

The kinematics of back-arc basins, examples from the Tyrrhenian, Aegean and Japan Seas

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Abstract: Three classical examples of marginal basins are explored to show the respective contributions of body forces and far-field stresses to the extension mechanism. Extensional stresses can be provided by: (1) the slab-pull force, which induces a retreat of the slab and which originates in the density contrast between the subducting slab (oceanic or continental) and the asthenosphere; (2) by lateral density contrasts within the crust (due to crustal thickening), which induce crustal spreading; and (3) by far-field stresses due to intra-continental deformation (continent–continent collision, for example). Slab pull is probably the most efficient extensional force to provide but its effects are modulated by the contributions of the two other forces. The Japan Sea opened along the eastern margin of the Eurasian continent because extensional boundary conditions were provided by the retreating Pacific subduction zone. The opening stopped as soon as the central Japan triple junction was established in its present position which resulted in a more efficient coupling between the Pacific and Eurasian Plates through the Philippine Sea Plate. The geometry of opening was further controlled by large-scale dextral strike-slip faults that run oblique to the subduction zone along >2500 km, and which are far-field effects of the India–Asia collision. The Northern Tyrrhenian Sea opened because of the retreat of the Adriatic continental slab. The strong slab-pull force is probably due to phase changes within the subducting lower crust. Crustal delamination leads to a warm lower crust which localizes the extensional strain. This extending domain migrated with time from west to east as the delamination and slab retreat proceeded. Upper crustal units incorporated in the Apennines accretionary wedge were later exhumed in the collapsing back-arc domain where their deformation and *P–T* history can now be observed. A similar history with an outward migration can be proposed for the Aegean Sea with, however, a stronger influence of crustal collapse over a larger domain. Here too continental collision (the Arabia–Eurasia collision) controlled the geometry of opening through the westward propagation of the North Anatolian Fault.

When back-arc basins form within the active margin of an intra-oceanic subduction they often present a clear pattern of magnetic anomalies. This is the case for some of the western Pacific back-arc basins which developed along the eastern border of the Philippine Sea Plate or the Australian Plate, such as the Lau Basin (Parson & Hawkins 1994) or the Shikoku–Parece Vela Basin (Mrozowski & Hayes 1979; Chamot-Rooke *et al.* 1987) (Fig. 1). In these cases, the kinematics of opening can be derived from the

geometry of magnetic anomalies. The average velocity of opening is *c.* 2–3 cm yr⁻¹, although it can be much faster in some extreme cases, e.g. the northern Lau Basin (Bevis *et al.* 1995). Back-arc basins located on continental margins have much more complicated geometries and the pattern of magnetic anomalies is usually not helpful. With the notable exception of the South China Sea (Taylor & Hayes 1983; Briais *et al.* 1993), magnetic anomalies are usually very poor, e.g. in the Japan (Tamaki *et al.* 1992) or the Tyrrhenian

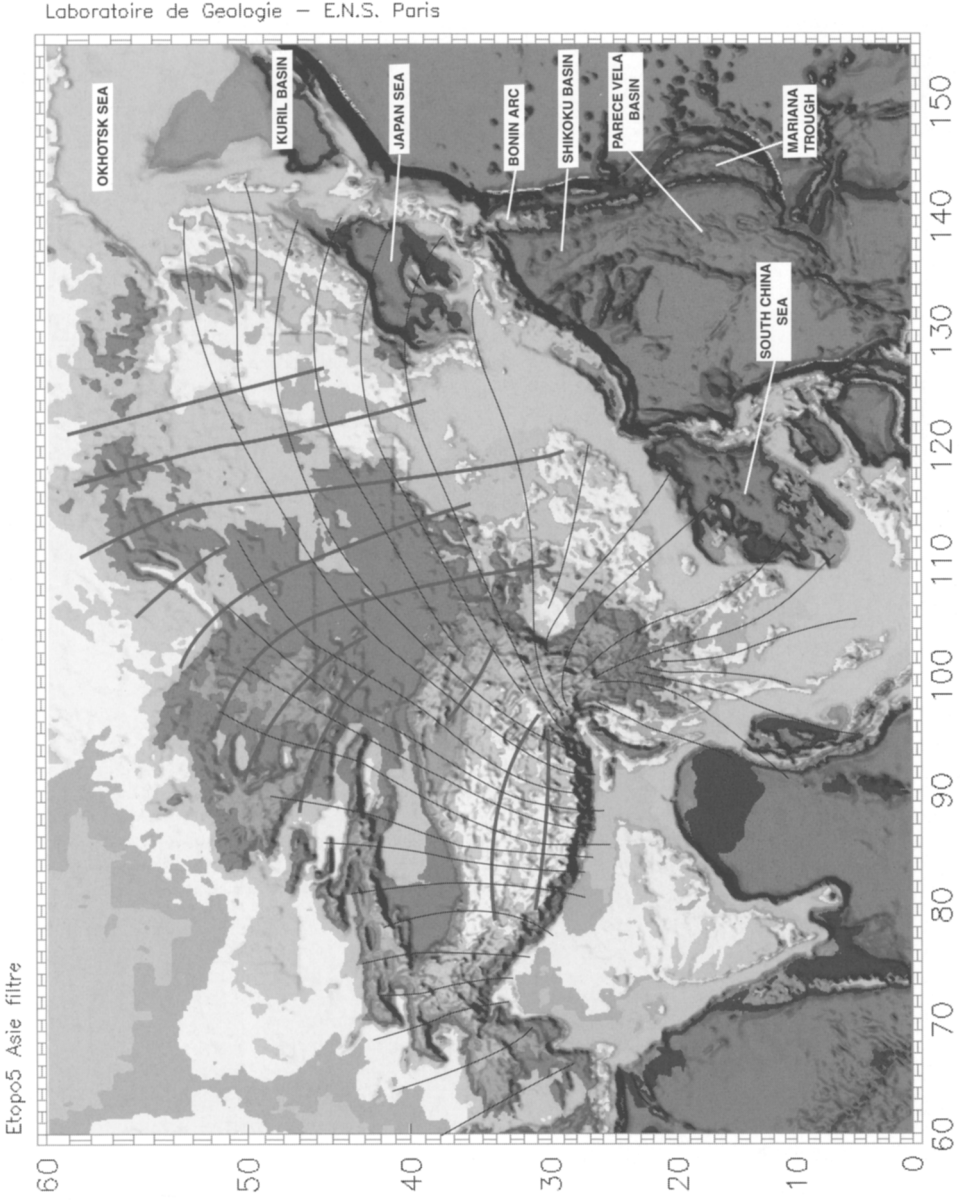


Fig. 1. Topographic map of Asia and the western Pacific region (etopo5) and the present-day stress field deduced from earthquake focal mechanisms. Thin lines, σ_{Tmax} ; thick lines, σ_{Tmin} .

Seas where they have not been identified (Sartori 1990). In some cases, oceanic crust is either too recent to produce significant anomalies (the Okinawa Basin) or non-existent (the Aegean Sea).

To reconstruct the kinematics of such basins it is necessary to work out the regional geological history, particularly the history of deformation of the margins during the opening. Although the geometrical constraints are likely to be more uncertain than in the case of a clear magnetic anomaly pattern, the tectonic history of the margin usually allows the proposal of a reasonable tectonic evolution and a discussion of the tectonic evolution in terms of the forces involved.

The formation of back-arc basins represents one of the major problems in plate tectonics and several mechanisms have been proposed to explain their evolution (Taylor & Karner 1983). The geometry and the rate of extension is thought to be controlled by the roll-back of the hinge of the subducting slab (Dewey 1980). From a dynamical point of view, this process is possibly connected with the density contrast of the oceanic lithosphere sinking down into the asthenosphere (Forsyth & Uyeda 1975; Uyeda & Kanamori 1979). In this case, the length of the slab should be directly related to the velocity of retreat in the mantle (Zhong & Gurnis 1995; Faccenna *et al.* 1996). The presence of lateral (Dvorkin *et al.* 1993; Russo & Silver 1994) or global (Ricard *et al.* 1991) mantle flow can also accelerate this process. The way the overriding plate reacts to the retreat of the slab hinge is still enigmatic [for a review see Taylor & Karner (1983)]. Nevertheless, it has been suggested that extensional stress could be more easily transmitted from the sinking to the upper plate when the coupling between the two plates is low (Bott *et al.* 1989; Shemenda 1994; Scholz & Campos 1995; Hassani *et al.* 1997).

Besides the slab roll-back force, which is always present, continental margins are subjected to a stress regime which is controlled by local body forces and far-field boundary forces. Described here are three examples of back-arc basins where other external forces contribute with the slab-pull force producing large-scale lithospheric extension in the upper plate. Chosen for examination are the Japan, the Aegean and the Tyrrhenian Seas, examples of back-arc basins where gravitational collapse and continental collision have contributed with the slab-pull forces to the extensional process.

The Japan Sea is an example where the continental crust was not thickened before the opening and where far-field stresses, due to intracontinental deformation (India-Asia collision), control the geometry of opening (Jolivet

et al. 1990a, 1994c; Kimura & Tamaki 1986; Tamaki 1988).

The Mediterranean back-arc basins (Tyrrhenian and Aegean Seas) (Fig. 2) opened during the Neogene in an overall context of collision between the African and European Plates (Dercourt *et al.* 1986; Dewey *et al.* 1989b). Back-arc extension is active together with collapse of thick crustal domains (Reutter *et al.* 1980; Le Pichon 1981; Dewey 1988; Serri *et al.* 1993; Gautier & Brun 1994a). Extension initiated after the formation of a thick Alpine-Hellenic crustal wedge. While in the Aegean and Southern Tyrrhenian Seas, subduction involved Mesozoic oceanic lithosphere (Lallemant *et al.* 1994), in the Northern Tyrrhenian Sea a thinned continental lithosphere underthrusts the Northern Apennines (Dewey *et al.* 1989a). Moreover, in the Tyrrhenian area, subduction developed perpendicular to the African motion suggesting that sinking of the Adria-Ionian lithosphere is simply 'passive' (Patacca & Scandone 1989). Despite the continental nature of the subducting lithosphere in the north, slab roll-back and back-arc extension were active during the Neogene and are still active today (Malinverno & Ryan 1986). The Mediterranean area is therefore a favourable site to investigate the contributions of near- and far-field stresses to the back-arc opening process.

The Japan Sea

Tertiary extension along the eastern margin of Asia has been attributed to two possible causes: (1) classical back-arc extension, thus a direct effect of the Pacific subduction (Uyeda & Kanamori 1979; Taylor & Hayes 1983); or (2) India-Asia collision (Tapponnier *et al.* 1982; Kimura & Tamaki 1986; Jolivet *et al.* 1990a). How subduction induces back-arc opening is not completely understood but a low mechanical coupling between the two plates is required. This low coupling can be due to retreat of an old, thermally mature, slab (Uyeda & Kanamori 1979) or to a decrease of the convergence velocity (Northrup *et al.* 1995). The Eocene-Miocene period was probably a period of slow convergence between the Pacific and Eurasia Plates (Northrup *et al.* 1995). The period of maximum extension (formation of back-arc basins between 30 and 15 Ma) is, however, not coincident with the period of slowest convergence (Eocene). Thus, this problem is not yet totally understood. The Japan Sea is probably the best studied back-arc basin, and it can therefore be used as a case example of the interaction between subduction and collision in shaping such basins.

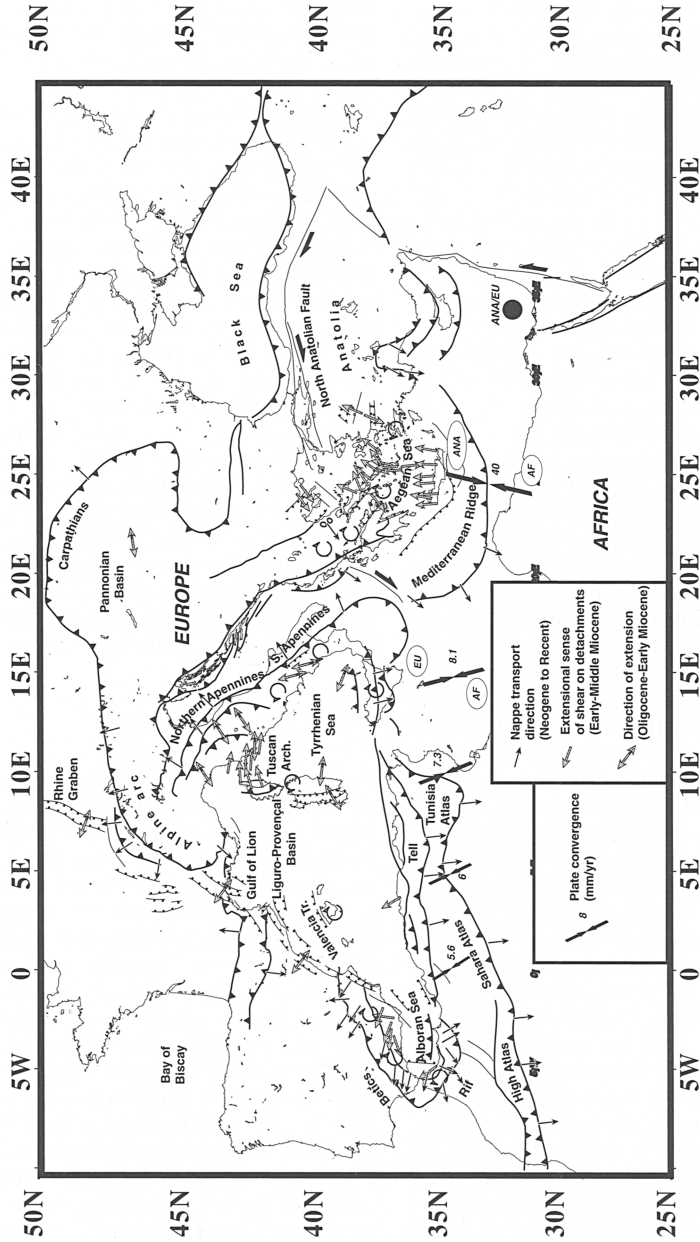


Fig. 2. General tectonic map of the Mediterranean region. Main thrust fronts, Miocene kinematic directions in the Mediterranean region from the Neogene to the Present (Frizon de Lamotte *et al.* 1991, 1995; Gautier & Brun 1994*a, b*, 1996, 1999; Jolivet *et al.* 1994*a, b*, 1995; Jolivet & Patriat 1999).

