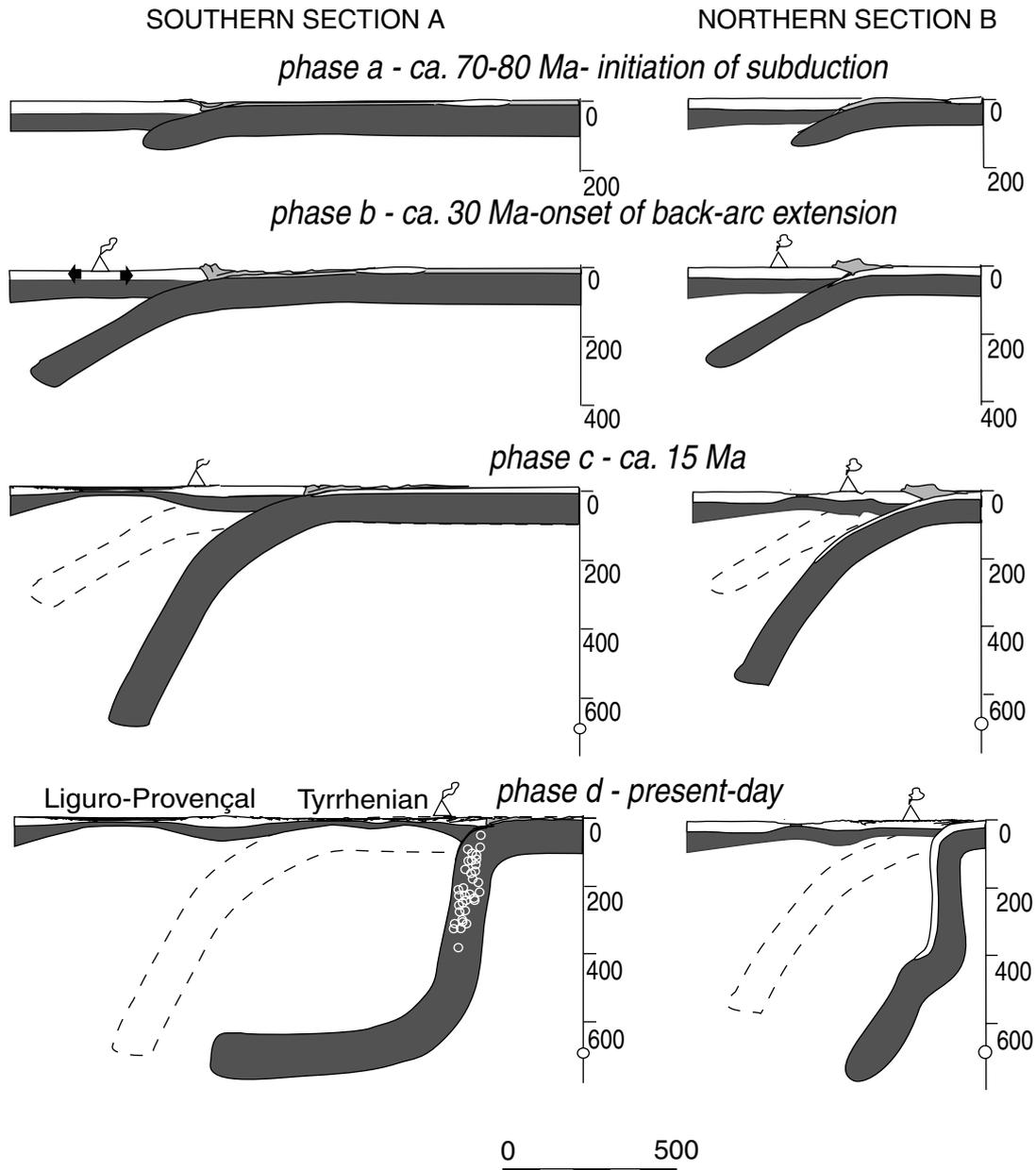


Figure 6. Four-phase reconstruction of the subduction process along the cross-sections of Fig. 2. The dip deep of the slab at phase C is inferred.

configuration the total amount of subducted material during the opening of the back-arc basins is then subtracted. Phases b and c in Fig. 6 show the slab configuration at the beginning of the Tyrrhenian and Liguro-Provençal rifting, respectively. Note that the deep dip of the slab during phase c is only inferred.

We divide the subduction history in the Central Mediterranean into three stages (Fig. 6).

(1) Phases a–b, from 80 to 30–35 Ma. In both sections, the amount of subduction was very small, ~ 400 km corresponding to an average velocity of 0.8 cm yr^{-1} . Arc volcanism developed when the slab reached a depth of 200–300 km, attaining a shallow dip.

