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Models of Mediterranean Back-Arc Basin Formation [and Discussion]

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Models of Mediterranean back-arc basin formation

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During and mainly after the Tertiary continental collisions, regions of extension and subsidence developed within the Alpine–Mediterranean orogenic belt in the rear part of the contemporaneously active arcs. Mediterranean back-arc basins are characterized by a hot upper mantle, overlain by crust transitional from continental to oceanic, which reflects their different stages of maturation. The four basins considered here are, in order of increasing maturity, the Pannonian Basin, the Aegean Basin, the Alboran–south Balearic Basin and the Tyrrhenian Basin. Back-arc extension is not a rigid plate opening but rather an areal expansion associated with progressive bending of the arc. The most successful models suggested for the evolution of Mediterranean back-arc basins imply updoming of the asthenosphere accompanied by lithospheric attenuation, stretching and dyke intrusion. The force behind asthenospheric updoming is under debate; active and passive mechanisms have been suggested. It seems, however, certain that gravity plays an important role in initiating and maintaining back-arc extension.

Basin subsidence is an isostatic response to structural changes of the lithosphere and to conductive decay of the associated heat anomaly. A quantitative model for basin formation can be obtained only if the subsidence history is well documented.

1. Introduction

The western Mediterranean Sea and parts of the eastern Alpine–Mediterranean region can be classified as back-arc basins, because they show the following features:

(*a*) back-arc and/or behind mountain arc setting;
(*b*) zones of extension and subsidence;
(*c*) calc-alkaline volcanism;
(*d*) transitional crust from continental to oceanic, and hot underlying upper mantle;
(*e*) shallow and intermediate depth earthquakes.

In this paper, four basins, the Pannonian, Aegean, Alboran and Tyrrhenian basins, will be considered. Their most important general features will first be summarized; current geodynamic models and their applicability in explaining these features will then be discussed.

2. General features of Mediterranean back-arc basins

Knowledge of the Mediterranean back-arc basins and associated arcs has increased considerably in recent years (see, for example, Nairn *et al.* 1977; Biju-Duval *et al.* 1979; Horváth & Berckhemer 1981), but it is still on very different levels. We summarize here, in seven points, only main features that appear to be generally valid for each basin and are of genetic importance.

(i) Back-arc basins have developed within the Alpine orogenic belt and are bounded by sharp arcs with high curvature. Using the term 'orocline', suggested by Carey (1958) for curved orogenic belts, we can say that back-arc basins are situated at the concave side of oroclines (figure 1). Where the arc is currently dismembered, the original continuity can readily be demonstrated.

In the following discussion, internal and external are terms expressing the position of the structural–statigraphical units of an orogen before deformation, relative to the continental foreland. External is toward the continent (continental margin), internal is toward the ocean (ophiolites, internal flysch and continental fragments derived from oceanic islands or from the opposing continent).

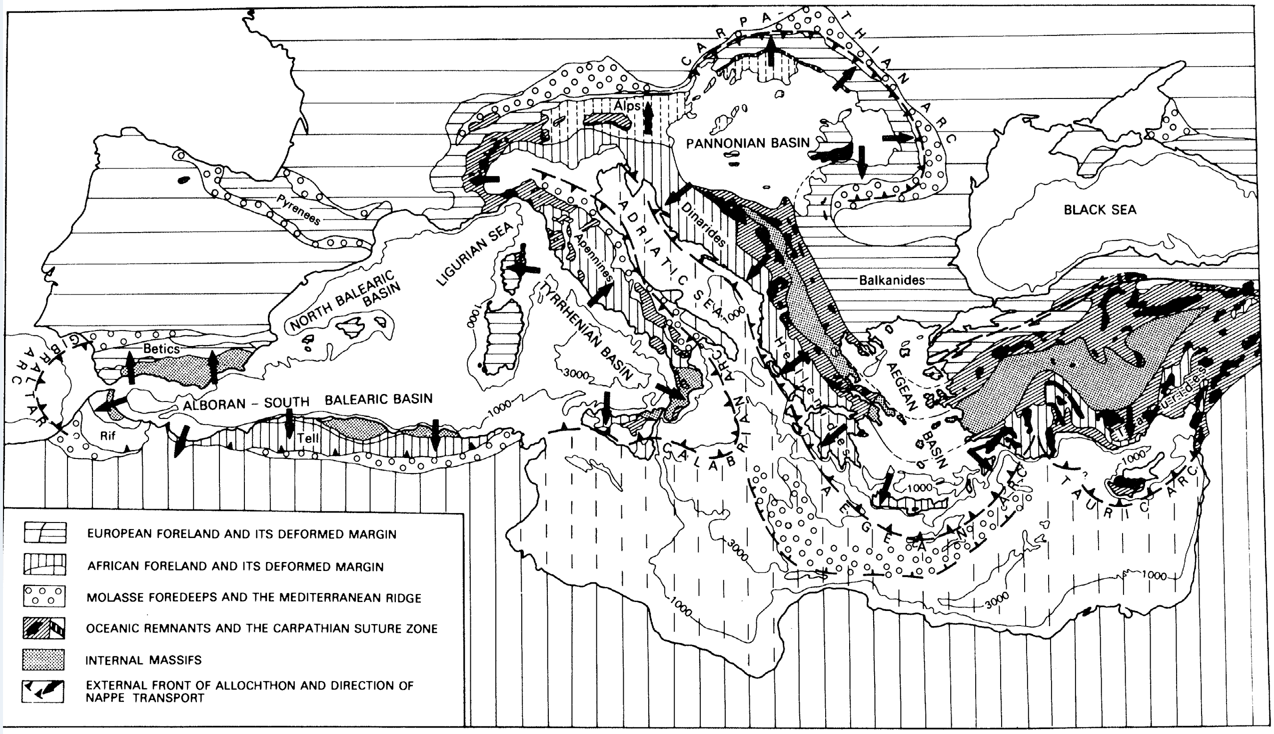

FIGURE 1. Geotectonic setting of Mediterranean back-arc basins (mainly after Channell *et al.* 1979).

The Pannonian Basin, and other smaller depressions at its periphery, are surrounded by the western, eastern and southern Carpathians. The western and eastern Carpathians are divided into an inner and outer belt. The inner Carpathians are discontinuous; large areas have subsided, and others are covered by Neogene volcanics. The outer, morphologically continuous, belt is made up from the Cretaceous–Palaeogene Flysch Carpathians (Mahel 1974; Channell *et al.* 1979).

The Aegean basin is bordered on the west by the Hellenides. Both the external and internal zones can be correlated through the Hellenic Arc and Aegean archipelago with the Taurides, Menderes massif and Pontides of Turkey (Aubouin *et al.* 1976; Brunn *et al.* 1976; Jacobshagen *et al.* 1978).

The Alboran Sea is a narrow basin of the western Mediterranean, which widens to the east toward the south Balearic basin. The Alboran–south Balearic basin is encircled by the Betic Cordillera of southern Spain, which curves sharply across the Gibraltar arc into the Rif and

Tellian Atlas of north Africa (Maghrebides). Internal massifs of the Betics and Maghrebides are situated along the coast of the Mediterranean. A Cretaceous–Palaeogene flysch complex and the '*Dorsale calcaire*' are situated to the exterior of the internal massifs. These two zones clearly demonstrate the original continuity of the Betic and Maghrebid orogens (Durand-Delga 1978).