Geological Structure

The Japan Sea is situated immediately to the north of the Eurasia–Philippine Sea–Pacific Plates triple junction (Figs 1, 3 and 4). Extension and formation of oceanic crust occurred during the Early and Middle Miocene, after which time the whole region entered a stage of east–west compression 10 Ma ago, ultimately leading to the incipient closure of the basin which started 2 Ma ago (Tamaki *et al.* 1992). The Pacific Plate underthrusts both the Eurasian and Philippine Sea Plates at a velocity of $c. 10 \text{ cm yr}^{-1}$ in the Japan and Bonin Trenches, while the Philippine Sea Plate subducts below the Eurasian Plate at a much slower velocity, between 2 and 4 cm yr^{-1} in the Nankai Trench (De Mets *et al.* 1990; De Mets 1992). Owing to the nearby position of the Philippine Sea Plate/Eurasia (PHSP–EU) rotation pole, the convergence rate of those two plates significantly increases from east to west along the Nankai Trench (Ranken *et al.* 1984; Huchon 1985; Seno *et al.* 1993). Vectors of the Pacific–Eurasia (PAC–EU) relative motion are almost perpendicular to the Japan Trench, while a significant obliquity is observed in the Nankai Trench for the PHSP–EU motion. This obliquity is partitioned between the nearly perpendicular convergence in the Nankai Trench and the dextral motion along the major tectonic boundary of southwest Japan, the Median Tectonic Line (MTL) (Fig. 4).

Deep waters are distributed within three deep basins (Fig. 4). The Japan Basin is flooded by oceanic crust (Tamaki 1988). Its triangular shape is the consequence of a propagation of oceanic rifting from east to west (Tamaki *et al.* 1992). The Yamato and Tsushima Basins are instead flooded with attenuated continental crust injected by basic magmas (Tokuyama *et al.* 1987).

A large rifted continental block, the Yamato Bank, remains isolated between the Yamato and Japan Basins. It shows a large aborted north–east–southwest trending rift in its middle part, and is bounded by large normal faults on its northwest and southeast sides. Normal faults with similar trends are found along the northern margin of the Japan Arc. They form dextral en echelon grabens along the western margin of Tohoku (Northeast Honshu) and Hokkaido. North–northwest–south–southeast trending dextral strike-slip faults transfer the motion from one graben to the next (Jolivet *et al.* 1991b).

The same trend of normal faults is found in the Tartary Strait between Siberia and Sakhalin, where they interfere with large north–south trending dextral strike-slip faults (Jolivet *et al.* 1992; Fournier *et al.* 1994; Worrall *et al.* 1996).

A large dextral pull-apart basin has been imaged on seismic reflection data in the middle of the strait [South Tartar Basin (STB); Fig. 4].

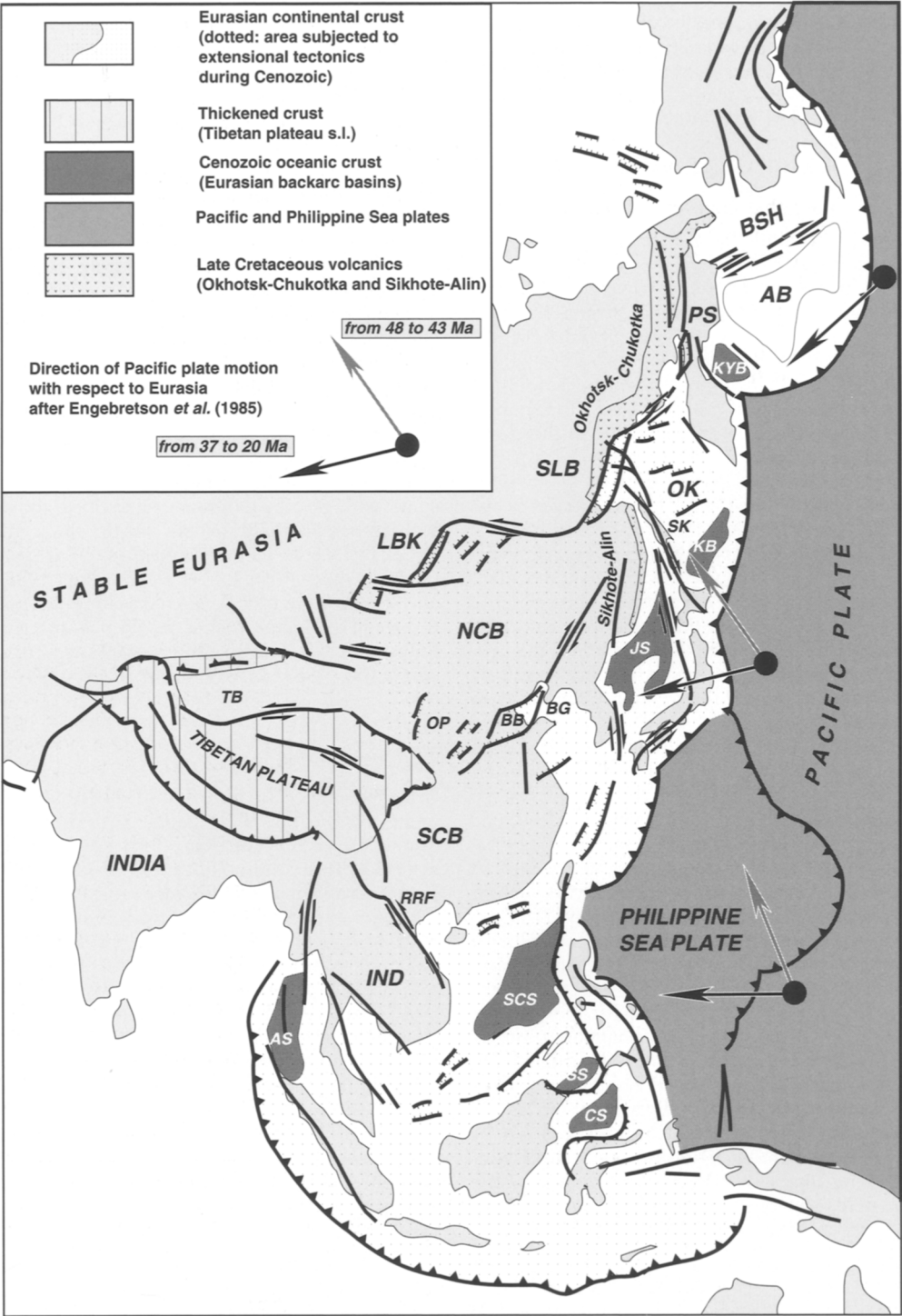
Extensional basins are also found along the MTL. They are filled with Middle Miocene sediments. Their geometry and the analysis of brittle deformation within them suggest that the MTL was a left-lateral, strike-slip fault with a large extensional component during the Middle Miocene (Fournier *et al.* 1995).

Major dextral shear zones bound the Japan Sea to the east and west. The Yangsan and Tsushima Faults were dextral during most of the Cenozoic (Sillitoe 1977). The Pohang and Tsushima (Ulleung) Basins formed during the Early and Middle Miocene as dextral pull-aparts (Lee & Poulet 1988; Yoon *et al.* 1997), and are still active (Jun 1990). The Sakhalin–Hokkaido Shear Zone runs from central Japan to northern Sakhalin and further north to the mainland of Siberia for >2500 km (Jolivet *et al.* 1994c; Worrall *et al.* 1996). It is transtensional in the south with en echelon grabens described above. It is transpressional in the north along the Hidaka Ductile Shear Zone in Hokkaido (Kimura *et al.* 1983; Jolivet & Miyashita 1985), and several large faults and en echelon folds and thrusts in Sakhalin, such as the Tym–Poronaysk Fault which is known along >600 km (Rozhdestvenskiy 1982, 1986; Fournier *et al.* 1994).

Tectonic timing

Palaeomagnetic data were first used to obtain the timing of opening of the Japan Sea (Otofujii & Matsuda 1983; Otofujii *et al.* 1985, 1991). Large-scale clockwise rotation of southwest Japan indicated fast opening during a very short time span, <1 Ma, some 15 Ma ago. More recent analysis seem to indicate a longer time span and some differential rotation between several blocks in southwest Japan (Jolivet *et al.* 1995; Ishikawa 1997). Analysis of the tectonic history of the margins, the sedimentation in the basin and the depth–age relations, has suggested an opening from the Late Oligocene to the Middle Miocene (Tamaki 1986; Ingle *et al.* 1990; Tamaki *et al.* 1992). Drilling of the basement during two ODP cruises (127 and 128) confirmed those conclusions. The end of opening occurred $c. 10 \text{ Ma}$ ago, when extensional deformation was replaced by east–west compression. Some 2 Ma ago, thrusting started along the eastern margin of the Japan Sea.

It is difficult to reconcile what is known about the relative kinematics of the Pacific and Eurasia Plates with the dextral shear. In any case, it



cannot simply be a matter of shear partitioning behind an oblique subduction zone as the dextral strike-slip system extends far inside the Asian continent obliquely to the Japan–Kuril Trenches. This leads to examination of the influence of the India–Asia collision.

The influence of the India–Asia collision

The distance that collision-related deformation has propagated within the Asian continent is still an open question. A class of models emphasizes the crustal thickening responsible for rising of the Tibetan Plateau (England & Molnar 1990), whereas others put forward strike-slip faults which have propagated progressively to the north as far as the Stanovoy Ranges (Tapponnier *et al.* 1982; Davy & Cobbold 1988). A quantitative study of the amount of sediments deposited within and around Asia during the collision process suggests that between 30 and 60% of the collision process was not accommodated by crustal thickening and formation of significant relief. Lateral migration of continental blocks and extension might thus have accommodated a significant part of the convergence (Métivier 1996). The opening of the Japan Sea has been envisaged as a consequence of the India–Asia collision, also partly controlled by subduction roll-back extensional forces (Jolivet *et al.* 1994c). A recently published tectonic map of the Sea of Okhotsk (Worrall *et al.* 1996) extends further to the north collision-related tectonic features, and confirms the relation between back-arc opening in the Japan–Okhotsk Sea region and continental collision, and the importance of the extensional boundary condition to the east. Most studies of the deformation of the Asian continent were focused on its southern part near the collision zone. Since Tapponnier & Molnar's (1976, 1977) pioneer work little attention has been paid to far-field effects of the collision. Studies of active deformation and quantification of instantaneous motion in Asia recently enriched the debate for or against extrusion tectonics in the immediate vicinity of the indenter (Avouac & Tapponnier 1993; Molnar & Gipson 1996). Several papers, however,

proposed relating the opening of the Japan Sea to the India–Asia collision (Kimura & Tamaki 1986; Jolivet *et al.* 1990a, 1994c). Analogue models, scaled for gravity (Davy & Cobbold 1988; Fournier 1994), have shown that indentation tectonics can be produced far from the collision and dextral shear zones parallel to the eastern 'free boundary'. An additional component of extension (introduced in the experiments by gravitational spreading) induces the formation of basins controlled by those dextral shear zones (Fournier 1994). Those experiments show that the opening of the Japan Sea, partly controlled by strike-slip faults produced by indentation tectonics, is a physically feasible mechanism. Numerical models have so far been unable to reproduce the fault pattern observed in Asia. Any model describing the deformation of the Asian continent should include the formation of the major back-arc basins and major strike-slip faults which are first order facts, besides the spectacular crustal thickening in Tibet.