(2) Phases b–c, from 30–35 to 15 Ma. In both sections, the velocity of subduction and the slab dip increased rapidly during the opening of the Liguro-Provençal basin. At the end of the

Sardinian drifting phase (Liguro-Provençal opening) the slab reached a depth of ~ 600 km. In the southern section, the subduction process stopped at ~ 15 Ma. In the northern section, subduction velocity was always slower than in the South and is observed to decrease progressively with time.

(3) Phases c–d, from 15 Ma to the present day. In the southern section, the slab is deformed, folded at the 660 km-transition zone and again increases its dip. This is indicated by a pause in subduction for about 5 Myr and the subsequent acceleration of the rollback velocity up to 6 cm yr^{-1} during the opening of the Tyrrhenian back-arc basin. In the northern area, conversely, subduction slowed permanently.

In conclusion, we infer that subduction started very slowly. After the initiation phase, the velocity of subduction increased sharply with time during the descent of the slab into the upper mantle while the dip of the slab increased and arc volcanism

developed 40–50 Myr after the onset of subduction. Soon after the opening of the Liguro-Provençal basin the slab reached the transition zone. Along the southern section, where only oceanic material was being subducted, the subduction velocity decreased, dropping temporarily to zero. After a few million years, however, the slab was probably re-bent at depth, it steepened again and the trench rolled back and re-accelerated during the rifting and spreading of the Tyrrhenian basin. Conversely, along the northern section, a progressive decrease of the subduction velocity and no re-acceleration is observed.

This scenario, of a long subduction process leading to the opening of two back-arc basins, is in agreement with previous geological models formulated on the basis of different data sets (e.g. Principi & Treves 1984; Knott 1987; Bortolotti *et al.* 1990; Jolivet *et al.* 1998; Rossetti *et al.* 2001). The geodynamic aspects are elaborated on later. In the scenario presented here the width of the oceanic lithosphere is limited to 400–500 km in the northern area and 700–800 km in the southern area. These values are consistent with previous palaeogeographic reconstructions (Dercourt *et al.* 1986; Schmid *et al.* 1996).

5.1 Comparison with previous models

An alternative palaeotectonic scenario suggests that subduction in the Central Mediterranean initiated at a later time than proposed here and only after a polarity flip with an earlier, SE-dipping subduction zone took place. The timing of this event has been put at the Cretaceous–Tertiary boundary (65 Ma; Dercourt *et al.* 1986), the Palaeocene (50 Ma, Boccaletti *et al.* 1971) or the Oligocene (30 Ma, Doglioni *et al.* 1997). The main objective of these alternative models is to explain the development of west-verging nappe structures in Alpine Corsica and the east-verging Apenninic structures. However, all of these models are at odds with the observation that HP metamorphic units are also present inside the internal Apenninic nappe and that these units are progressively younger moving towards the east from 65 Ma up to 25–22 Ma. This indicates a continuous evolution from the development of subduction to back-arc extension (Jolivet *et al.* 1998) (Fig. 3). Furthermore, the east-verging siliciclastic deposition that accompanies the development of the Apenninic trench is a continuous process starting from the Late Cretaceous up to the present-day (Fig. 3; Principi & Treves 1984). These data support the model of a double-vergent Alpine–Apenninic orogenic wedge, related to the same northwestward subduction from the very first stage of deformation, with an along-strike change in the dip of the subduction zone moving towards the Alps.

6 DISCUSSION

The reconstruction of Fig. 6 can be presented in the form of a diagram showing the amount of subduction versus time for the southern (Fig. 7a) and the northern (Fig. 7b) cross-sections. Starting from the present, the current length of the slab as inferred from the tomographic model is plotted. The error bars shown are estimates of uncertainties that are considered conservative upper bounds. Other data in Fig. 7 come from calculating the amount of extension. Under the assumption that convergence is negligible (see above), extension is identical