The Tyrrhenian basin is bordered by the southern Apennines to the east and Sicily to the south. Although the Calabrian internal massif obscures the physical continuity of the southern Apennines toward Sicily, the correlation of external and internal units is well established (D'Argenio & Pialli 1975; Caire 1978).

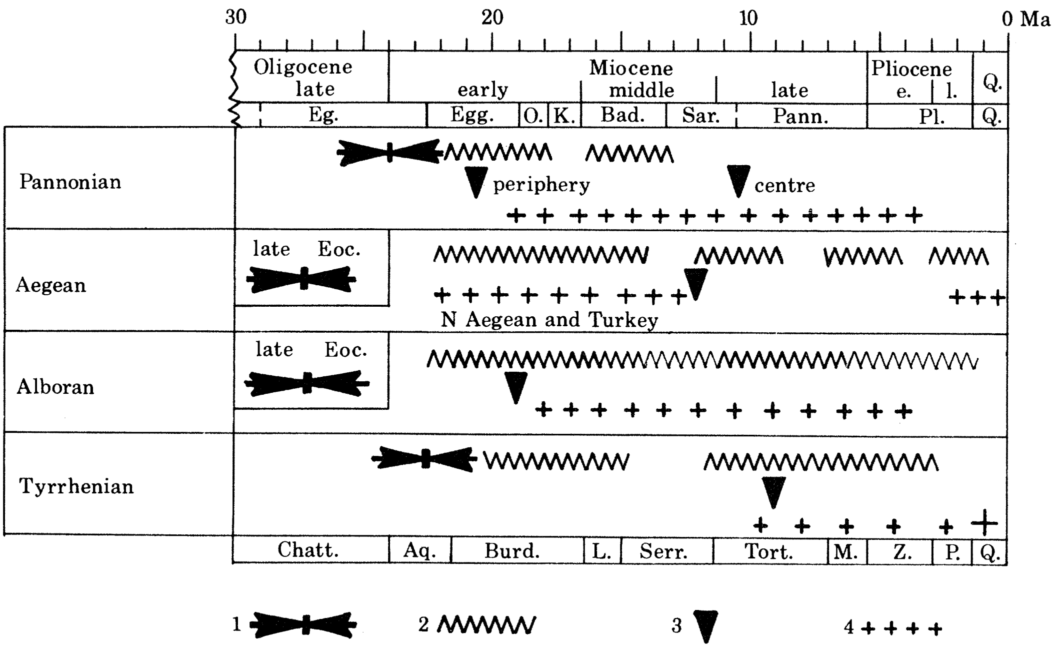

FIGURE 2. Diagram showing the time of main tectonic events of back-arc basins and associated arcs (Horváth & Berckhemer 1980; Le Pichon & Angelier 1979; Jacobshagen *et al.* 1978; Biju-Duval *et al.* 1979; Arana & Vegas 1974; Di Girolamo 1978). The central Paratethys regional stages used for the Miocene of the Pannonian basin are as follows: Eg., Egerian; Egg., Eggenburgian; O., Ottnangian; K., Carpathian; B., Badenian; Sar., Sarmatian; Pann., Pannonian (*sensu stricto*). Keys: 1, continent–continent collision; 2, post-collisional deformation of the external zones; 3, onset of extension and subsidence of the back-arc basin (note that this is very controversial in the Tyrrhenian Basin; see discussion by Hsü 1977); 4, calc-alcaline volcanism (large ' + ' indicates that the peri-Tyrrhenian volcanism is mostly Quaternary).

(ii) In each deformed arc the internal zones are thrust onto the external zones, an event that is usually interpreted as the result of continent–continent collision. Figure 2 gives the timing of continent–continent collision and the initiation of the appropriate back-arc basin formation. Formation of back-arc basins post-dates the initiation of continent–continent collisions and they are superimposed on the collisional suture. They can therefore be called post-collisional episutural basins (Bally 1975).

The Pieniny and part of the outer Carpathian flysch was deposited in a wide trough, the deepest part of which was probably floored by oceanic or suboceanic crust (Radulescu & Sandulescu 1973; Mahel 1978). Shortening of the flysch by subduction went on (possibly

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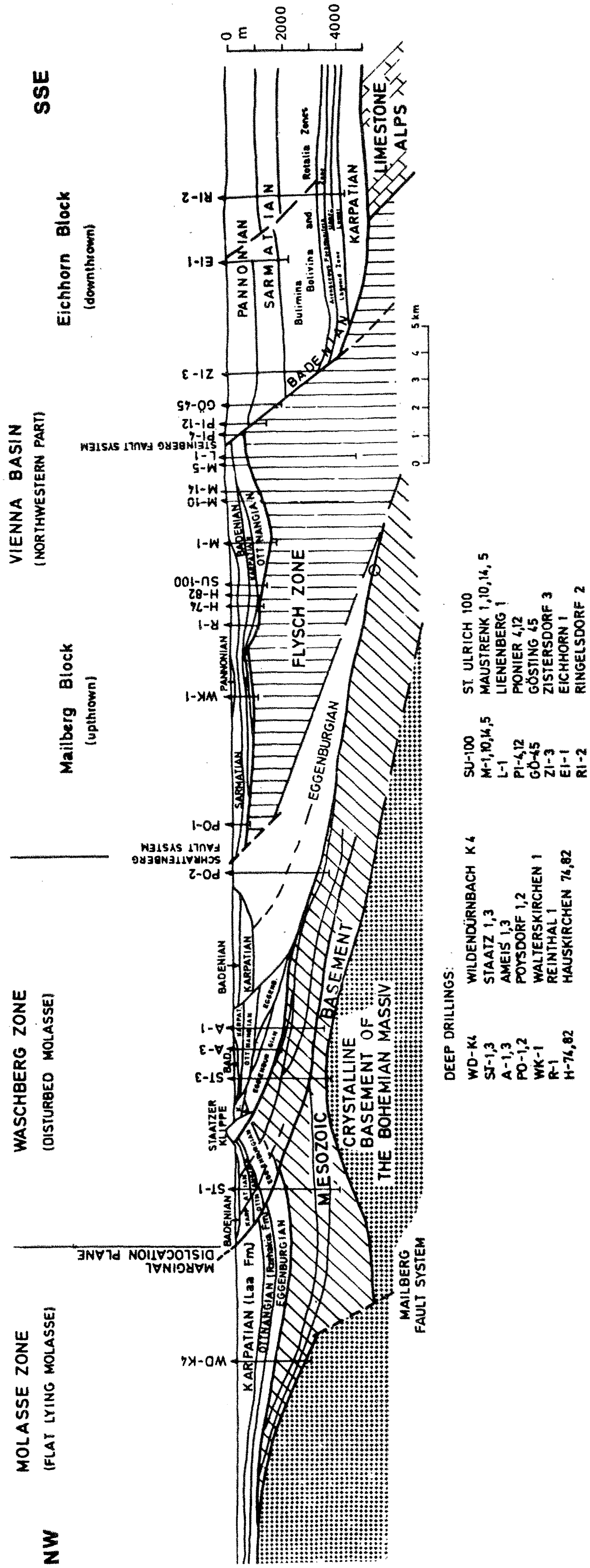