The Pamir–Baikal–Okhotsk Shear Zone. A simplified version of a new tectonic map of the Sea of Okhotsk (Worrall *et al.* 1996), together with the Japan Sea region (after Jolivet *et al.* (1994)), shows the entire dextral strike-slip system from southwest Japan to the northern Sea of Okhotsk where it meets a large-scale left-lateral system (Fig. 4). A striking feature is the 500 km long Shantar–Liziansky Basin along the northwestern margin of the Sea of Okhotsk, where basin-forming extension was active from the Eocene to the early Miocene. Extension is transferred at the southern and northern ends of the graben to strike-slip displacement along the left-lateral Stanovoy Fault (Tapponnier *et al.* 1982), which connects the Shantar–Liziansky Basin to the Baikal Rift. A steeply dipping shear zone of inferred left-lateral strike-slip origin, parallel to the northern margin of the Sea of Okhotsk, connects the Shantar–Liziansky Basin to Tertiary faults in the Pustarets and Penzhina Basins. Seismic data show that normal faults in the Shantar–Liziansky Basin accommodated a minimum of 15–20% extension (Worrall *et al.* 1996). The apparent offset of the Late Cretaceous

Fig. 3. Tectonic map of Asia from the India–Asia collision zone to the Bering Strait [modified after Worrall *et al.* (1996)] and the directions of Pacific–Eurasia relative motion (Engelbreton *et al.* 1985). AS, Andaman Sea; BB, Bohai Basin; BG, Bohai Gulf; AB, Aleutian Basin; BSH, Bering Shelf; CJCZ, central Japan collision zone; CS, Celebes Sea; DD, Derugin Deep; IP, Izu Peninsula; ISTL, Itoigawa–Shizuoka Tectonic Line; JS, Japan Sea; KYB, Komandorsky Basin; KB, Kuril Basin; LB, Liziansky Basin; LBK, Lake Baikal; MTL, Median Tectonic Line; NCB, north China; OK, Sea of Okhotsk; OP, Ordos Plateau; PS, Pustarets Basin; RRF, Red River Fault; SAF, Sikhote Alin Fault; SB, Shantar Basin; SCB, South China Block; SCS, South China Sea; SK, Sakhalin; SLB, Shantar Liziansky Basin; SS, Sulu Sea; STB, South Tartar Basin; TB, Tarim Basin; TB, Tsushima Basin; TTL, Tanakura Tectonic Line; YB, Yamato Basin; YBK, Yamato Bank; YF, Yangsan Fault.

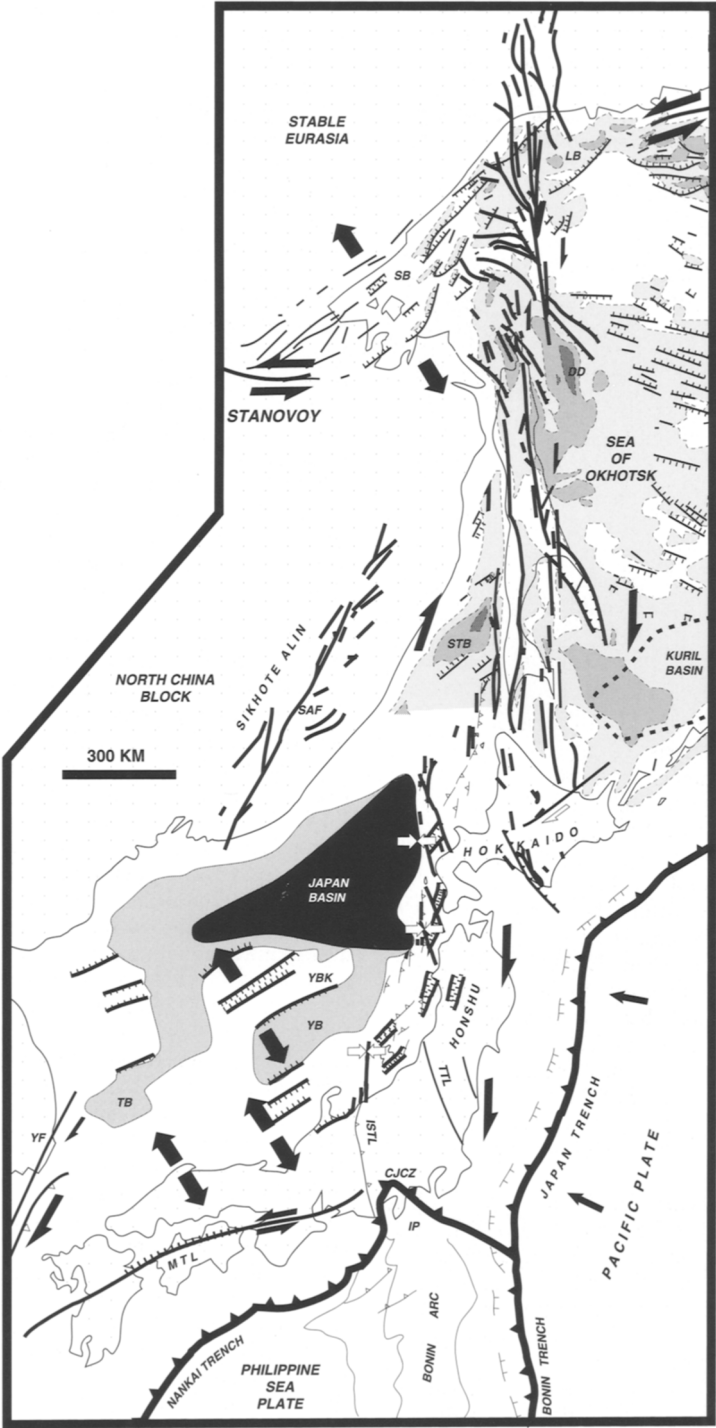


Fig. 4. Tectonic map of the Japan and western Okhotsk Seas (Jolivet *et al.* 1994c; Worrall *et al.* 1996). White convergent arrows along the eastern margin of Japan Sea indicate direction of recent shortening. For abbreviations see legend of Fig. 3.

Okhotsk–Chukotka and Sikhote Alin volcanic belts (Fujita & Newberry 1982; Parfenov *et al.* 1993) likewise suggests left-lateral shear along the northern edge of the Sea of Okhotsk. This succession of extensional basins and strike-slip faults is kinematically compatible with the left-lateral shear zone that runs from the Pamir Range to Lake Baikal (Davy & Cobbold 1988).

Right-lateral Shear Zones along the Pacific Margin. North trending dextral shear zones controlled the opening of intracontinental or back-arc basins as dextral pull-aparts (the Bohai Basin, the Japan Sea) during the Tertiary (Kimura *et al.* 1983; Lallemand & Jolivet 1985; Jolivet *et al.* 1994c). This situation prevailed in Japan until the active east–west compression took over in the late Miocene at 10 Ma. The Sakhalin–Hokkaido Shear Zone extends into the northern Sea of Okhotsk, where it cuts through the Shantar–Liziansky Basin and dies out further north. Transensional dextral motion until the middle Miocene changed to transpressional motion from the late Miocene to the Present in the north (Worrall *et al.* 1996). The dextral shear zone is thus traceable for a distance of >2500 km. Parallel to the Pacific margin in the south, it cuts right through the continent further north. On the basis of a comparison with analogue experiments (Davy & Cobbold 1988), it has been proposed that the left- and right-lateral shear zones both accommodated the internal deformation of Asia due to collision (Jolivet *et al.* 1990a). The new data in the Okhotsk Sea reinforce the comparison with those experiments, and some other more recent studies (Fournier 1994; Jolivet *et al.* 1994c) which show the contemporaneous formation of north–south dextral shear zones with an extensional component along the eastern border of Asia, and northeast–southeast sinistral ones linking the northwest corner of the indenter (western Himalayan syntaxis) to the northeast corner of the indented lithosphere (Okhotsk Sea region).

Timing of Deformation in Northeast Asia. The extrusion model for the India–Asia collision was originally based upon plasticine experiments (Tapponnier *et al.* 1982) which suggested the progressive extrusion of continental blocks along left-lateral shear zones, with a northward propagation of deformation from the Red River Fault to the Stanovoy Ranges. Three lines of evidence now seem to show that the Pamir–Baikal–Stanovoy Shear Zone might instead be an early product of the India collision. Firstly, the age of the inception of the Baikal Rift is clearly pre-Neogene (Kashik & Mazilov 1994; Logatchev &

Zorin 1987), and arguably Eocene; see reviews in Logatchev (1993) and Houdry–Lémont (1994) (Table 1). Two stages of Baikal Rift development are generally recognized. An early stage of slow extension, starting between Late Cretaceous and Late Oligocene is followed by a second stage of more rapid subsidence and extension in the Pliocene–Quaternary. Secondly, dextral shear in Sakhalin and in the Tatar Basin may have started as early as the Eocene, and exposed and subsurface rift grabens in northern Sakhalin contain Eocene non-marine infill above a sharp basal unconformity (Kosygin & Sergeyev 1992; Worrall *et al.* 1996). Finally, subsurface correlations suggest that the early fill in Shantar–Liziansky is also Eocene in age (Worrall *et al.* 1996). While not absolutely conclusive, available evidence now suggests that the Pamir–Baikal–Okhotsk Shear Zone had its inception in Eocene times, approximately coincident with the India collision, and is thus coincident with the earliest, not the latest, development of collision-related shear in northeastern Asia.

Evolving Kinematic Boundary Conditions. The evolution of boundary conditions can be viewed by using stress-field variations in Asia (Fig. 5). The present-day stress field can be roughly deduced from focal mechanisms of earthquakes (Zoback 1992). Figure 1 shows the directions of σ_{Hmax} and σ_{Hmin} deduced from focal mechanisms. The strike of σ_1 is generally northeastward in northeast Asia, except near Japan where it changes to become perpendicular to the subduction zone. In most of northeast Asia, a strike-slip regime is dominant, whereas in Japan, a compressional regime is recorded. The 25 Ma map (Fig. 5) was constructed by using the orientation of large-scale structures, such as rifts of this age as well as more detailed studies of palaeostresses (Fournier *et al.* 1995). The geometry of the stress field is similar to the present one except near Japan. Interpretation of the 45 Ma stage was obtained from the strike of large-scale structures only. The result is consequently less reliable and can be used only for comparison. An older stage (pre-45 Ma) could also be proposed that would show a totally different scheme in northeast Asia, where most of the faults parallel to the eastern border were left-lateral. Thus, major changes appear at 45–43 Ma if the hypothesis for the formation of the large-scale left-lateral Pamir–Baikal–Okhotsk Shear Zone c. 10 Ma, when east–west compression began in Japan, is followed:

- *From 45 to 10 Ma.* The geometry of deformation in the early stages of collision (45–10 Ma) is very similar to that predicted by

Table 1. *Tectonic history of Asia and its eastern boundaries*

Age (Ma)	Pamir-Baikal-Okhotsk shear zone	Himalaya-Tibet	South China Sea	Japan Sea Sakhalin	Northern Sea of Okhotsk	Bering Shelf
45	Left-lateral motion and extension—First episode of rifting	Inception of collision	Rifting	Inception of Pacific subduction	Extension and left-lateral motion	Right-lateral motion and basin formation
32	Rifting and sinistral motion continues	Crustal thickening	First oceanic crust	Rifting and dextral motion	↓	↓
25	↓	Crustal exhumation and first exhumation	Spreading	First oceanic crust, dextral motion	↓	↓
20	↓	South Tibetan detachment; fast exhumation	Kinematic change; inception of subduction	Spreading and extension, dextral motion	↓	↓
12	↓	↓	End of spreading Subduction	End of spreading End of dextral motion, inversion in Tsushima	End of extension	Subsidence and dextral deformation slow
8	↓	Lithospheric root detachment and uplift?	↓	Inception of east-west compression	Continued slow regional subsidence	Minor deformation
2	Fast extension on the Baikal rift; second stage of rifting	Continuous thickening and extension on the plateau	↓	Inception of Japan Sea subduction	Faster extrusion of Okhotsk block?	Deformation ceases

analogue models scaled for gravity (Davy & Cobbold 1988; Jolivet *et al.* 1990a). As in other types of models (Tapponnier *et al.* 1982; Dewey *et al.* 1989a, England & Molnar 1990; Huchon *et al.* 1994; Rangin *et al.* 1995), the eastern boundary is free of stress other than lithostatic.

Such conditions require a frontal subduction of old lithosphere east of Asia (Uyeda & Kanamori 1979). This situation was in fact established only after the abrupt change in the absolute motion of the Pacific Plate at 43 Ma. This major kinematic change in plate motion was recently challenged (Norton 1995), based on a Farallon–Pacific–Antarctica–Africa–North America–Asia Plate circuit. Norton's original observations are that the 43 Ma elbow exists only along the Hawaii–Emperor Chain and that no major tectonic event ever occurred on the margins of the continental plates surrounding the Pacific Ocean. Those two assertions can, however, be disputed: (1) the Louisville Ridge in the southwest Pacific shows the same elbow, as well as most other hotspot lines (Fleitout & Moriceau 1992); (2) a number of events along the Pacific margins could be associated with the 43 Ma event, such as the possible capture of the Philippine Sea Plate (Hilde *et al.* 1977), the inversion of strike-slip sense along the Bering Shelf (Worrall 1991) or the absence of Eocene sedimentation in northeast Japan, as well as the cessation of activity of major left-lateral faults along the margin of northeast Asia (Otsuki & Ehiro 1978). Palaeomagnetic data obtained on the Emperor seamounts suggest that the hotspot might have moved at several cm yr^{-1} during the early Tertiary (Tarduno & Cottrell 1997). The data still allow a significant northward motion of the Pacific Plate before the Hawaiian–Emperor bend. This discussion is placed under the hypothesis that the 43 Ma event actually happened.