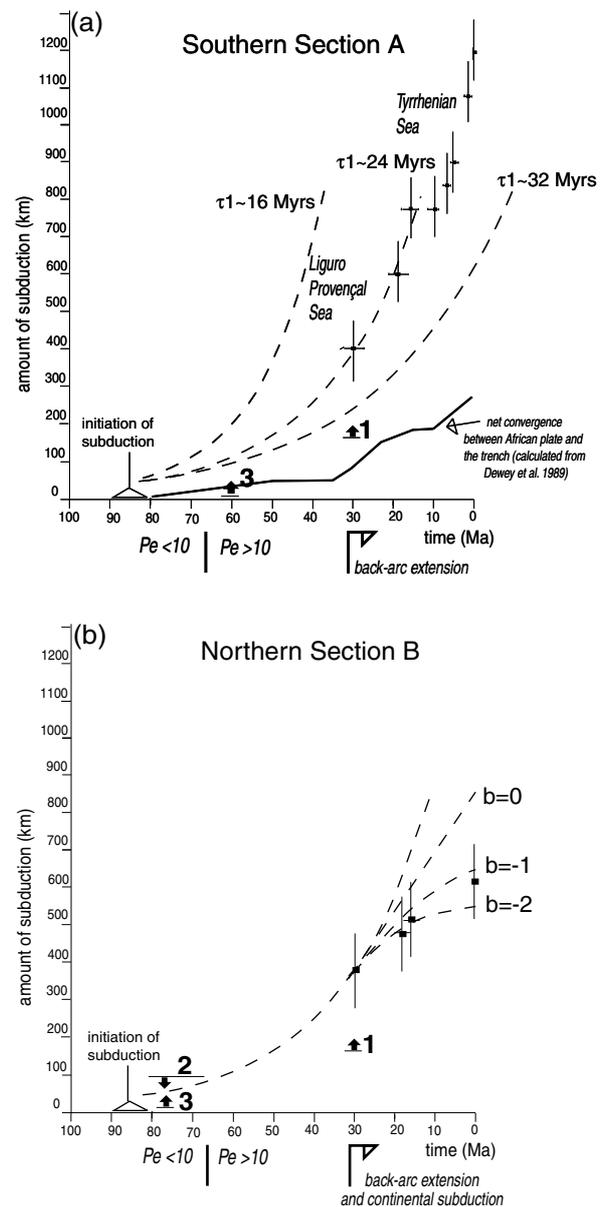


Figure 7. Amount of subduction as a function of time for the Central Mediterranean subduction zone along the cross-sections of Fig. 6: (a) southern cross-section and (b) northern cross-section. Curves are drawn from eqs. 1 (left plot, exponential increase part) and (3) (right plot, slowdown of subduction after inclusion of continental material) with the parameters mentioned in the text. Arrows denote the constraints from: (1) onset of volcanism (2) duration of the continuous sedimentation of foredeep deposition (see text for details), and (3) first appearance of HP metamorphic assemblages.

to the amount of subduction. Further constraints are the onset of arc volcanism, which gives a minimum amount of subduction of 150–200 km at that time (using 45° as the maximum slab dip, as constrained by the arc-trench gap) and in the northern section, the maximum amount of subduction during the deposition of the first foredeep deposits.

In the next section the first, accelerating part of the slab depth versus time curve that can be observed along both cross-sections is discussed.

6.1 The first stages: initiation and development

Before the slab reaches the 660 km discontinuity, the data in Fig. 7 can be interpreted using an argument of Becker *et al.* (1999). These authors show that many first-order observations in subduction zones can be explained by simple heterogeneous flow models. In particular, Becker *et al.* (1999) observed that the subduction velocity increases roughly exponentially with time for slab instabilities which were free to sink under their negative buoyancy (i.e. no ridge push) before impedance to flow at the transition zone became important. In both numerical and laboratory convection models the data could be fitted by equating the viscous dissipation per unit length due to the bending of oceanic lithosphere to the rate of potential energy change by slab pull. This approach was motivated by the work of Conrad & Hager (1999).

The depth extent of the slab at time t , $H(t)$, then scales as

$$H(t) = H_0 \exp\left(\frac{t}{\tau_1}\right), \quad (1)$$

where H_0 is the initial length and τ_1 is a characteristic timescale that follows from the scaling argument and is given by

$$\tau_1 = \frac{\mu R^2}{C \Delta \rho_{oc} g r^3}. \quad (2)$$

Here, $\Delta \rho_{oc}$ is the density contrast between oceanic plate and mantle, μ is the effective viscosity of the bending plate, g is the gravitational acceleration, r is the bending radius, R is the half-thickness of the plate, and C is a fitting constant, which was ~ 0.28 for the numerical experiments (Becker *et al.* 1999). For eq. (1) it is assumed that only the viscous bending of the oceanic plate and the gain in buoyancy forces matter, that R and r are constant during subduction, and that the entrainment of upper plate lithosphere as well as the varying dip of the slab are not important. Accordingly, eq. (1) is only a first-order approximation, an end member case against which to test observations.