FIGURE 3. Section through the Molasse zone of the westernmost Carpathians and the northern Vienna basin (after Steininger *et al.* 1975).

episodically) during the latest Cretaceous–Palaeogene. The piling up of the flysch nappes and their thrusting onto the foredeep molasse started in the late Oligocene–earliest Miocene, which suggests that the onset of collision of the inner Carpathian–Pannonian plate and the European foreland began at this time (Horváth *et al.* 1977). The Vienna basin started to develop in the early Miocene (Eggenburgian) and is clearly superimposed on the collisional suture (figure 3). The major part of the Pannonian basin lies behind the collisional front. Its basement was deformed mainly during the earlier, i.e. mid-Cretaceous and latest Cretaceous–Palaeocene tectonic phases (Mahel 1974; Channell *et al.* 1979). The subsidence was initiated here during the mid-Miocene with formation of several grabens. The general subsidence of the region started in the late Miocene (Pannonian).

On both sides of the Alboran basin the Betic and Maghrebian ranges have been continuously active since Eocene time. The rather narrow Tethyan arm was subducted under Iberia and, by the late Eocene, the collision of Africa and Iberia occurred, as indicated by the thrusting of internal massifs onto the external zones. The post-collisional extension and subsidence started in the early Miocene (Burdigalian) and affected mainly the allochthonous internal units (Auzende *et al.* 1973; Biju-Duval *et al.* 1979).

In the Hellenic–Aegean region the subduction of the Vardar ocean (the main Tethys) occurred mainly during late Jurassic–early Cretaceous time. Deformation of the internal zones continued in the Middle to Upper Eocene. When obduction of the Vardar ophiolites onto the Pelagonian platform had been completed, the Pindos trough subducted below the Pelagonian platform, which in turn was thrust toward the external zones (Jacobshagen *et al.* 1978). This suggests the disappearance of all oceanic-type lithosphere and the onset of continent–continent collision (Aubouin *et al.* 1976). The subsidence in the Aegean region (northern and southern Aegean) started at around the Middle–Late Miocene boundary and was superimposed on the deformed orogenic belt (Le Pichon & Angelier 1979).

In the western Mediterranean region, an ocean separated Europe and the Adriatic continent during the Cretaceous. From Eocene time, the ocean was subducted to the north, which gave rise to the separation of the Calabria–Sardinia block from Europe and to the opening of the Ligurian Sea. In the late Oligocene or early Miocene, the Calabrian–Sardinian unit collided with the Adria margin, initiating the Apenninic–Sicilian deformation. After the collision, major extension and subsidence began in the late Miocene (mid-Pliocene?) which led to the formation of the Tyrrhenian basin (Hsü 1977; Channell *et al.* 1979). This extension was superimposed onto the suture zone of the Alpine and Apenninic–Sicilian nappe system (figure 4).

As the nature of the eastern Mediterranean crust is not clear, the original connection of the external Sicilian–Apenninic and Hellenic–Aegean units with the African continent is a matter of debate. Several authors believe that these external units have been separated from Africa by an 'external ocean', part of which is preserved in the eastern Mediterranean (Dewey *et al.* 1973; Biju-Duval *et al.* 1977; Hsü 1977). There are, however, several arguments that speak against this possibility.

(*a*) The suggestion that even the most external ophiolitic nappes (Cyprus, Antalya, Crete) came also from the north (from the Vardar and Izmir–Ankara suture of the Tethys), and that their relative autochthon (e.g. Bey Daglari) can be correlated with the Arabian platform and Adriatic foreland (Ricou *et al.* 1975; Brunn *et al.* 1976; Aubouin *et al.* 1976), although this view is not yet generally accepted.

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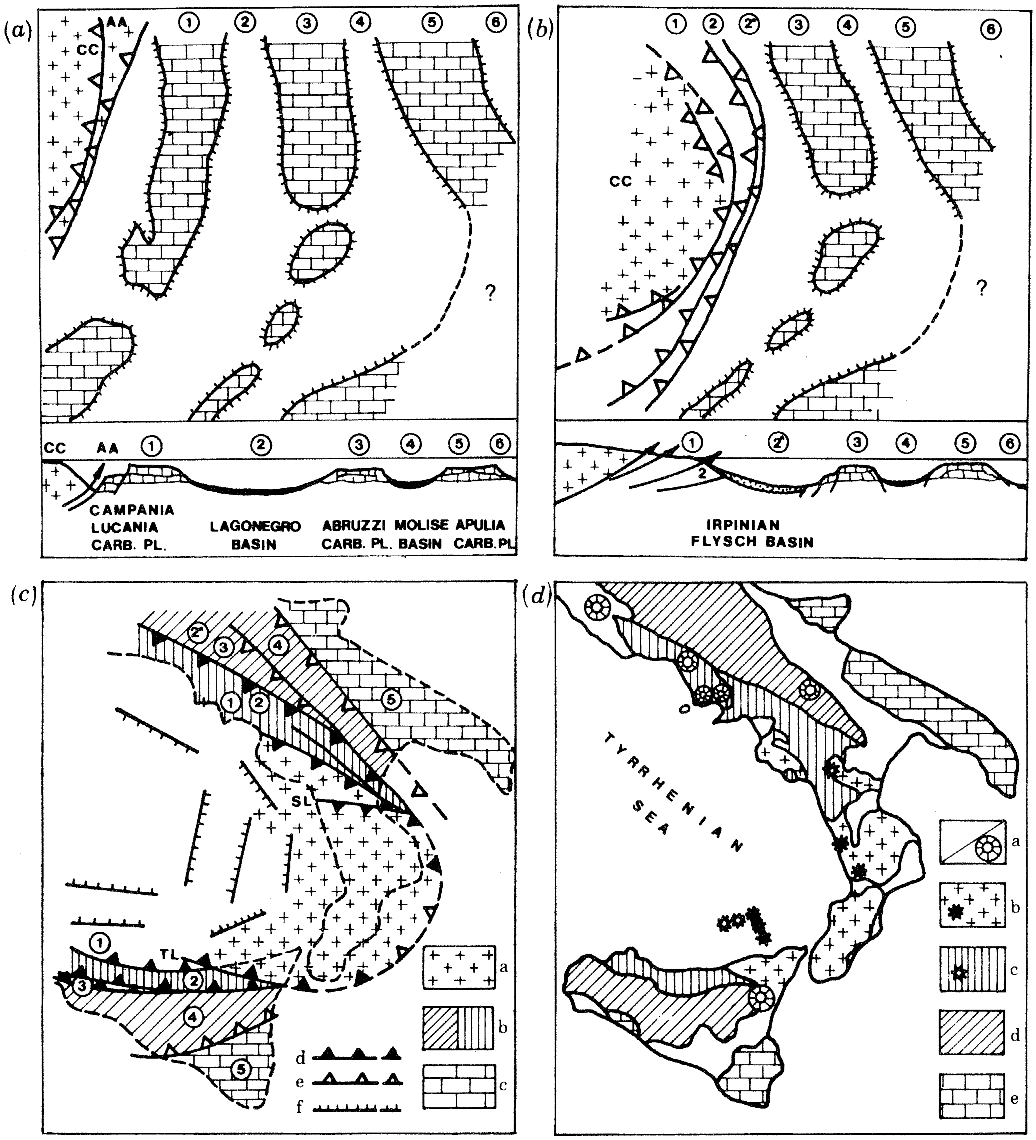

FIGURE 4. Stages of evolution of the Calabrian orocline and tectonic map of southern Italy: (*a*) late Oligocene; (*b*) mid-Miocene; (*c*) late Pliocene; (*d*) Recent (after D'Argenio *et al.* 1980).