The Pacific–Eurasia relative motion was oblique on the plate boundary before 43 Ma

(Engebretson *et al.* 1985). This motion was recorded on land by numerous left-lateral faults (Otsuki & Ehiro 1978). Effects of the oblique subduction were felt far inside the Asian continent, suggesting a strong mechanical coupling between the two plates. It is only when frontal subduction was established that the stress-free boundary

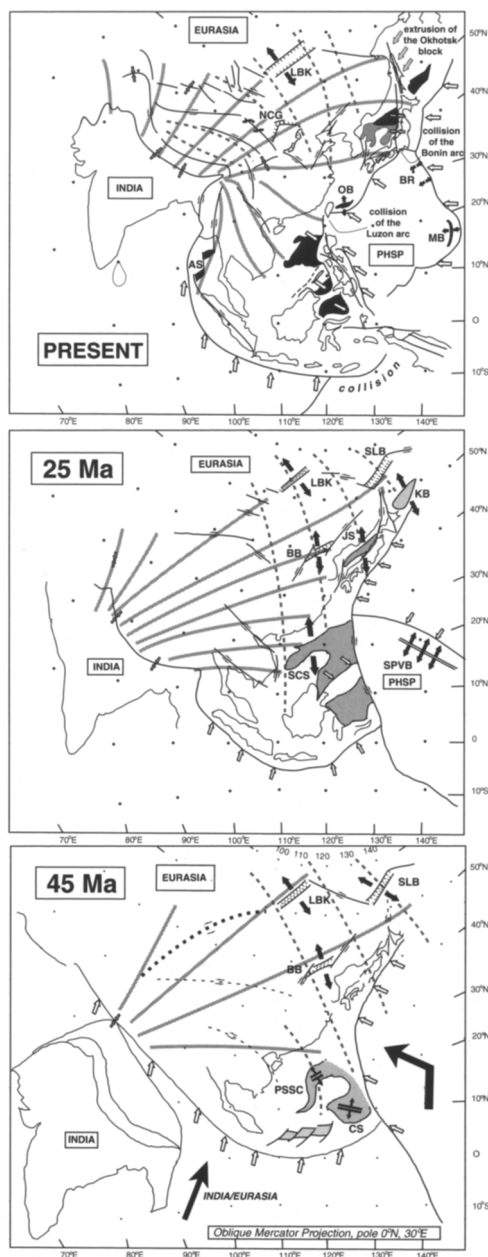


Fig. 5. Possible tectonic evolution of Asia from the first collision to the Present with emphasis on the deformation in Far East and the stress-field variations [principal stress directions in the Present-day configuration were inferred from focal mechanisms taken from Dziewonski *et al.* (1987), Parfenoy *et al.* (1987), Fournier *et al.* (1994) Huchon *et al.* (1994)]. Thick lines represent σ_{Hmax} . See Fig. 3 for abbreviations; SPVB, Shikoku-Parece Vela Basin; PHSP, Philippine Sea Plate; MB, Mariana Basin; PSSC, Proto South China Sea; BR, Bonin Rift; OB, Okinawa Basin.

conditions were met. It can be speculated that at this stage, as soon as these conditions were established, a left-lateral shear zone propagated inside Asia at least as far as the Sea of Okhotsk and perhaps to the margin of the Arctic Ocean.

- *From 10 Ma to the Present.* Compression has been recorded in the marginal basins since the middle Miocene (in the south) or late Miocene (in the north) (Tamaki *et al.* 1992). In Japan, it can be due to the progressive collision of the Bonin Arc, since the late middle or early late Miocene (Taira *et al.* 1989), and to a faster extrusion of the Okhotsk block in the Quaternary (Riegel *et al.* 1993), which might have produced a more efficient compressional component along the dextral Sakhalin–Hokkaido Shear Zone. The most reasonable explanation is that the arrival of the Philippine Sea Plate between the Pacific and Eurasia Plates below central Japan has simply coupled the two larger plates, inducing east–west compression parallel to the Pacific Plate motion in the eastern margin of Asia. Although a large part of the Philippine Sea Plate is thermally mature (in its southwestern part), its northern part is young (<12 Ma for the youngest) and was still younger and buoyant 10 Ma ago. The Pacific Plate is much older all along the western Pacific subduction zone. The arrival of the buoyant Philippine Sea Plate between the Eurasian and Pacific Plates might have considerably changed the mechanical coupling.

Conclusions

Before the present-day tectonic setting, characterized by crustal thickening in Tibet, strike-slip and extensional tectonics dominated in Asia. The left-lateral shear zone that links the western Himalayan syntaxis to the Baikal and Stanovoy Ranges can now be extended through the Sea of Okhotsk and perhaps as far as the margins of the Arctic Ocean. The western margin of the Sea of Okhotsk is characterized by the 2500 km long dextral Sakhalin–Hokkaido Shear Zone, running from central Japan to the northern margin of the Sea of Okhotsk. The deformation related to collision reached the Baikal Graben and the Sea of Okhotsk, and perhaps the Bering Shelf, by Eocene times. The early propagation of a left-lateral shear zone across far eastern Russia might have been favoured by the creation of a stress-free, or extensional boundary, condition east of Asia after the change in the absolute

motion of the Pacific Plate at 43 Ma. The distribution of sinistral and dextral shear zones in northeast Asia is similar to that predicted by analogue models of collision scaled for gravity.

When compared to more simple back-arc basins, such as the Shikoku–Parece Vela Basin, the Japan Sea shows similarities in terms of timing. Both basins opened at the same time, a period when extension prevailed along the entire western Pacific margin. Basin geometries are instead very different and the deformation history of the Japan Basin cannot be explained solely by extension due to slab roll-back. Models involving asthenospheric injection (Tatsumi *et al.* 1989) are alternative explanations for back-arc opening which cannot take into account the asymmetry of the basin and the existence of large-scale strike-slip faults. Intracontinental deformation only can explain the observed strike-slip component. The simplest model, described here, involves the India–Asia collision which modifies the stress regime in the back-arc region and leads to the observed strike-slip history. Extension is fundamentally a consequence of subduction, slab roll-back and/or asthenospheric injection, and strike-slip a consequence of collision.

Northern Tyrrhenian Sea

Geological setting

The Tyrrhenian Sea formed during the last 15 Ma in the back-arc region of the Apennines–Calabrian subduction system (Kastens *et al.* 1988; Kastens & Mascle 1990; Sartori 1990) (Figs 2, 6 and 7). Extension affected a continental crust partly thickened during the Alpine Orogeny, as opposed to the nearby Ligo-Provençal Basin where no such thickening has occurred (Burrus 1984; Faccenna *et al.* 1997). A continuum of extension is recorded from the Gulf of Lion and Provence margin in the early Oligocene or even late Eocene, to the present extension in the Apennines (Chamot-Rooke *et al.* 1999; Jolivet *et al.* 1998). Oceanic crust was formed in the Ligo-Provençal Basin during the Early Miocene when Corsica and Sardinia rotated counter-clockwise about a pole located in the Ligurian Sea (Montigny *et al.* 1981; Vigliotti & Kent 1990; Mauffret *et al.* 1995).

Extension migrated eastward and reached Corsica by the Early Miocene, the Corsica Basin in the Middle Miocene, Elba Island in the Late Miocene, and Tuscany and the Apennines in the Quaternary (Jolivet *et al.* 1990b, 1998) (Fig. 8). This continuous evolution is

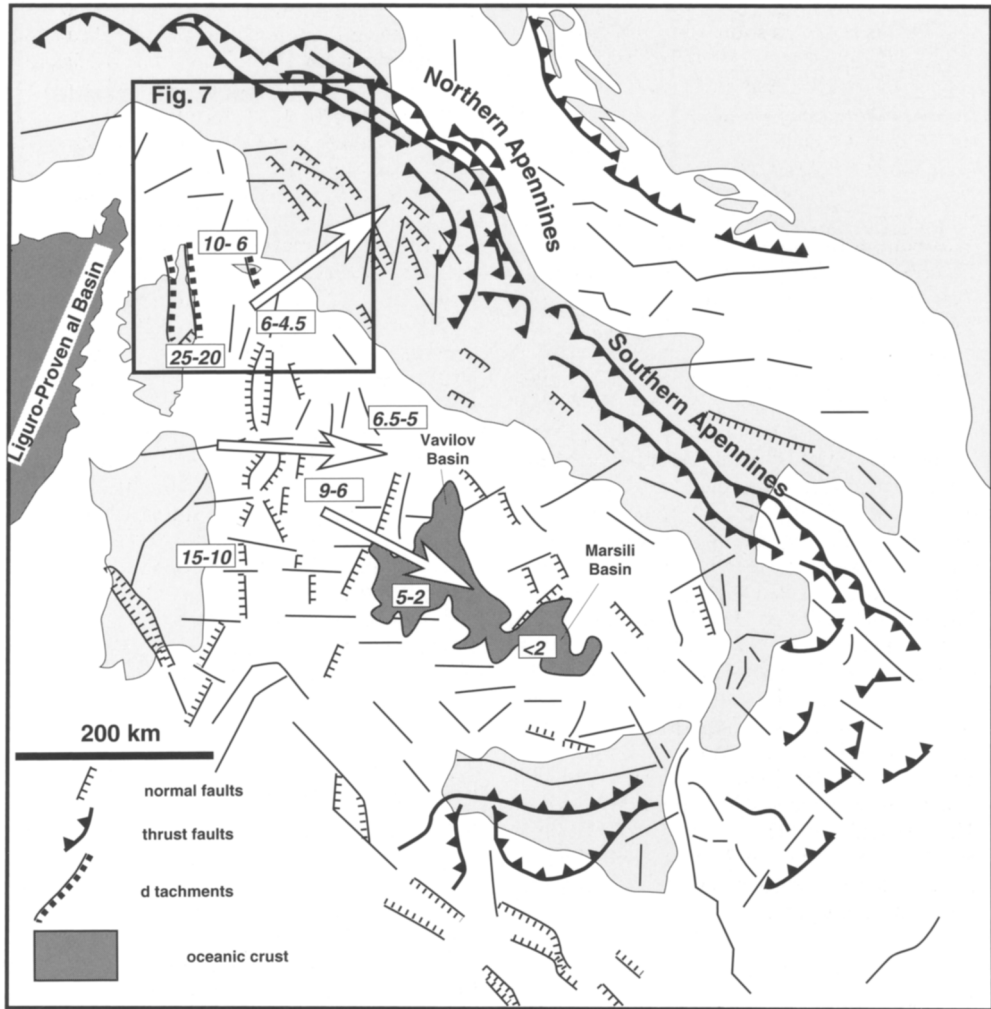


Fig. 6. Tectonic map of the Tyrrhenian Sea and the Apennines. Numbers in square boxes are the ages of extension in Ma (Kastens & Mascle 1990; Jolivet *et al.* 1999).

recorded in the ages of synrift deposits, radiometric ages of exhumed metamorphic rocks and of magmatic events (Elter *et al.* 1975; Serri *et al.* 1993; Bartole 1995; Brunet *et al.* 1997; Chamot-Rooke *et al.* 1997; Jolivet *et al.* 1998). A contemporaneous migration of the magmatic arc is recorded along the same direction from the western margin of Corsica to the western Apennines (Serri *et al.* 1993). During the same period, the Apennines belt was formed as a large accretionary complex at the expense of the sedimentary cover of the Adriatic Plate (Lavecchia & Stoppa 1989; Patacca *et al.* 1990; D'Offizi *et al.* 1994).

Despite this simple kinematic history, the geometry and the mode of extension is different between the two basins: a 'narrow' and 'single-rift' style of extension of the Liguro-Provençal basin contrasts with the 'wide' and 'multi-rift' style of the Tyrrhenian Basin. It has been proposed (Faccenna *et al.* 1997) that these two styles of extension depend upon the pre-rift rheology linked with its geological heritage: the Tyrrhenian extension, in fact, occurred on thickened and weakened Alpine crust whereas the Liguro-Provençal Basin developed on a cold and resistant Hercynian crust. Oceanic crust formed only in the southern Tyrrhenian Sea, north of

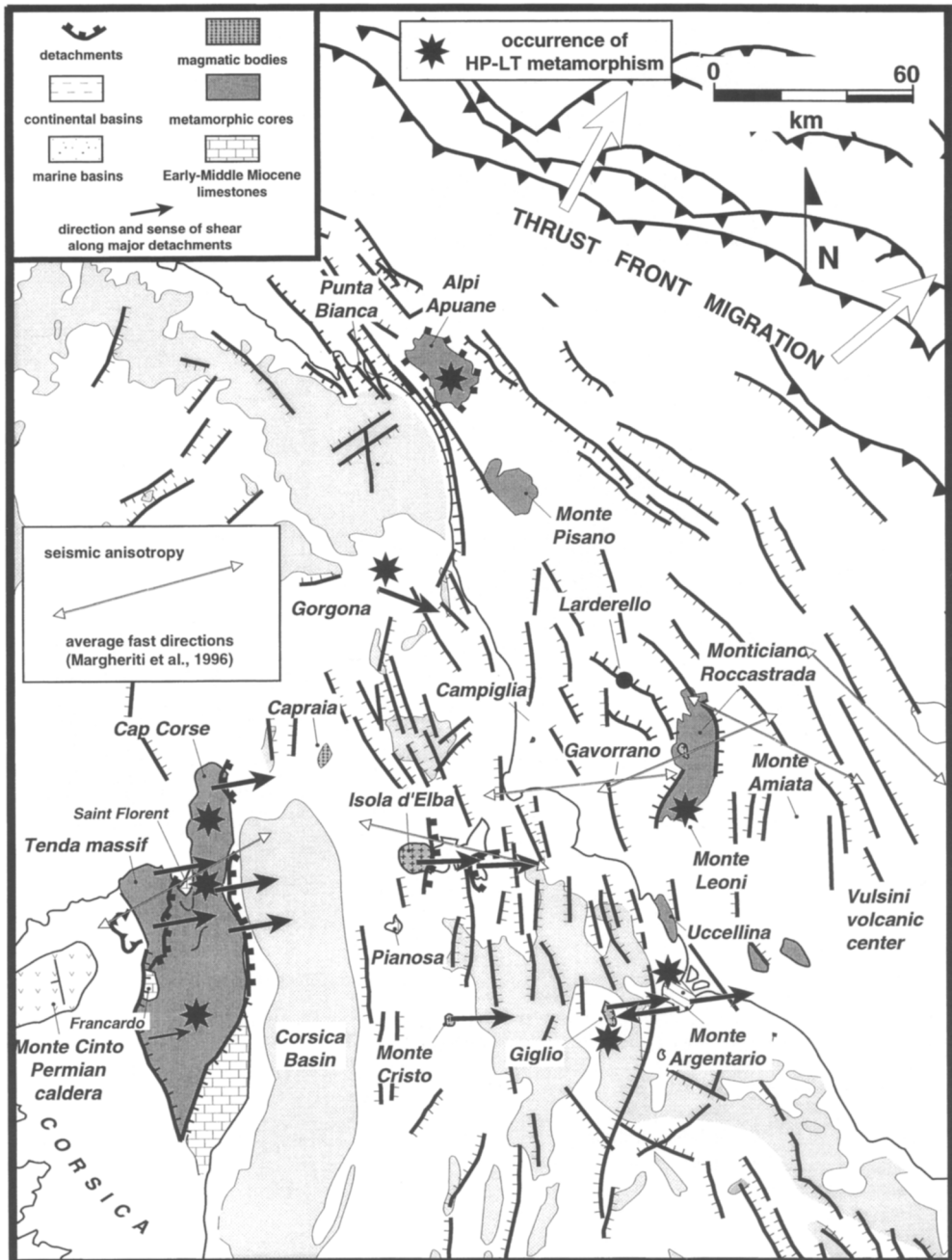


Fig. 7. Tectonic map of the Northern Tyrrhenian Sea [modified after Jolivet *et al.* (1998)].