In Fig. 7(a), we compare eq. (1) with our data for a range of different timescales τ_1 . The trend observed for the Central Mediterranean subduction zone can be fitted if τ_1 is equal to 24 Myr (Fig. 7a). Using a value of 125 km for r (as measured from the present-day configuration), 45 km for R (Suhadolc & Panza 1989), and 10^{23} Pa s for the viscosity of the lithosphere (see Ranalli 1995), we can solve for $\Delta \rho_{oc}$ in eq. (2), finding a value of ~ 50 kg m $^{-3}$.

Considering our data uncertainties, it cannot be argued that there is any particular non-linear functional dependence of H on t in a statistical sense. Furthermore, eq. (2) shows that τ_1 strongly depends on the geometrical factor r^2/R^3 , which is poorly constrained by geological data and might change substantially with time. This diminishes the resolution of the scaling argument with respect to parameters such as $\Delta \rho_{oc}$ and μ . Other possible important effects that were not considered include viscous dissipation in the mantle and the effects of non-Newtonian rheology, which might be poorly described by our effective viscosity approach.

However, we consider robust our estimate of a timescale $\tau_1 = 24$ Myr. Whatever process might be responsible for the subduction dynamics has to yield a similar number. The viscous bending argument gives the correct timescale if one accepts that a density contrast of around 50 kg m $^{-3}$ and a plate thickness

of 90 km are realistic estimates (e.g. Cloos 1993) for the Central Mediterranean subduction zone. This seems plausible since the plate was at least 70 Myr old at initiation (see e.g. Abbate *et al.* 1984, for the age of the oceanic plate). An effective viscosity of 10^{23} Pa s for the lithosphere is consistent with estimates of a maximum factor of 100–500 contrast between lithospheric and mantle viscosities (Conrad & Hager 1999; Becker *et al.* 1999) when we use a canonical postglacial rebound value of 10^{21} Pa s for the mantle.

Along the northern section we observe a decrease in the subduction speed from 30 Myr onward. This can be due to continental material entering the trench (Faccenna *et al.* 1997). Stretching the Becker *et al.* (1999) argument further, we therefore modify the balance between viscous dissipation and slab pull energy release that has led to eq. (1) to allow for the addition of positive buoyancy to the slab column. If we stick to all simplifications from above and assume that the change in the density contrast from oceanic ($\Delta \rho_{oc} > 0$) to continental ($\Delta \rho_{con} < 0$) material occurred at time t_1 , we can deduce that the slab depth relative to $H(t=t_1)$ will proceed as

$$h'(t') = \frac{1}{b} \left(\exp\left(\frac{t'}{\tau_2}\right) - 1 \right), \quad (3)$$

where $h'(t') = H(t')/H(t=t_1) - 1$, $t' = t - t_1$, $b = \Delta \rho_{con}/\Delta \rho_{oc}$ and τ_2 is a second timescale, related to τ_1 as follows:

$$\tau_2 = \frac{\Delta \rho_{oc}}{\Delta \rho_{con}} \tau_1 = \frac{\tau_1}{b}. \quad (4)$$

Imposing the condition that continental material entered the trench ~ 30 Ma and assuming that all other parameters in the system remain constant, it was found that the decrease of trench retreat (and hence the decrease in the amount of subducted material) can be modelled by continental material with $b \sim -1.5$ (Fig. 7b). This corresponds to $\Delta \rho_{con} \sim -75$ kg m $^{-3}$. Such a value is comparable to those proposed for a thinned, continental lithosphere such as the Adriatic (Cloos 1993).

The finding that our observations are well described by the simple scaling argument above lends credibility to the idea that the main resisting force for subduction arises from the viscous drag within the bending oceanic plate at the trench (Conrad & Hager 1999; Faccenna *et al.* 1999; Becker *et al.* 1999). The implication of the results presented here for subduction zones in nature is that slow unstable growth in non-converging margins should be expected, followed by an exponential increase of the subduction velocity as a consequence of the increasing slab pull before interaction with the phase change at 660 km becomes important. A possible incursion of continental material at the trench, conversely, will lead to a slowdown of the subduction process on timescales given by τ_2 .

In settings with slow convergence, we can expect that thermal diffusion will spread out gravitational instabilities and therefore counteract subduction initiation. This effect will become important if convective timescales are greater than diffusive ones. As a measure of this effect, the Peclet number can be used

$$Pe = \frac{u_c l_c}{\kappa}, \quad (5)$$

where u_c and l_c are characteristic velocity and length scales, respectively, and κ is thermal diffusivity. Thus, high Pe means

that convection is effective and small Pe indicates that diffusion will tend to smear out buoyancy anomalies.