(*a*) Palinspastic restoration of southern Italy shortly before Sicilian–Apennine continental margin deformation. CC, Crystelline Calabria nappe system; AA, Longobucco–Longi–Taormina tectonic units (*chaîne calcaire*), back-thrusting against the Sicilian–Apennine palaeogeographic units. Apennine: 1, Campania–Lucania carbonate platform and surrounding or intervening margin-to-basinal areas; 2, Lagonegro Basin; 3, Abruzzi–Campania carbonate platform; 4, Molise–Marsica Basin; 5, Apulia carbonate platform; 6, Gargano basin. Sicily: 1, Panormide carbonate platform; 2, Imerese Basin; 3, Trapanese pelagic carbonate platform; 4, Sicanian Basin; 5, Iblean–Saccense pelagic carbonate platform.

(*b*) Palinspastic restoration of southern Italy after the early deformation of the Sicilian–Apennine continental margin, but before the late Miocene rotations of earlier thrust fronts.

(*c*) The Calabrian orocline after the Lower Pliocene tectonic phase: a, Calabrian Arc tectonic units; b, Apenninic tectonic units originating from the pre-Tortonian (right) and Tortonian – early Pliocene (left) deformation; c, Foreland; d, pre-Tortonian thrust fronts; e, Tortonian – early Pliocene thrust fronts; f, normal faults. SL–TL, Sangineto and Taormina tectonic lines. Encircled numbers refer to palaeogeographic units, here largely deformed. Note the progressive antithetic rotation of thrust fronts, SL and TL.

(*d*) Tectonic sketch of southern Italy: a, foredeep and main marine and continental Quaternary deposits and volcanics; b, Calabrian Arc and associated tectonic units, also showing some of their tectonic windows of c; c, Apenninic and Sicilian units deformed before the Tortonian and tectonic window of d; d, Apenninic and Sicilian tectonic units of Tortonian to early Pliocene age; e, foreland.

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(*b*) Stratigraphic, palaeobiogeographic and tectonic evidence for the continuity of the Periadriatic orogenic belt and palaeomagnetic evidence for the similarity of movement of Adria with Africa since the early Mesozoic (Channell & Horváth 1976; Channell *et al.* 1979).

(*c*) Geophysical evidence for the presence of thick Mesozoic carbonates below the eastern Mediterranean and their correlation with shallow water carbonates on the African foreland on one hand, and with the same rocks in Ragusa and Apulia on the other (Mulder 1973; Burollet *et al.* 1978; Morelli 1978).

(*d*) 'Granitic' rather than oceanic lower crustal velocities (5.0–6.7 km/s); thick and cold lithosphere, and the most recent surface-wave data, which indicate 30–40 km crustal thickness for the eastern Mediterranean (Morelli 1978; Cloetingh 1979).

If this interpretation is correct, then the foundering of the eastern Mediterranean is a neotectonic process, which affected an attenuated and altered continental margin lithosphere (Morelli 1978; Channell *et al.* 1979; Brunn & Burollet 1979). Under these circumstances only limited subduction of such a lithosphere is possible.

It is to be emphasized that the controversial nature of the eastern Mediterranean crust does not affect the thesis that Mediterranean back-arc basins are associated with continent–continent collisions, as only the extent of the southern colliding continent is debatable. The problem is vital, however, if the role of oceanic subduction in back-arc basin formation is being considered.

(iii) The time of back-arc extension and subsidence overlaps the time of post-collisional deformation of the arc (figure 2). It suggests that the extension of the back-arc region is compensated for by compressional deformation in the external zones of the arc. The acute bending of the arcs largely occurs during this compressional deformation. In other words, back-arc extension is related to oroclinal bending. Brunn & Burollet (1979) arrived at the same conclusion by studying the Calabrian, Aegean and Ceram 'inducted arcs'.

The main reason for believing in oroclinal bending is offered by the serious overlap problem encountered if palinspastic sections are restored perpendicular to the present bend of the Mediterranean arcs. Further evidence for progressive bending of the Calabrian Arc (D'Argenio *et al.* 1980) is given by:

(*a*) the opposite vergence of contemporaneous thrust fronts of Calabria over the Apennines (Sangeneto line) and over Sicily (Taormina line);

(*b*) the progressive orientation of the younger thrust fronts to NW–SE in the Apennines and NE–SW in Sicily;

(*c*) the increased shortening toward the Calabrian massif;

(*d*) palaeomagnetic evidence for clockwise rotation of thrust sheets in Sicily and anti-clockwise rotation in the southern Apennines.

The evolution of the Calabrian orocline is shown in figure 4. It is an important characteristic that not the complete lithospheric column, but a decoupled upper sliver (in other words, basement and/or sedimentary cover nappes) moves outward. That is, the outward migration and oroclinal bending of the arc is largely an intralithospheric process.

The importance of intralithospheric splitting in explaining the post-collisional deformations of the Alps has long been recognized (Laubscher 1971; Oxburgh 1972). The mechanism is physically reasonable, because decoupling makes it possible that the heavy subcrustal lithosphere can sink further into the asthenosphere while the more buoyant crustal lid is locked by the collision. (The surface of decoupling is not necessarily the Moho, but rather the rheological boundary between the brittle (upper) and plastic (lower) lithosphere.)

Most recently, Reutter *et al.* (1980) argue that after the late Oligocene collision of Corsica and the northern Apennines, the lower lithosphere of the latter was decoupled and asthenospheric material intruded into the gap. We think that this mechanism can offer a reasonable explanation for the volcanism (see point (vii) below) and crustal attenuation well behind the front of collision (e.g. Tuscany, Latium and Campania).