Sicily, within the Marsili and Vavilov Basins from 5 Ma to the Present (Kastens *et al.* 1988). This fast opening in the southern regions is coeval with fast rotations about vertical axes in the southern Apennines and Calabrian Arcs (Scheepers *et al.* 1993).

Extension in the Northern Tyrrhenian Sea has been active for most of the Neogene (Fig. 8). As opposed to its southern counterpart, no large-scale rotation of crustal blocks is recorded in the palaeomagnetic record (Mattei *et al.* 1996). Only an east-west extension is observed

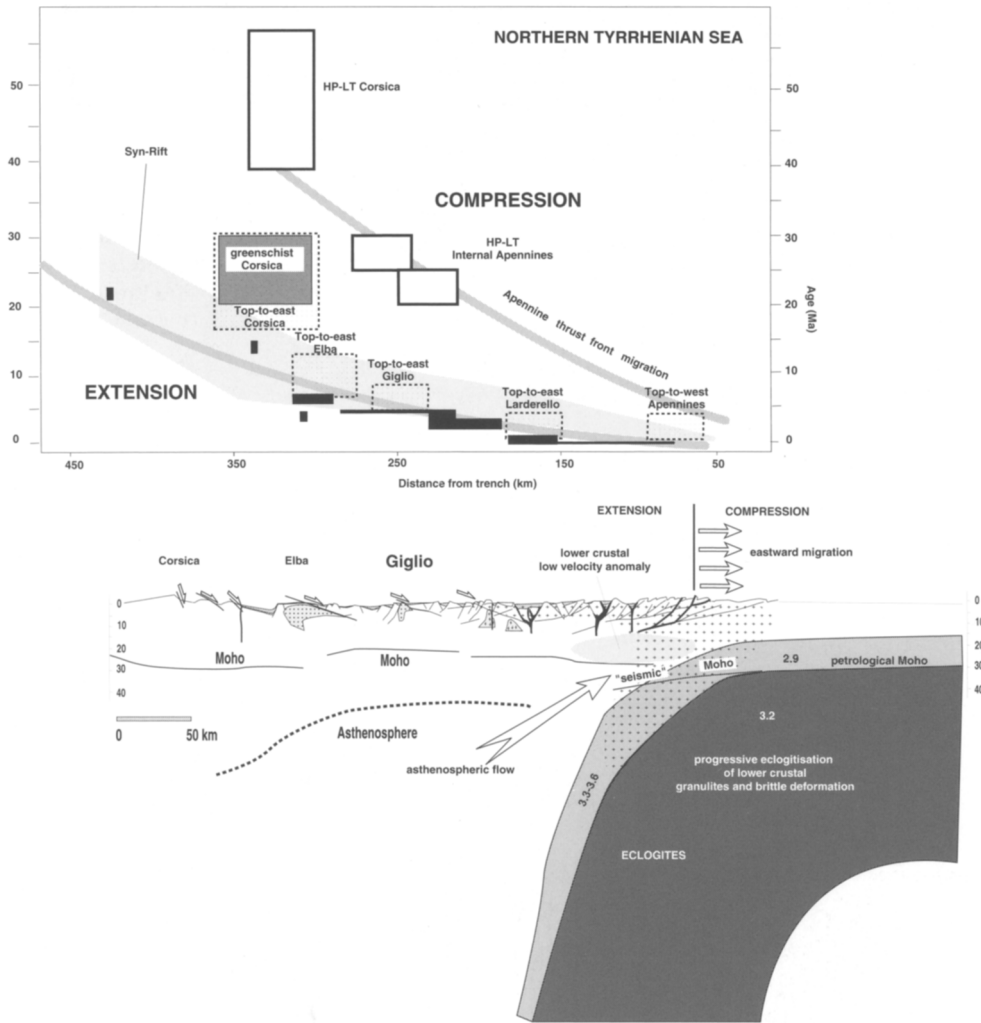
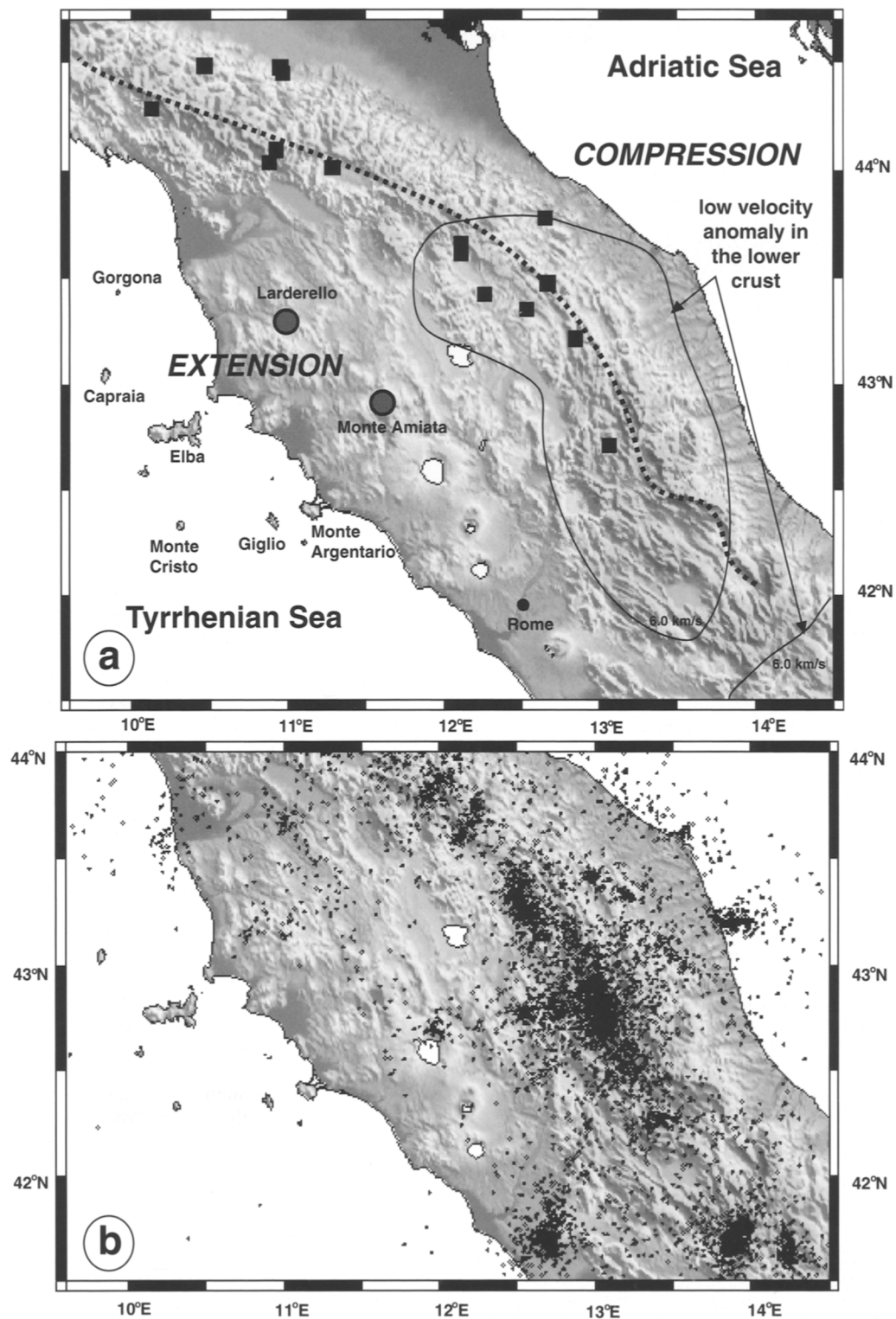


Fig. 8. Cross-section through the Apennines and Tyrrhenian Sea, and the migration of extension and magmatism. (a) Diagram showing the migration of magmatism (black boxes), synrift deposits (grey area), HP-LT metamorphism (white boxes) and activity of extensional shear zones (dotted boxes). (b) Lithospheric-scale cross-section (Wigger 1984; Della Vedova *et al.* 1991; Ponziani *et al.* 1995), seismicity (Selvaggi & Amato 1992), geological structure (Zitellini *et al.* 1986; Lavecchia *et al.* 1987), age of syn-rift deposits (Bossio *et al.* 1993; Martini & Sagri 1993; Bartole 1995), radiometric ages of magmatism (Serri *et al.* 1993) and radiometric ages of metamorphic rocks (Brunet *et al.* 1997).

from Corsica to the Apennines (Jolivet *et al.* 1994b, 1998). The direction of extension is attested by the overall geometry of grabens, and by the shear direction recorded in exhumed metamorphic core complexes of Corsica and the Tuscan Archipelago (Jolivet *et al.* 1998). A clear migration of the locus of extension is evidenced from the Early Miocene in Corsica to the Present in the Apennines. Migration of exten-

sion is coeval with a similar migration of the compressional front to the east and a concomitant migration of the volcanic arc always along an east-west direction. The large-scale geometry is quite simple and can be considered as 2D.

It has been suggested that the difference in style and amount of extension in these two regions corresponds to a difference in style of



subduction (Faccenna *et al.* 1997). From the Oligocene to the Present, only continental lithosphere has been subducting below the Northern Apennines (Coli *et al.* 1991), whereas Ionian oceanic lithosphere is still subducting below the Calabrian Arc (Selvaggi & Chiarabba 1995).

Slab roll-back (Malinverno & Ryan 1986; Royden 1993; Keller *et al.* 1994) is considered the most plausible mechanism for extension in this region. Modelling of the shape of the slab shows that topographic load cannot be responsible for the observed geometry and gravimetric signal, and that subsurface loads at mantle depth (slab) are the acting driving forces for the recent period (Royden 1993). On the other hand, it has been shown that exhumation of high-pressure (HP) and low-temperature (LT) metamorphic rocks was active in the frontal domain until the middle Miocene and that HP–LT metamorphism had also migrated eastward with time (Jolivet *et al.* 1998). Extension was thus set on a thick crust until at least the Middle Miocene. Topographic loads and body forces leading to crustal collapse were thus active in the early history of the Tyrrhenian Sea. It can be shown that both compression in the south (African collision) and crustal collapse are necessary to initiate the subduction process (Faccenna *et al.* 1996). As subduction and extension proceed, the slab-pull component increases as the length of the subducted slab increases.

However, from the Oligocene to the present only continental lithosphere has subducted below the Northern Apennines (Coli *et al.* 1991). The geodynamic situation is thus very different from a classical Mariana-type subduction. Furthermore, although subduction and compression are still active along the trench in the Adriatic Sea, and a drastic shortening observed in the sediments offscraped from the Adriatic crust, no significant crustal thickening is recorded at present (Chiarrabba & Amato 1996). Thus, the continental crust has to be partly subducted (D'Offizi *et al.* 1994; Pialli *et al.* 1995). Eclogitization of the subducted crust could have reduced its buoyancy and permitted its subduction (Pialli *et al.* 1995). The recent discovery of intermediate earthquakes (down to 90 km) below the Apennines, as well as tomographic images, confirm this conclusion (Selvaggi & Amato 1992) (Fig. 9). This deep seismic activity somehow apparently contradicts the presence of a low-velocity anomaly in the lower crust below the Northern Apennines

(Chiarrabba & Amato 1996) (Fig. 9) and a severe attenuation of Pn and Sn waves in the upper mantle (Mele *et al.* 1997).

Compression v. extension and subduction of continental lithosphere

Summarized here are some of the most important aspects of the geodynamics of this region.