Arbitrarily picking $Pe = 10$ as a benchmark and taking $l_c = 100$ km and $\kappa = 10^{-6}$ m² s⁻¹ as characteristics numbers, it was found that velocities should be larger than ~ 0.3 cm yr⁻¹ for convection to dominate. [More sophisticated treatments of similar problems can be found in the literature, e.g. Conrad & Molnar (1997), but an order of magnitude estimate will suffice here.] Therefore, diffusion can be expected to have been important during the very earliest, slow stages of subduction (~ 20 Myr). However, other effects such as strain rate and/or grain-size-dependent viscosity (Riedel & Karato 1997) or the presence of a pre-existing weak zone might enhance the rate of instability growth and accelerate initiation. Eventually the slab pull must have been strong enough to overcome limiting factors since it is known that subduction initiated in the Mediterranean.

A comparison with other subduction zones indicates that the model presented here applies strictly to a slowly converging setting because the onset of arc volcanism post dates the onset of subduction by only a few millions of years in fast-converging areas such as New Zealand (Stern & Holt 1994).

6.2 Episodic opening of back-arc basins and interaction with the 660 km transition zone

In the southern cross-section, the initiation and development stage is followed by a decrease in subduction velocity when the slab reaches the 660 km transition zone (Fig. 7a). This coincides approximately with the end of the first back-arc spreading. Afterwards, the velocity of subduction increases again with time during the opening of the second back-arc basin. This sequence is interpreted as indicating that the opening of the first back-arc basin occurred as a consequence of the increased velocity of subduction. This was driven by the negative buoyancy of the oceanic plate; once the instability reached a significant depth (200–300 km), its total slab pull increased causing extension, rifting and break-up of the overriding plate (Faccenna *et al.* 1996; Becker *et al.* 1999). In addition, the decrease in the African absolute motion ~ 35 –30 Myr ago (Silver *et al.* 1998) may have further contributed to decrease the convergence velocity allowing a net increase in the velocity of trench retreat (Jolivet & Faccenna 1999).

In our reconstruction, the end of the first spreading episode and the initiation of the second spreading phase is due to the interaction between the slab and the 660 km transition zone (Faccenna *et al.* 2000). This is likely to produce a slowing down of the subduction velocity (e.g. Zhong & Gurnis 1995; Fucicello *et al.* 1999) and a decrease in the dip of the slab which can in turn cause a decrease of the in-plane extensional stress on the upper plate (Tao 1991). After few million years, the slab steepens again and re-accelerates, leading to the opening of the second back-arc basin.

The seismic tomography models have been interpreted such that a continuous flow of subducted material has piled up temporarily at 660 km. This is congruent with the expectation that slab rollback will delay slab penetration into the lower mantle (e.g. Griffith *et al.* 1995; Christensen 1996). In a non-converging margin setting, it can be implied that restricted upper mantle convection seems to have been the dominant mode during the whole evolution of the slab.

7 CONCLUSIONS

The evolution of the subduction of the Central Mediterranean represents a unique opportunity to unravel the way the upper mantle evolved over an 80 Myr time-span. We find that subduction was dominated by slab-pull in a restricted, upper mantle convection environment. Jointly interpreting geological constraints and a seismic tomography model allows the reconstruction of a non-steady state, three-stage process: (1) formation and nucleation of the first instability; (2) development of subduction in the upper mantle and (3) episodic opening of back-arc basins during the interaction between the slab and the transition zone at 660 km. The first stage is characterized by the initiation of subduction with a slow growth of the instability, resisted by thermal diffusion. The second stage is characterized by the development of subduction, where the slab accelerated during its sinking into the upper mantle. This episode can be successfully modelled by equating the viscous dissipation per unit length due to the bending of oceanic lithosphere to the rate of potential energy release by slab pull. Finally, the third stage is controlled by the interaction between the slab and the 660 km transition zone resulting, in the southern region, in an important re-shaping of the slab and intermittent pulses of back-arc extension. In the northern region the decrease in the trench retreat can be explained by the convergence of positively buoyant continental material at the trench.

ACKNOWLEDGMENTS

We thank Renato Fucicello for continuous encouragement during all stages of this work. Particular thanks go to Domenico Giardini for his support in clarifying and stimulating the main concepts expressed here. CF and TWB are indebted to Richard J. O'Connell and for discussion and continuous support and to invite CF to Harvard where part of this work was carried out. This paper also benefited from discussions with Giorgio Ranalli, Fabio Speranza, and Francesca Fucicello. TWB was supported by DAAD under a Doktorandenstipendium HSP-III. This work was supported by CNR (programme SubMed, R. Fucicello).

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