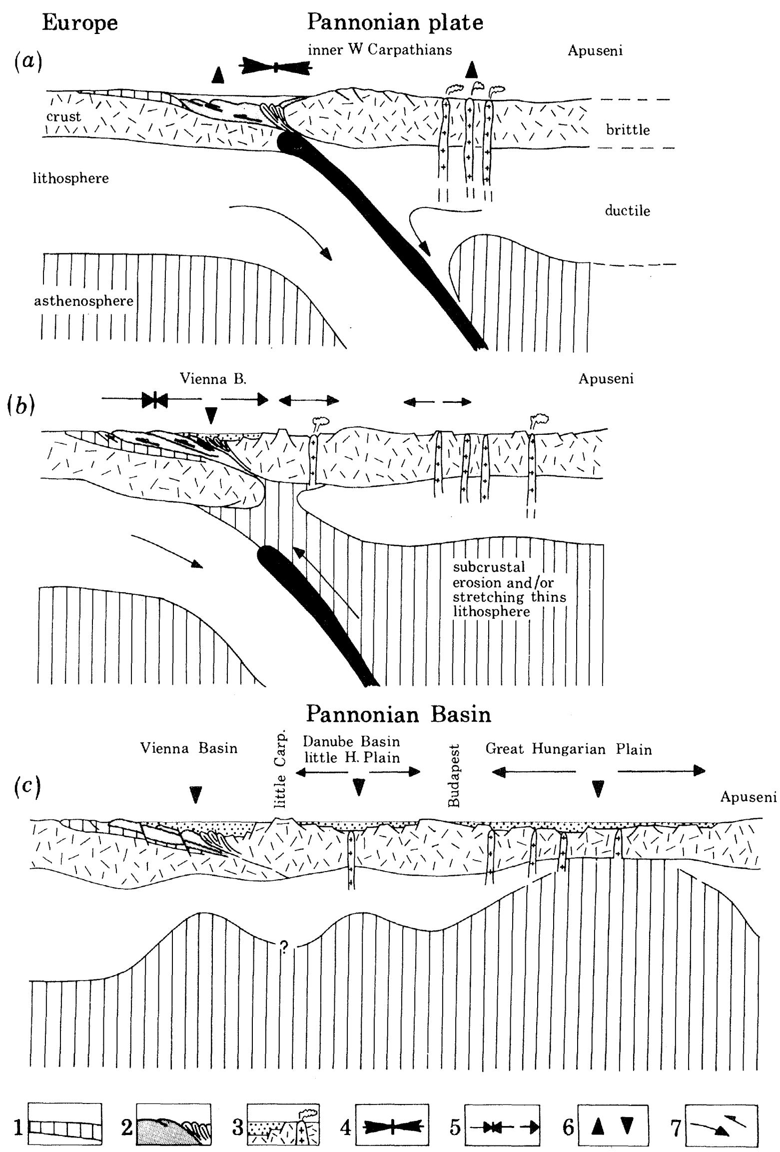

FIGURE 5. Schematic evolutionary model of the westernmost segment of the Carpathian Arc and the Vienna and Pannonian back-arc basins: (*a*) early Miocene; (*b*) early–mid-Miocene; (*c*) late Miocene–Pliocene. See text for further explanation. Key: 1, foredeep molasse; 2, outer Carpathian flysch and subduction zone mélange (Pieniny Klippen belt); 3, Neogene–Quaternary basin fill and calc-alkaline volcanism; 4, continent–continent collision; 5, zone of compression and extension, respectively; 6, uplift and major phase of subsidence, respectively; 7, subduction and overthrusting, respectively.

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The Neogene evolution of the westernmost segment of the Carpathians may give another example of this process (figures 3 and 5). Here, after continent–continent collision, the flysch complex overrode the foredeep molasse towards the end of the early Miocene. This was followed by the collapse of the region of crustal thickening, which resulted in the formation of the Vienna basin. This is a peculiar process because the consequence of crustal thickening ought to be uplift and not subsidence. We speculate that downward bending of the subducted plate and intralithospheric splitting can explain the observed subsidence right after the strong compression. After collision, the leading edge of the European plate bent down and asthenospheric material was able to intrude into the gap. Attenuation and stretching of the thick crust occurred, accompanied and followed by subsidence for isostatic reasons.

Analysis of Landsat images of areas adjacent to the Alboran Basin reveals a regular pattern of lineaments, bounded by two major ENE-trending lines in Spain and in north Africa. The set of lineaments suggests WSW movement of the 'Alboran plate-mosaic' associated with progressive bending of the Gibraltar arc.

Le Pichon & Angelier (1979) have recently worked out a quantitative kinematic model for the late Miocene to Recent deformation of the Aegean region. The major result of this model is that in late Miocene time (13 Ma) the Aegean Arc was less curved and shorter than at present. The progressive bending of the arc is associated with compression in the arc and trench, and extension in the back-arc basin. Moreover, Le Pichon & Angelier suppose that the boundary of the non-rigid Aegean 'plate' is rigid, and go on to calculate its pole of rotation. It suggests that the western branch of the Hellenic trench is a purely consuming boundary, while the eastern branch is mostly transform. They argue that this has been so since the late Miocene (13 Ma B.P.). This rotation of the rigid plate boundary, however, can be considered only a poor approximation of the actual movement of the Hellenic arc. One of the authors, for example, has shown previously that the extensional grabens of the Cretan sea tend to be orientated parallel to the Hellenic arc all along its length, suggesting radial movement of the arc (Angelier 1978). Radial compression in the arc (Angelier 1978; Brunn & Burollet 1979) and in the Mediterranean Ridge (Finetti 1976) is also well documented.

(iv) Back-arc basins are not uniform depressions but a system of basins separated by elevated ranges of tectonic and volcanic origin. This 'basin and range' morphology reflects the intrinsic nature of lithosphere during extension. Lithospheric extension is associated with vertical movements, the history of which may be determined from the sedimentary record. On the basis of numerous seismic sections, onshore and offshore drillings, including the D.S.D.P. holes, the sedimentary sequence of the Mediterranean is fairly well known. Unfortunately, however, no reliable subsidence history can be derived at present because the bathymetric conditions of the Mediterranean during and before the Messinian are strongly debated (see, for example, Hsü *et al.* 1978). This is the major problem one encounters in attempting to formulate quantitative evolutionary models for the Mediterranean back-arc basins. This is, however, not so in the Pannonian basin, where the subsidence history is well understood (Sclater *et al.* 1980; Royden & Sclater, this symposium). There are two major types of subsidence history: one is characteristic for the deep graben-like depressions at the periphery, the other is characteristic for the more isometric depressions at the centre of the intra-Carpathian region. At the periphery there occurred a very fast initial subsidence followed by slow subsidence, as opposed to a single phase of moderately fast and remarkably linear subsidence at the centre. Syn-sedimentary faulting played major and minor roles, respectively, in these basins (Vass 1979). Deep grabens around the central basin are also present in the Tyrrhenian and Aegean basins. Further data may show whether the contrasting subsidence history of the periphery and centre is a general feature of the Mediterranean back-arc basins or is peculiar to the Pannonian Basin.

The major part of the Pannonian Basin was emergent before the basinal development. Massive pre-basinal updoming of several kilometres is well documented in the Aegean region. During the Miocene, Barrovian metamorphism took place in the central Aegean, culminating locally in anatexis and magma generation. These were followed by rapid uplift of the central Aegean, including Attica and the Menderes massif, towards the end of the Miocene (Dürr *et al.* 1978). Arching and faulting is still going on locally. The uplift was associated with the compression of central Aegea and, further on, phases of compression interrupted the general late Miocene to present extensional régime of the back-arc area (figure 2) (Mercier 1977; Angelier 1978). Local compressions during the general extensional period can be observed also in the Pannonian basin (Horváth & Berckhemer 1981).