- Before the Early Oligocene, extension was active only on the future northern margin of the Gulf of Lion (Burrus 1984), more easterly regions were subjected to compression. The nappe stack and the associated HP–LT metamorphism of Alpine Corsica is evidence of this essentially compressional stage (Matter *et al.* 1981; Caron 1994). From the Late Oligocene onward the Apennines thrust front started to move eastward and basins formed behind by extension. The most spectacular result of this extension is the rifting and spreading of the Liguro-Provençal Basin and concomittant counter-clockwise rotation of Corsica and Sardinia in the Late Oligocene and Early Miocene (Burrus 1984). Extension started in Alpine Corsica some 30 Ma ago (Jolivet *et al.* 1991a, 1998; Daniel *et al.* 1996). It was accommodated by shallow east dipping extensional shear zones at the brittle–ductile transition and normal faults above. Evidence for extension is found further east with a similar geometry and a progressive younging toward the Apennines: Middle Miocene in the Corsica Basin, Late Miocene in Elba and Monte Cristo, Pliocene in Giglio (Rossetti *et al.* 1999), Late Pliocene and Early Pleistocene in Tuscany. At present, extension is active in the Apennines (Frepoli & Amato 1997). During the whole Neogene the direction of extension observed in exhumed metamorphic cores remained unchanged, trending approximately east–west. Recent data on the seismic anisotropy along a transect from Corsica to the Apennines shows that the whole Tyrrhenian lithosphere has been stretched along the same direction (Margheriti *et al.* 1996) (Fig. 7). These results also show a progressive change of the fast direction from approximately east–west in the extended domain to northwest–southeast in the Apennines. This evolution shows the

Fig. 9. Topography and seismicity in the Northern Apennines and Tuscany. (a) Deep earthquakes (Selvaggi & Amato 1992) (black squares) and a low-velocity anomaly in the lower crust (Chiarrabba & Amato 1996). The dotted line represents the drainage divide in the Apennines. (b) All earthquake epicentres.

progressive deformation of the mantle and lower crust from the front of the belt to the back-arc domain.

- The thrust front has migrated eastward with a somewhat lower velocity (Patacca & Scandone 1989). It can be followed on a map by the migration of the HP–LT metamorphic rocks and by the migration of preserved thrusts in the Apennines. Along the entire transect from Corsica to the Apennines there is little evidence that the continental basement has been involved in the crustal thickening process. Crustal thrust sheets are included in the Schistes Lustrés Nappe of Alpine Corsica (Mattaue *et al.* 1981) but further east the Palaeozoic basement crops out only in the Apuan Alps, with no indication of deep crustal lithologies (Carmignani & Kligfield 1990). In all metamorphic complexes observed in the Tuscan archipelago and in Tuscany, HP–LT metamorphic rocks are metasediments of Permo-Triassic age (Verrucano facies) without any trace of a continental basement. The frontal Apennines are made of imbricated thrust sheets of sedimentary origin scrapped off the subducting Adriatic crust (D'Offizi *et al.* 1994). Nowhere is the Adriatic continental crust involved in the shortening process as seen at the surface. The upper crust might still be involved at depth below the Apennines but at least a part of the crust, and certainly the lower crust, has to be subducted in the asthenosphere.
- The transition between compression and extension coincides today with the highest elevation all along the Apennines belt (D'Agostino *et al.* 1998). The example of the Gran Sasso Range is very characteristic of this behaviour. The Gran Sasso basal thrust is reworked at shallow depths by a shallow west dipping normal fault which crops out on the internal flank of the mountain range. Steeper west dipping normal faults, which affect the entire brittle crust, are then found at close proximity further west. Major short-wavelength (30 km) topographic gradients are related to active normal faults. This geometry is different to that observed in the Himalayan–Tibetan belt or the Andes, where the transition between compression and extension coincides with a given topographic contour (Molnar & Lyon-Caen 1988). Regions situated above 3000 m show active extension, while compression is recorded below. This might suggest a general sudden collapse of the entire belt (Platt & England 1994). The geometry observed in the

Apennines instead suggests that a wave of extension progressively lowers the altitude of the belt being formed further east. This is compatible with the eastward migration described above for the Neogene period.

- The present day Apennines form a 100 km wide topographic bulge which culminates at Gran Sasso where most of the shallow seismic energy is released. This domain is supported by a crust not thicker than 40 km. It also corresponds to a low-velocity anomaly in the lower crust with P-wave velocities $< 6.0 \text{ km s}^{-1}$ (Chiarraba & Amato 1996). The lateral extent of this anomaly is very close to that of the topographic bulge. To obtain these low velocities, high temperatures must exist in the lower crust below the Apennines at present. This anomaly is also located east of the most intensely extended domain in the Northern Apennines.

The location of this anomaly is somehow in contradiction with the location of the recent volcanic arc further to the west. Volcanic rocks recently erupted have a clear mantle signature with almost no crustal contamination (Serri *et al.* 1993). During the Late Miocene and the Pliocene a larger crustal component was present in magmatic rocks. The granitoids of Elba, Monte Cristo and Giglio contain cordierite, and are thought to derive from crustal anatexis. This suggests that the position of the volcanic arc is controlled by the geometry of the slab at depth and the partial melting domain in the lower crust is due to some other mechanism (see below). The absence of a high heat flow at the surface, above the low-velocity anomaly, also suggests that the anomaly is quite recent. It can be speculated that it was previously further west and that it has migrated eastward to the present position only recently. The Quaternary age of the main phase of uplift, where extension was already established, and the geographical distribution of thermal anomalies in the lower crust and upper mantle, suggest a thermal origin for the main topographic bulge of the Apennines.

Despite a hot lower crust, deep earthquakes are found at mantle depths down to 90 km below the belt (Selvaggi & Amato 1992). Four of those earthquakes provided first-motion focal mechanisms: three are extensional and one is compressional. A question then arises: in which material are those deep earthquakes produced? Tomographic data suggest that the slab is continuous below the Apennines, down at least to

250 km (Amato *et al.* 1993). Why then would the earthquakes be recorded only down to 90 km if the subducted continental mantle is continuous down to deeper depths?

- As shown on Fig. 8, the eastward migration of extension and compression proceeds at two different rates, extension migrating faster than compression. The distance between the volcanic domain and the trench thus decreases with time. This decrease can suggest a progressive verticalization of the slab. This is evidence for a major role played by slab pull in the trench retreat process.

Eclogitization of the lower crust

The recent discovery of pseudotachylites associated to eclogitization of granulites in western Norway north of Bergen (Austrheim 1987; Austrheim & Boundy 1994; Austrheim *et al.* 1997), suggest that brittle deformation might occur at great depths during metamorphic phase transformation. This brittle deformation is seen only during the first stages of eclogitization of intermediate to basic granulites. While the transformation proceeds, lower crustal material becomes less resistant and more ductile. Only ductile deformation is seen when the transformation is complete. It is uncertain whether these deformations are the result of sudden volume change or whether they correspond to a more regional strain field imposed by the geometry of plate convergence. The existence of eclogites below the Moho, below the Tibetan Plateau, has also been postulated (Le Pichon *et al.* 1997; Sapin & Hirn 1997), in spite of the presence of a hot and weak lower crust.

Lower crustal material brought to eclogite conditions above 13–14 kbar pressure (P) [at 550°C temperature (T)], can attain high densities, similar to what is expected for mantle rocks (Bousquet *et al.* 1997). The density increase is controlled by the appearance of garnets >500°C and by the transformation of plagioclase into pyroxene. It has been proposed for the Caledonides of western Norway (Dewey *et al.* 1993; Andersen 1999), the Himalayas (Henry *et al.* 1997; Le Pichon *et al.* 1997) and the Alps (Bousquet *et al.* 1997), that a part of the continental crust might be hidden below the Moho which would then mark the phase transition between amphibolite facies and eclogites.

One can postulate that the deep earthquakes seen below the Apennines are within the subducted continental crust and that they show the progressive equilibration of lower crustal material to the P – T conditions of the eclogite

facies. This hypothesis can be paralleled to that proposed by Piali *et al.* (1995).

Lower crustal delamination

This hypothesis has several important implications (Fig. 10).

- It offers the possibility of a delamination of the lower crust together with the subcontinental Adriatic mantle (D'Offizi *et al.* 1994; Piali *et al.* 1995). The delaminated lid has a higher density than the surrounding asthenosphere and it sinks under its own weight leading to roll-back of the slab.
- The delaminated crust is replaced by asthenosphere flowing upward. The remaining crust is thus put in direct contact with the hot asthenosphere and partial melting ensues. The low-velocity anomaly would then be a consequence of this delamination controlling the position of the topographic buldge.
- Partially molten lower crust leads to the diapiric ascent of anatectic granitoids. It also acts as a velocity discontinuity which will localize extensional strain in the lower crust and control shear senses along the brittle–ductile transition (Jolivet *et al.* 1998). During slab roll-back the molten lower crust also migrates eastwards, while the crust molten in previous stages cools back and becomes more resistant. Extension then follows the migration.
- Upper crustal rocks underthrust below the Apennines tend to rise up again once the lower crust has delaminated between the frontal thrust and the first extensional fault (Chemenda *et al.* 1995).

The acceleration of the eastward migration with time is evidence that the slab-pull component is more and more efficient with time, and overcomes the extensional forces due to collapse of the thickened crust. This is also compatible with the progressive steepening of the slab, suggested by the faster migration of extension rather than compression. This smaller contribution of body forces can also be seen in the delay between the inception of extension at a given point along the transect and the extrusion of magmatic rocks at the surface: it suggests that crustal thickness has decreased with time. The maximum pressure recorded in HP–LT metamorphic cores also decreases eastward, which also suggests a thinner crust (Jolivet *et al.* 1998). This is further compatible with the observation that the anatectic component in magmas decreases both with time and eastwards (Serri *et al.* 1993). Last but not least, the exhumation

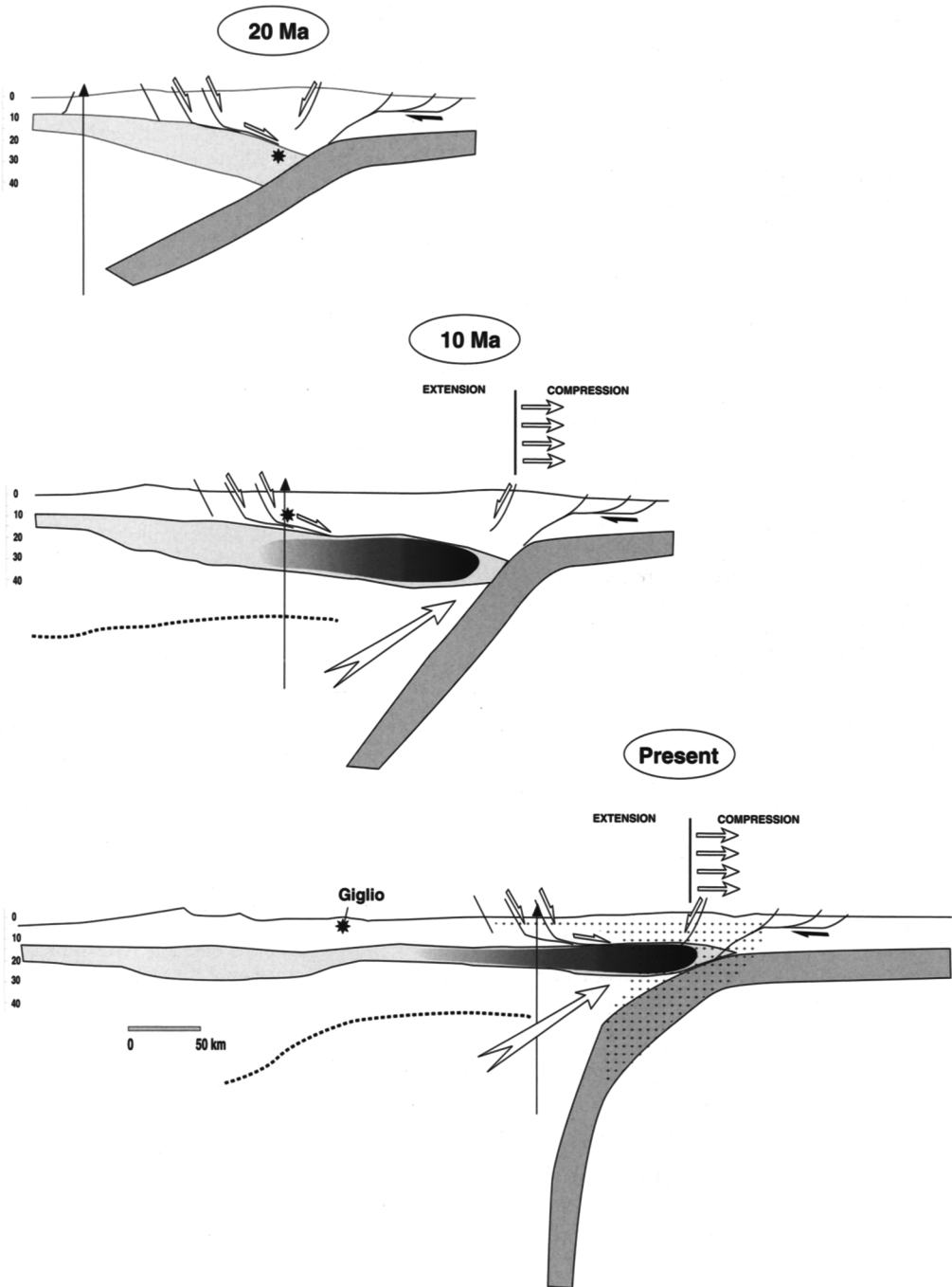


Fig. 10. Schematic evolution of the Corsica-Apennines transect.

of well-preserved HP–LT parageneses in the Tuscan archipelago during the Miocene suggests that the crust was thick enough to prevent reheating before the rocks were exhumed.