TABLE 1. SOME GEOPHYSICAL DATA FOR THE MEDITERRANEAN BACK-ARC BASINS

(Mainly after Finetti & Morelli (1973), Morelli (1975), Makris (1977), Panza *et al.* (1978) and Horváth & Berckhemer (1981).)

	average water depth and sedimentary thickness/km	Moho depth/km (crustal velocities/(km/s))	thickness of the lithosphere/km	heat flow mW/m²	intermediate and deep seismicity
Pannonian	0 4†, 2‡	26–30 (5.7–6.9)	56	40–60† 90–110‡	localized (Vrancea), 70–150 km
Aegean	1†, 0.5‡ 1–2†, 0.5‡	20–30 (6.0–6.8)	thin	50–80† 90‡	fairly and poorly defined Benioff zone, 0–180 km
Alboran	1.0 2.0	16 (6.0–6.3)	70–90	60–80	localized (Granada), 600 km
Tyrrhenian	1†, 3‡ 3–8†, 2‡	12‡ (6.9–7.4‡)	30	110‡	discontinuous Benioff zone, 200–300 km

† Peripheral part. ‡ Central part.

(v) The Mediterranean back-arc basins are characterized by crust transitional from typical continental to typical oceanic. As each basins developed from more or less similar continental crust, the thinner their present crust the more mature their present stage of evolution. The thickness of the lithosphere is also less than normal and, as a consequence, high heat flow and high seismic attenuation occur (table 1).

The wide belt of crustal attenuation combined with the diffuse zones of extensional features indicate that back-arc extension is not a rigid-plate opening but an areal expansion of the lithosphere.

The Pannonian Basin is characterized by thin continental crust. An average figure for the thickness of the upper crust (without the sedimentary blanket) is 17 km and for that of the lower crust is 8 km. It indicates that the attenuation of the lower crust is more pronounced than that of the upper crust (Stegena *et al.* 1975). Basaltic dykes are not known with certainty but they may occur below the sedimentary blanket, as indicated by long linear magnetic

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anomalies. Lithospheric thickness is determined by reflexion seismic techniques, and the value obtained (56 km) is constrained by magnetotelluric soundings and deep temperature calculations (Horváth *et al.* 1979). Heat flow is high in the central basin and normal or low in the peripheral depressions (table 1).

The Aegean Basin also exhibits thinned-out continental crust, which is 20 km in the Sea of Crete and increases to 30–32 km in the central and north Aegean (Makris 1977). Magnetic anomalies that are distributed along the Aegean basin are caused by either ophiolites or volcanics. The Saros graben of the northern Aegean Sea is associated with very strong magnetic anomalies. No reliable data for lithospheric thickness are available. Nevertheless, low sub-Moho velocities (7.6–7.8 km/s), low upper mantle density, travel-time residuals and strong uprise of isotherms indicate a shallow asthenosphere (Makris 1977). The few available heat flow data can describe only the broad features. Heat flow highs are associated with the Aegean volcanic arc and with the Turkish coast of the Aegean Sea. A heat flow map of Europe shows that the high heat flow area is the continuation of the Turkish hyperthermal régime (Cermak & Rybach 1979).

Under the Alboran Basin the depth of the Moho is only 16 km, increasing abruptly towards both the Betics and Rif. The mean crustal velocity lies between 6.0 and 6.3 km/s and the sub-Moho velocity is about 7.5 km/s. The velocities and the thickness observed are characteristic of a thinned continental crust overlying an anomalous upper mantle (Working Group 1978). This observation calls into question the validity of the rigid-plate opening model of the Alboran Basin, which is based on the postulate that the portion of the basin deeper than 1000 m consists of new surface created by separation of two rigid plates (Le Pichon *et al.* 1972). Magnetic anomalies of the Alboran Basin are correlated with several volcanic mounts and banks (Galdeano & Rossignol 1977). A strong linear magnetic anomaly is associated with the Alboran rise. The thickness of the lithosphere is less than normal, about 90 km in the western Alboran Basin, rapidly decreasing towards the Balearic Sea (Panza *et al.* 1978). The heat flow has not been well determined, the few measurements indicating values greater than normal (Cermak & Rybach 1979).

The depth of the Moho below the central part of the Tyrrhenian Sea is about 12 km, and may be characterized as oceanic. D.S.D.P. hole 373A penetrated more than 200 m of olivine tholeiite, very similar to that on the Mid-Atlantic Ridge, indicating that the Tyrrhenian abyssal plain has been produced by sea-floor spreading (Barberi *et al.* 1978). The very shallow asthenosphere and exceptionally high heat flow (table 1) suggest that the extension is still active. The crust between the shelf edge and the coast is continental, and the Moho depth there is 20–25 km. In contrast to the relatively thin sedimentary cover of the abyssal plain, the periphery is characterized by deep grabens filled by Miocene–Recent sediments. Strong magnetic anomalies occur that are related to volcanic seamounts. Non-magnetic basement highs are frequently observed. It was shown by dredging (Selli & Fabri 1971; Heezen *et al.* 1971) that they consist of continental fragments formed during the extension of the former continental crust.

(vi) The Alpine–Mediterranean region is a site of high seismic activity. It is remarkable that all intermediate and deep earthquakes are associated with the four back-arc basins discussed here (McKenzie 1972). In the Tyrrhenian and Aegean region the tendency is that deeper earthquakes occur closer to the centre of the basin. It is widely believed that this is a manifestation of active subduction of the eastern Mediterranean lithosphere towards the centre of the basins (McKenzie 1972; Caputo *et al.* 1970; Papazachos 1973). As a matter of fact, however, these Benioff zones are discontinuous and/or poorly defined. Figure 6 shows a section through the Hellenic Arc. It can be seen that earthquake foci are diffuse and most are situated in the crust. This is a region where well developed subduction is certainly not occurring and geophysical data may be explained in terms of collison of two continental plates associated with limited underthrusting of the less buoyant. There is a curious absence of earthquake foci between 50 and 200 km along the postulated subduction zone of the Calabrian arc (McKenzie 1972). The nature of volcanism and the sporadic focal depths suggest that there is a detached lithospheric slab which is the relict of the subduction associated with the convergence of Calabria and the Apenninic margin (Channell *et al.* 1979). The Carpathian Arc is a site of relatively low seismic activity. The only exception is the rather confined Vrancea zone at the junction of the eastern and southern Carpathians. Earthquake foci delineate a nearly vertical slab of 130 km deep with a seismic gap between 40 and 70 km (Fuchs *et al.* 1979). This is one of the few examples of localized intracontinental seismic activity.