Conclusion

Several mechanisms have been proposed to explain the evolution of the Northern Tyrrhenian–Apennine system. It has been suggested that the roll-back of the hinge of the slab, driven by subsurface loads or a slab-pull force, is able to explain the gravimetric signal of the area and the formation of deep foredeep basins (Malinverno & Ryan 1986; Royden 1993). An alternative model proposed by other Carmignani *et al.* (1994) suggests that the collapse of a thickened, post-orogenic crust can contribute to produce extension in the hinterland and compression in the foreland. Topographic loads and body forces leading to crustal collapse were indeed active in the early history of the Tyrrhenian Sea, and can contribute together to the fast retreat of the hinge of the slab (Faccenna *et al.* 1996). However, they cannot alone explain the progressive evolution and migration of the extension–compressional system, especially in most recent times when crustal thickening did not occur.

Some authors (Reutter *et al.* 1980; Lavecchia & Stoppa 1989; D'Offizi *et al.* 1994; Keller *et al.* 1994; Piali *et al.* 1995) proposed that the delamination of the lower crust–upper mantle from the upper crust can better explain the architecture of the Apenninic wedge, allowing a passive subduction of the continental lithosphere. In agreement with this model (D'Offizi *et al.* 1994; Piali *et al.* 1995), it is suggested here that phase transformation and eclogitization of the lower crust is a fundamental requisite for the evolution of the Northern Tyrrhenian–Northern Apennine area, leading to the delamination of the continental crust and stable subduction of the lower crust–upper mantle during the last 30 Ma. The eclogitization of the lower crust is a possible mechanism to explain the limited occurrence of deep earthquakes in the high-velocity zone present below the Apennines. As already proposed for other mountain belts (Andersen & Jamtveit 1990; Dewey *et al.* 1993; Le Pichon *et al.* 1997), this hypothesis suggests that most of the eclogitized lower continental crust is subducted. This does not conflict with the rule that continental crust is generally buoyant and is not subducted. Eclogites are rather rare rocks on the surface of continents and eclogitized granulites are even less frequent. Old portions of buoyant continental crust are not

made of eclogitized granulites but of granulites or high temperature gneiss with some partial melting, remains of post-orogenic collapse of ancient mountain belts.

The occurrence of deep earthquakes below the Northern Apennines is interpreted as the result of brittle deformation during the progressive equilibration of the underthrust Adriatic continental lower crust to the P – T conditions of the eclogite facies. The delamination of the eclogitized lower crust can provide an additional force to drive slab roll-back and extension in the upper plate. In this hypothesis, the Moho seen below the Northern Apennines would be intra-crustal, separating amphibolites from eclogites.

An immediate comparison with the Alboran Sea can be seen (Platt & Vissers 1989; Seber *et al.* 1996). In both cases, the driving mechanism for back-arc extension is the roll-back of a delaminated lid of sinking lithosphere. As argued by previous authors, this mechanism is the only one that can explain the formation of arcs without simple kinematic relations with the Africa–Eurasia convergence. The phase changes invoked here below the Apennines can only enhance the efficiency of a backward motion of the subducting plate. It also has the advantage of solving the problem of the fate of the lower crust which is not found in the accreted units forming the Apennines.

This example shows that slab roll-back is feasible even with the subduction of a continental lithosphere. In the case of the Tyrrhenian Sea, as opposed to the Aegean or Japan Seas, there is no real convergence between the two plates and subduction is only the consequence of passive sinking of the Adriatic lithosphere in the mantle, at least for recent periods. Extension within the upper plate is mainly a consequence of slab roll-back and delamination, its detailed geometry is controlled by that of the pre-existing crustal wedge which is now collapsing. The active faulting in the Apennines is then driven by two concurrent mechanisms: (1) extension induced by the slab roll-back; (2) gravitational collapse of the topographic bulge located above the low velocity anomaly. The relative importance of the two mechanisms depends on the crustal structures undergoing extension: thicker crust will reflect a greater component of the gravitational collapse while extension in the thinner crust would be driven by slab roll-back only.

The Aegean Sea

The tectonic history of the Aegean Sea is partly similar to that of the Tyrrhenian Sea as it

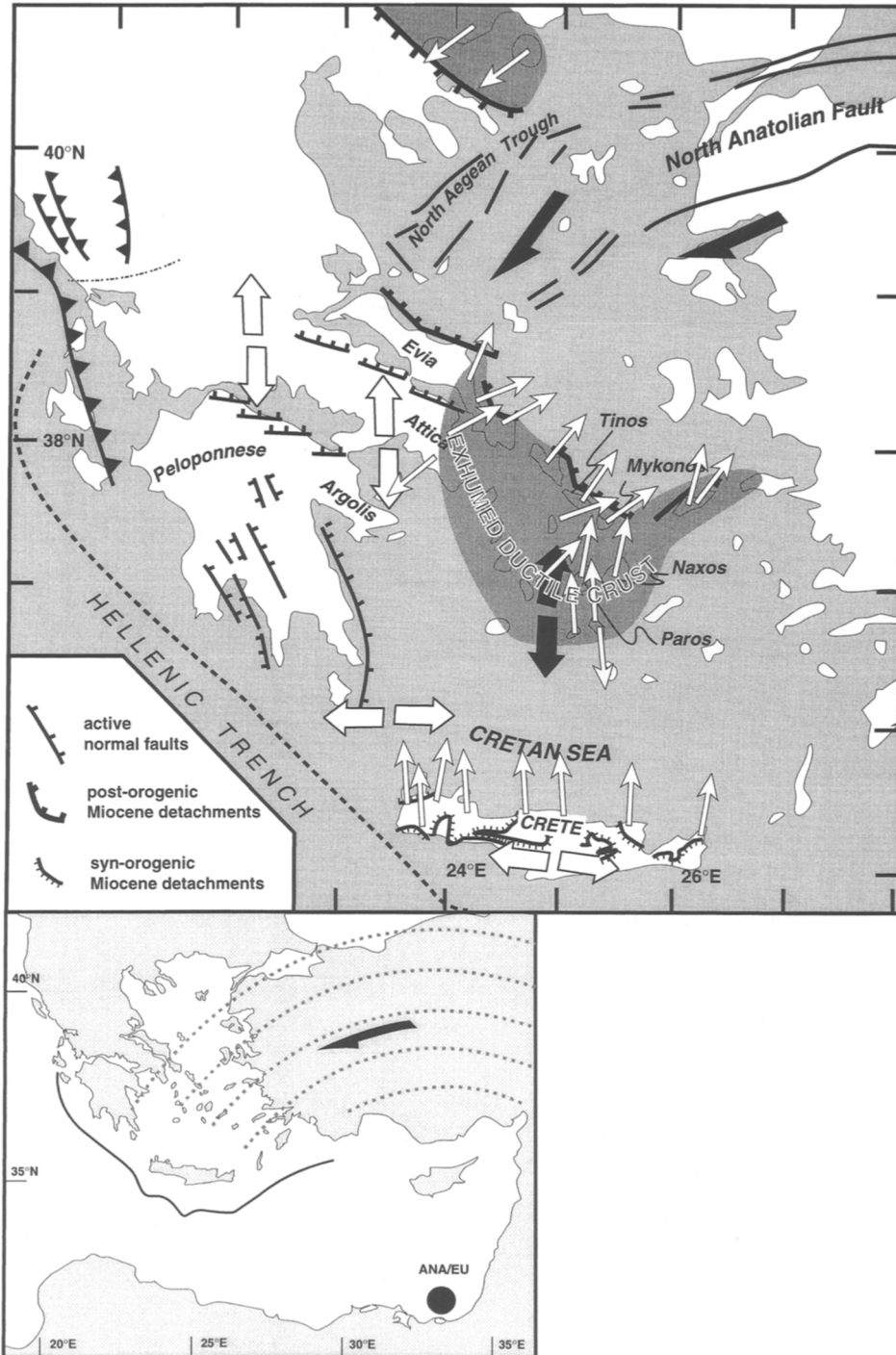


Fig. 11. Schematic tectonic map of the Aegean Sea showing the exhumed ductile crust in the Aegean region and the direction and sense of ductile shear (Buick 1991; Faure *et al.* 1991; Dinter & Royden 1993; Sokoutis *et al.* 1993; Gautier & Brun 1994a, Jolivet *et al.* 1996). The insert shows the rigid extrusion of the Anatolian block documented by geodetic displacements as well as the calculated rotation pole (Le Pichon *et al.* 1994).

involves the collapse of a thickened crust and the outward migration of the subduction front. However, two major differences are to be noted: (1) extension is more widely distributed in the Aegean Sea, which suggests a larger component of gravitational collapse, and (2) the recent geometry of extension is partly controlled by dextral strike-slip faults which guide the westward extrusion of the Anatolian block.

Geological setting

The Aegean Sea (Figs 2, 11 and 12) formed in the back-arc region of the Hellenic Trench during the Late Oligocene until Present (Le Pichon 1981; Le Pichon & Angelier 1981; Gautier *et al.* 1993). Subduction of the African Plate beneath the Anatolian block proceeds at a fast rate of $c. 4\text{--}5\text{ cm yr}^{-1}$ (Le Pichon *et al.* 1994). Extension has taken place on a previously thickened continental crust, the deep parts of which are now exhumed at the surface (Lister *et al.* 1984). Subduction of oceanic crust started at least 40 Ma ago, as attested by seismic tomography data (Spakman 1990). Transition from continental collision to oceanic subduction might have released the compressional stresses and favoured gravitational collapse of the thickened crust.

Active extension is principally localized along the outer Hellenic Arc (Crete), in the Corinth Gulf region and around Volos (McKenzie 1972, 1978; Taymaz *et al.* 1991; Hatzfeld *et al.* 1993; Armijo *et al.* 1996; Rigo *et al.* 1996), leading to an intense seismicity. The central part of the extended domain, the Cyclades Islands, is less seismic though it was subjected to high extensional strain during the Miocene.

Large finite extension is recognized in regions of thin continental crust in the North Aegean Trough and the Cretan Sea where the deepest waters occur. It is also important in the Cyclades islands where Cordilleran-type metamorphic core complexes have recently been studied (Lister *et al.* 1984; Buick 1991; Faure *et al.* 1991; Gautier *et al.* 1993; Gautier & Brun 1994a; Jolivet *et al.* 1994a; Jolivet & Patriat 1999). The exhumation of metamorphic rocks is partly the consequence of the Aegean extension, even though a large part of it could be contemporaneous with compressional tectonics (Avigad *et al.* 1997; Jolivet & Patriat 1999). The exhumation of the eclogites and blueschists in the Cyclades Islands is probably older than the formation of the Aegean Sea. The Aegean extension probably exhumed the amphibolites and greenschist facies rocks cropping out in the core of metamorphic core complexes such as those of Naxos and Paros.

Exhumation of HP–LT metamorphic rocks occurred in Crete in the Early–Middle Miocene, while HT–LP rocks were exhumed further north in the Cyclades Islands (Fassoulas *et al.* 1994; Jolivet *et al.* 1996). Exhumation in Crete is interpreted as the result of extension in the upper parts of a thick accretionary complex near the thrust front.

Extension is further controlled by the North Anatolian Fault, a dextral strike-slip fault which runs along the northern border of Anatolia from eastern Turkey to the Aegean Sea (McKenzie 1978; Le Pichon *et al.* 1994; Armijo *et al.* 1996). It connects to the Aegean Trench through a number of east–west grabens, among which the Gulf of Corinth is the most active (King & Ellis 1990; Rigo *et al.* 1996). Current displacements reveal an almost rigid rotation of the Anatolian block, including the Aegean region, about a pole located in the southeast Mediterranean Sea (Le Pichon *et al.* 1994) (Fig. 11). Internal deformation of the Aegean domain is at present a second-order phenomenon in terms of velocities relative to Eurasia. It could have been a first order phenomenon before the initiation of the north Anatolian Fault in the middle Miocene.

Geometry of extension

Extension was more widely distributed during the Miocene than at Present (Gautier & Brun 1994a). It is recognized in the whole Aegean region as well as in western Turkey in the Menderes Massif (Hetzl *et al.* 1995a,b). The progressive localization with time is probably due to the formation of the North Anatolian Fault and its southward propagation (Armijo *et al.* 1996).

However, active extension in the Corinth Gulf and fossil extension seen in the Cyclades Islands show a very similar geometry with shallow, north dipping ductile shear zones (Jolivet *et al.* 1994a, Jolivet & Patriat 1999; Patriat *et al.* 1999). Active north dipping normal faults that control sediment deposition in the Corinth Gulf are planar down to the depth of the brittle–ductile transition, e.g. 8–10 km (King *et al.* 1985; Jackson & White 1989). Seismogenic low-angle, north dipping normal faults have been recognized in the Corinth Gulf (Bernard *et al.* 1997). They are then relayed by shallow, north dipping extensional shear zones which produce micro-earthquakes (King *et al.* 1985; Rigo *et al.* 1996). Geophysical studies show that the deep crust in the same region is rich in fluid circulation (Chouliaras *et al.* 1997).

Exhumed ductile shear zones of Miocene age in Tinos or Paros show exactly the same features

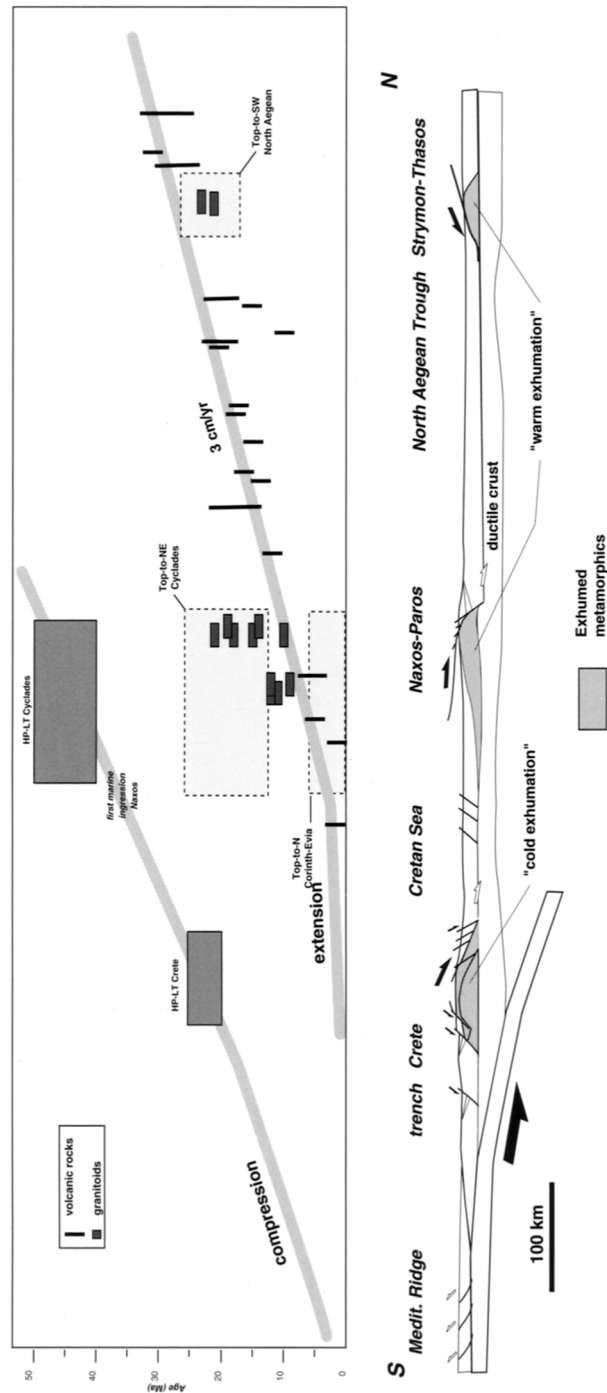


Fig. 12. Cross-section through the Aegean domain showing the major core complexes and the migration of metamorphic and magmatic events. (Altherr *et al.* 1982; Seidel *et al.* 1982; Fytikas *et al.* 1984; Kyriakopoulos *et al.* 1988; Wijbrans *et al.* 1993; Jolivet *et al.* 1996, 1999; Wawrzenitz and Krohe 1997).

(Gautier *et al.* 1993; Patriat 1996; Jolivet & Patriat 1999). Top-to-the-north shear criteria are concentrated along a prominent north dipping shear zone where deformation evolves with time from ductile to brittle. The ductile deformation is contemporaneous with the crystallization of greenschist parageneses that rework earlier (Eocene) HP–LT ones. The shear zones are further characterized by intense fluid circulations best seen in cataclastic levels immediately beneath the brittle fault.

Similar north dipping detachments are observed in most islands of the northern Cyclades Islands, including Mykonos, Paros and Naxos (Avigad & Garfunkel 1989; Buick 1991; Faure *et al.* 1991; Gautier & Brun 1994b). The metamorphic conditions active during extension in those three islands was much hotter than in Tinos and Andros, reaching the amphibolite facies and anatexis. The island of Ios shows a complex interaction between north and south directed shear zones (Vandenberg & Lister 1996). In the northern Aegean region, the southern Rhodopian Massif (Strymon detachment) (Dinter & Royden 1993) and Thassos Island (Sokoutis *et al.* 1993; Wawzenitz & Krohe 1999) also show extensional shear zones with top-to-the-northeast sense of shear.

Recent palaeomagnetic studies of the Aegean granitoids suggest that some of the stretching lineations shown in Fig. 11 have been rotated since the Miocene by some 25° or so, either clockwise (Tinos and Mykonos) or counter-clockwise (Naxos) (Morris & Anderson 1996; Avigad *et al.* 1999). Though it might be too early to have a complete picture of the pre-rotation lineation pattern, it remains that from the central Aegean to Crete all senses of shear point to the north as opposed to those of the Northern Aegean domain which point to the south.

While post-orogenic extension was active in the Cyclades Islands, coeval with high temperature evolution, synorogenic extension occurred in Crete in a much colder environment (Jolivet *et al.* 1996) (Fig. 12). Data from the islands of Syros and Sifnos suggest that a similar history occurred in the Eocene when HP–LT rocks were being exhumed in the Cyclades below north dipping detachments in a cold environment, and HT–LP rocks were exhumed further north in the Rhodopian Massif.

Gravitational collapse v. slab roll-back

A simple model involving the progressive migration of the subduction front from north to south

can be proposed with synorogenic extension above the accretionary complex accompanying post-orogenic extension in the back-arc region (Jolivet *et al.* 1994b; Jolivet & Patriat 1999). The migration of the magmatism documents the roll-back of the subducting slab (Fytikas *et al.* 1984) (Fig. 12).

Rocks were first buried within the accretionary complex near the trench where they recrystallized in the blueschist or eclogite facies. They are then soon exhumed at the surface by the activity of large-scale detachments that dip toward the back-arc region and its low topography. Those which are exhumed soon enough will preserve their HP–LT parageneses, while those which remain at depth longer will see a more or less complete high temperature overprint. During the southward migration of the slab, the accretionary complex grows at the expense of sediments carried by the subducting slab as in the case of the Apennines. The once frontal zones are then progressively transferred to the back-arc domain where the heat flow is higher. The thick crust is then heated and its resistance decreases. It thus collapses and HT–LP metamorphic rocks are exhumed. This process went on from the Eocene–Present more or less continuously.

As shown in Fig. 12, extension is coeval with the emplacement of granitoids. As opposed to the volcanic rocks which migrated regularly from north to south at $c. 3 \text{ cm yr}^{-1}$, granitoids mostly intruded the Cyclades Islands and are grouped in a rather narrow domain which corresponds to the zone thickened during the Eocene.

Conclusions

As in the Tyrrhenian Sea, crustal collapse after initial thickening controls the geometry of extension. Extensional boundary conditions are provided by the retreat of the subducting African slab. Far field stresses due to the Arabia–Eurasia collision also played their part for the recent period as extension became localized and asymmetric as a consequence of the southward propagation of the North Anatolian Fault.

Discussion

Back-arc extension in all the three examples described above suggests that a large component of slab roll-back was involved, and that the geometry of extension and the distribution of strain within the extended domain is further controlled by long-distance effects of continental

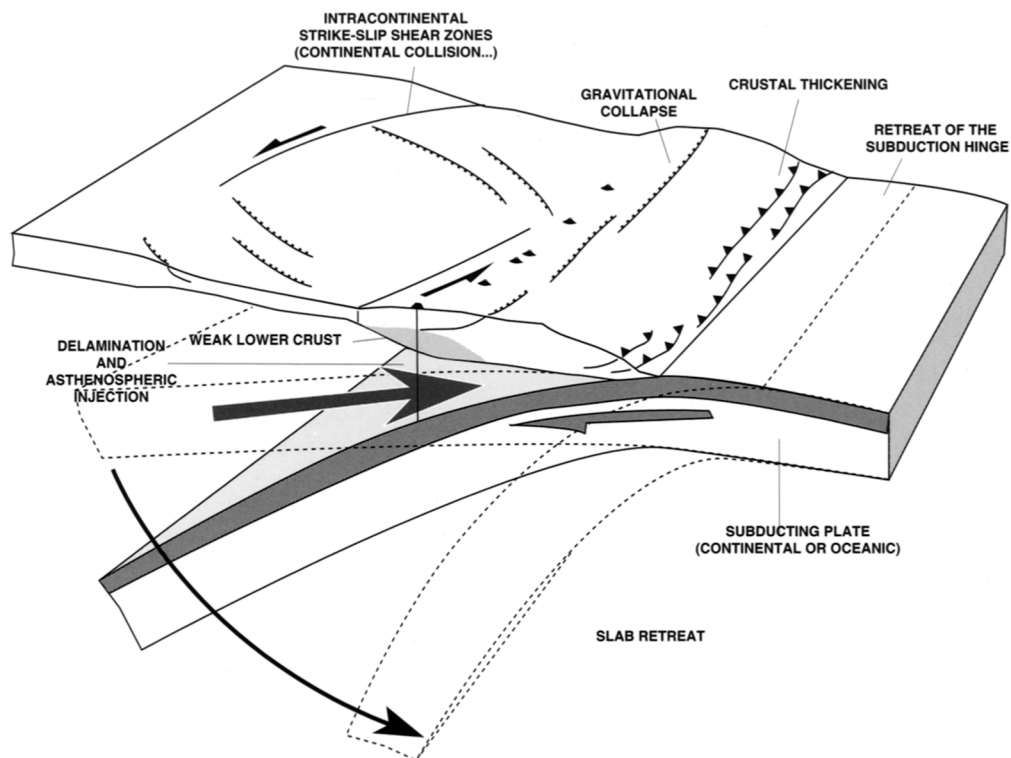


Fig. 13. Schematic diagram showing the various phenomena involved in the formation of marginal basins.

collision and/or body forces during collapse of mountain belts (Fig. 13).

Slab roll-back

The Japan Sea opened during the Miocene above a retreating Pacific subduction, which also promoted the opening of the Shikoku-Parece Vela Basin along the eastern margin of the Philippine Sea Plate during the same period. The end of opening in the Late Miocene and the recent incipient closure of the Japan Sea affects only the Eurasian margin. Extension still prevails above the Bonin (Taylor 1992) and Mariana Trenches at present. Slab roll-back is still active but the degree of mechanical coupling between the Eurasian and Pacific Plates was modified in the Late Miocene by the formation of the central Japan triple junction. The introduction of the young Philippine Sea Plate between the two larger plates, and the collision of the Bonin Arc with central Japan, might have coupled their displacements and induced compression in the upper plate strictly parallel to the convergence vector.

The Northern Tyrrhenian Sea extension is presently active in the Apennines as the con-

tinuation of a long-lived process which started as early as the Oligocene. Slab roll-back has been active during this entire period, even though there has been little east-west convergence between the European and the Adriatic Plates. Passive sinking of the Adriatic continental lithosphere is still active at present. Eclogitization of the lower continental crust might be one factor which increases the density of the lithosphere and permits sinking and roll-back.

The Aegean Sea started to form (30 Ma ago) soon after the inception of an oceanic subduction which replaced continental collision (40–45 Ma ago). The old oceanic lithosphere of the eastern Mediterranean has been subducted below the Anatolian Plate at a fast rate ($4\text{--}5\text{ cm yr}^{-1}$) since the Late Miocene at least. Before the formation of the North Anatolian Fault the rate of convergence Africa–Eurasia was probably lower.

Crustal collapse

The collapse of a thick and warm continental crust leads to widely distributed extension (Wernicke 1992). The Mediterranean back-arc basins

make no exception to that rule. The striking difference between the Liguro-Provençal Basin, where extension is localized and oceanic crust formed in rather early stages and the Northern Tyrrhenian Sea, where a Basin-and-Range topography is observed, is due in large part, to the pre-extension crustal thickness (Faccenna *et al.* 1997; Jolivet *et al.* 1998). In the Tyrrhenian Sea oceanic crust formed in the south only in the latest stages of extension after extreme thinning of the continental crust. Post-orogenic extension lead to the exhumation of metamorphic rocks equilibrated in HT–LP conditions. Exhumation of blueschists and eclogites occurs within the accretionary complex.

Extension is even more widely distributed in the Aegean Sea and core complexes more spectacular than in the Tyrrhenian Sea, although the tectonic histories of the regions are quite similar. Crustal collapse was probably initiated after the transition from continental collision to oceanic subduction some 40 Ma ago, leading to a sudden crustal collapse of the entire thickened domain while the subduction started its southward retreat.

Continental collision

Long-distance effects of continental collision are recorded both in the Japan and Aegean Seas. The recent propagation of the North Anatolian Fault in the Aegean Sea has led to a localization of strain along the fault and at its southwestern termination. The formation is this fault is due to the extrusion of the Anatolian block as a consequence of the Arabia–Eurasia collision (McKenzie 1972).

The geometry of opening of the Japan Sea can be explained as a consequence of large-scale strike-slip shear zones due to the indentation of India into Asia. The asymmetry of opening from the beginning of rifting to the end of spreading is a consequence of the India–Asia collision. This is possible only because a low-stress boundary condition to the east (Pacific subduction) allowed the propagation of strike-slip faults up to the margin. After the formation of the central Japan triple junction, deformation in the upper plate became controlled by the Pacific–Eurasia relative movements and east–west compression ensued.

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