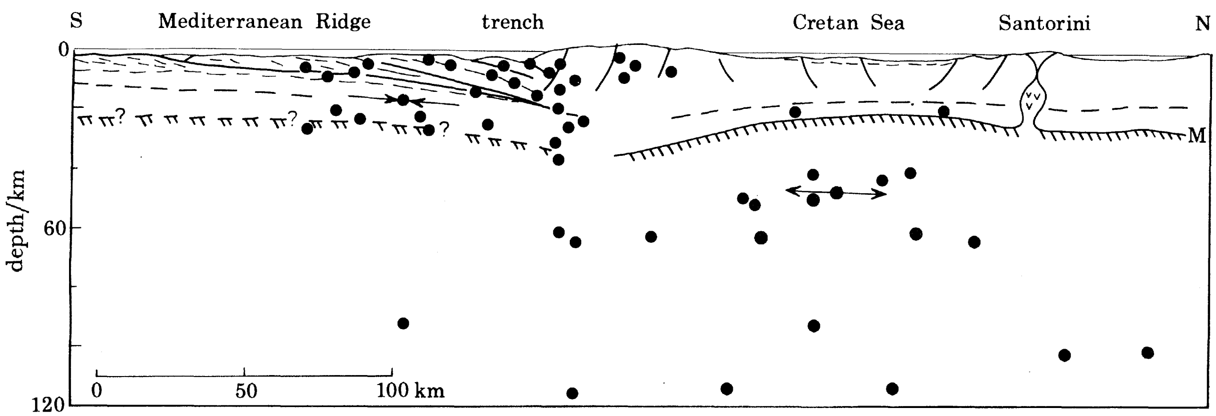

FIGURE 6. North–south section through Crete from central Aegea to the eastern Mediterranean, based on neotectonics, deep seismic soundings (Makris 1977) and seismicity (Richter & Strobach 1978).

The exceptionally deep seismic events of southern Spain are not well understood. They are related to either a relict slab of northward subducted Tethys (Arana & Vegas 1974) or to lithospheric slab underthrust during westward movement of the 'Alboran plate' (Udias & Lopez Arroyo 1972).

(vii) Calc-alkaline volcanism associated with Mediterranean back-arc basins post-dated the continent–continent collision (figure 2) and took place over a wide belt (i.e. not only in front of, but on and behind the suture of the former subduction zone). In other words, Benioff-zone magmatism cannot satisfactorily explain the observed space–time relation of calc-alkaline volcanism.

Lexa & Konecny (1975) demonstrated that the space–time–composition relation of the Neogene calc-alkaline volcanism of the Carpatho–Pannonian region does not show the features that one would expect by analogy with active island arcs. They suggested that the source of calc-alkaline magma was the thermal mantle diapir (updoming asthenosphere) associated with the intra-Carpathian basins. The dominance of rhyolitic ignimbrites within the Pannonian

Basin and rare-earth pattern of andesites (G. Pantó, personal communication) indicate that extensive crustal fusion must have occurred here.

In the Aegean region the late Pliocene–Quaternary volcanic arc is usually related to the subduction of the eastern Mediterranean lithosphere (Ninkovich & Hays 1972; Jacobshagen *et al.* 1978; Le Pichon & Angelier 1979). It was mentioned above that oceanic subduction here is controversial. In Turkey, however, where volcanism of similar age and character is much more abundant, it is clear that the volcanism cannot be connected with any descending

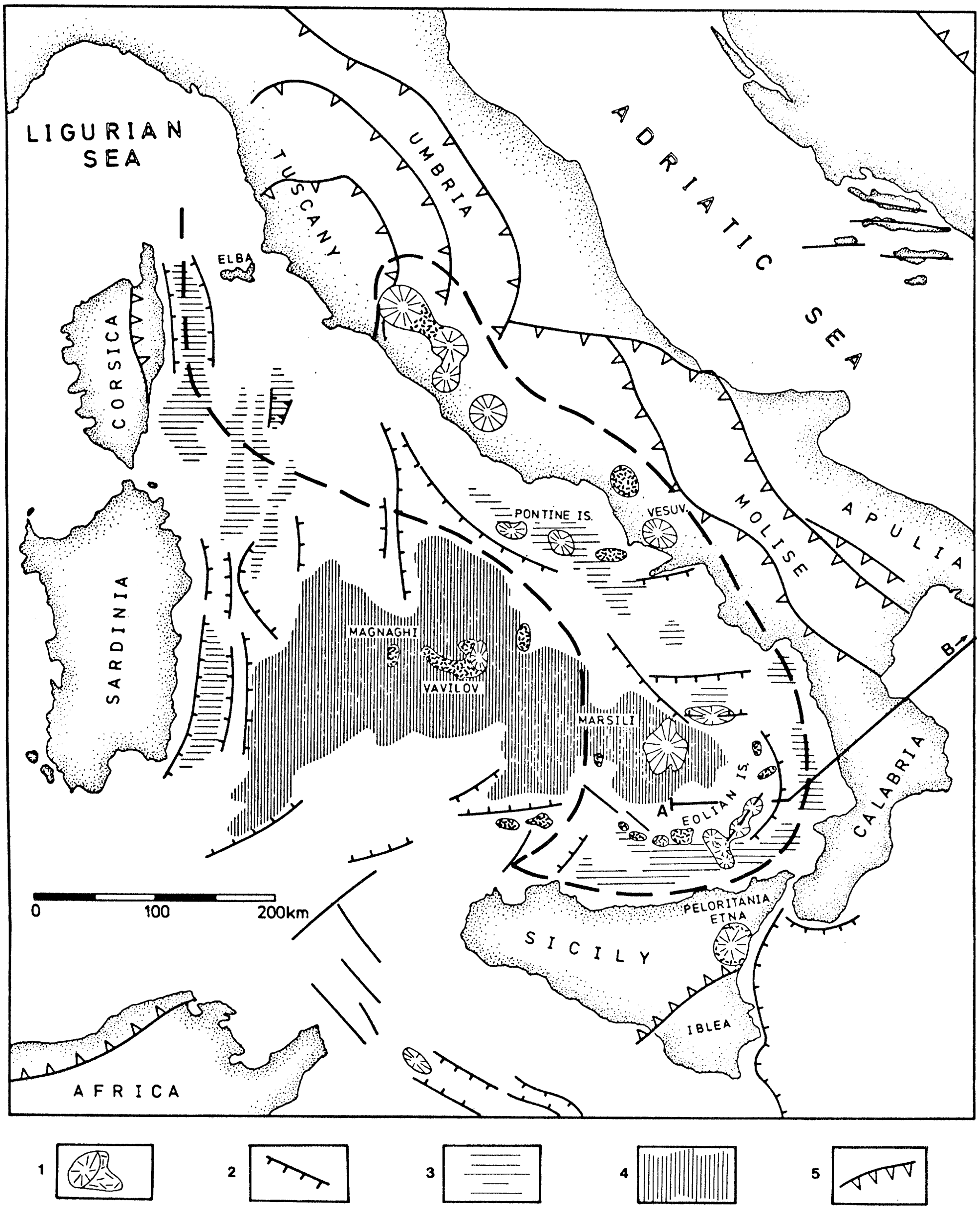

FIGURE 7. Tectonic scheme of the Tyrrhenian basin and surrounding areas (mainly after Boccaletti & Manetti (1978) and Di Girolamo (1978)). The broken line delimits the belt characterized by calc-alkaline and shoshonitic volcanism. Key: 1, Miocene to mostly Quaternary volcanism; 2, major normal faults associated with Miocene–Recent extension; 3, very thick (up to 8 km) Miocene–Recent sedimentary fill; 4, Tyrrhenian abyssal plain; 5, main thrusts.

slab. The observation that volcanics are largely confined to the highest part of the Turkish plateau and that there is a zone of seismic attenuation beneath the plateau suggests that calc-alkaline volcanics may be derived from the melting of the lower crust (Sengör & Kidd 1979). Intensive andesitic volcanism occurred in the central and northeast Aegean and western Anatolia during the early and mid-Miocene (22–13 Ma) (Jacobshagen *et al.* 1978). Its genesis and relation to the formation of the Aegean Basin is poorly understood.

The Alboran Basin, including northern Morocco and southeast Spain, was characterized by calc-alkaline volcanism from the mid-Miocene to early Pliocene, which is related to a northward dipping subduction zone, active during the mid-Miocene (Arana & Vegas 1974). Because the Tethys had been fully consumed here by the end of the Eocene (Biju-Duval *et al.* 1979), it is difficult to understand what ocean was subducted during the mid-Miocene.

A recent review of the volcanism associated with the formation of the Tyrrhenian basin has been published by Di Girolamo (1978). On the basis of major element geochemistry he points out that the convergent plate margin-type volcanites (calc-alkaline and shoshonitic associations) form an extended belt about 650 km long, from northwestern Sicily up to Elba and Tuscany (figure 7). The Aeolian volcanic arc is only small but an integral part of this belt. It is therefore clear that the frequently supposed active subduction along the Calabrian arc cannot offer an explanation for this extended volcanic belt. Moreover, it cannot be connected with the subduction, which preceded the early Miocene collision of Corsica–Sardinia–Calabria and the Apennine margin, as the age of the volcanism is mostly Quaternary. Di Girolamo (1978) concludes that volcanism may be related to the occurrence of magma fusion and magma ascent during the tensional evolutionary phase of the peri-Tyrrhenian region. We add that the deep-seated mechanism could have been the splitting of the leading edge of the Adriatic lithosphere (Reutter *et al.* 1980).

3. Discussion of current geodynamic models

A most interesting review of different ideas on the evolution of the Mediterranean region, up to the mid-1970s, has been given by Hsü (1977). We discuss here only the more recent models and those older ones that are of current interest.

A widely expressed and applied model invokes active mantle diapirism and gravity tectonics (Van Bemmelen 1972, 1973). According to Van Bemmelen, the driving force of orogeny is a mantle diapir, rising from the asthenosphere because of the buoyancy of segregated basaltic melts. The diapir first causes updoming of the crust; spreading of the dome under gravity causes a 'basin and range' type topography, volcanism and radially outward emplacement of nappes. During this process the original continental crust can be transformed into an intermediate type of crust by means of subcrustal erosion and basification. The main mass of the upwelling mantle material is squeezed sideways by the load of the cooling and crystallizing roof of the diapir and the buoyant main body of the diapir. This sideways motion and the gravity spreading cause the arc to override actively the fore-arc lithosphere, giving rise to a 'passive' subduction zone. The back-arc basin is formed by the collapse of the roof of the mantle dome, because of progressive cooling. This model encounters serious difficulties if it is considered the main driving force of the whole process of orogeny (see, for example, Jacobshagen *et al.* 1978). It is, however, quite successful if applied only to back-arc basin formation. As a matter of fact, active mantle diapirism and gravity spreading can well account for most of the observed results, if it is added that basification of continental crust is thought to be not a geochemical process but a magmatic one during the splitting apart of the continental crust. This type of model has been advocated by several authors (Selli 1974; Carey 1976; Makris 1977) and actually, explicit or not, this is the germ of many other more recent models.

An attempt to explain the problem of why and how mantle diapirs are initiated was made by the combination of mantle diapirism and subduction (Scholz *et al.* 1971; Stegena *et al.* 1975; Boccaletti *et al.* 1976; Smith 1978). It was shown, by using the Basin and Range province and the Pannonian and Tyrrhenian basins as examples, that subduction associated with the early evolution of the island or mountain arc could have initiated the updoming of the asthenosphere below the back-arc region. Mathematical modelling of the subduction process has shown that this process is physically tenable (McKenzie 1969; Andrews & Sleep 1974).

Among the different subduction models the recent version from Toksöz & Hsui (1978) seems to be of particular relevance. It is based on a simultaneous solution of the equation of motion and energy. Viscous drag exerted by the subducting slab induces convective flow in the overlying asthenosphere. However, once a lateral temperature gradient is established, thermal buoyancy effects become the more effective driving process, which may continue until counter-balanced by some resistive forces. In their model calculation, 5–10 Ma after the beginning of subduction the upwelling mantle flow generates a surface elevation of the order of 1 km some 200–300 km behind the volcanic arc and produces a tangential stress of 5–10 MPa, which may be sufficient to break and stretch the lithosphere. Although this model is physically correct some reservation is still justified. The principal difficulty is that in the Mediterranean region the actual relation of the time of subduction and back-arc basin extension does not agree with that predicted by the theory. On one hand, there is no evidence in the Aegean region that subduction preceded the beginning of back-arc extension. On the other hand, the formation of the Pannonian and Alboran basins started well after the termination of long-lasting subduction processes. This model also gives no explanation for the particular situation of back-arc basins within the orogen and for the remarkable connection of back-arc basin evolution with continent–continent collision.

In this respect, the rigid–plastic indentation model of continent–continent collision deserves attention (Molnar & Tapponier 1975). It suggests that the westward and southwestward movement of Turkey and Aegea is the consequence of plastic extrusion of continental slivers from the Bitlis collisional suture. But what is the cause of the Aegean extension? Tapponnier (1977) attributes it to a secondary extensional field appearing between the two southward extruded Aegean and Turkish arcs, which are therefore not Pacific-type consuming boundaries. Brunn's (1976) idea of inducted arcs is essentially the same. Dewey & Sengör (1979) argue that the westward motion along the north Anatolian transform is retarded by intracontinental locking between the Sea of Marmara and the northern end of the Hellenic trench, and the resulting east–west compression is relieved by north–south extension. McKenzie (1978*b*), however, questions this interpretation and argues that boundary forces cannot maintain the Aegean extension.

Gravity is a body force that certainly contributes to back-arc extension. The basic principle is very simple: gravitational energy is released if the crust becomes thinner. McKenzie (1978*b*), however, mentions that all continents could release gravitational energy in this way, yet this behaviour is rare. Significant gravity instability occurs, however, at continental margins. The idea is that even in isostatic equilibrium a lithosphere of variable thickness and/or density produces considerable horizontal stresses that try to equalize weight at each level above the depth of compensation (Bott 1971; Artyuskov 1973). Berckhemer (1977) applied this idea of gravity spreading to active continental margins, particularly for the Hellenic Arc. He showed that the difference in elevation between the Aegean sea crust and the eastern Mediterranean floor results in a gravitational stress of 80–90 MPa, acting outward from the Hellenic arc. This is a large force, which may not be counterbalanced by the internal strength of the crust; Aegean crust therefore spreads onto the eastern Mediterranean floor. The overridden lithosphere will be suppressed and may be pulled down by gravity if its density is higher than that of the asthenosphere. The gravity spreading of the back-arc region is accompanied by extension and updoming of the asthenosphere. A third gravitational force that drives back-arc extension may be given by the flowing apart of the lithosphere from the asthenospheric dome. This model is accepted also by Le Pichon & Angelier (1979) for the Aegean Basin.

Recently, two thermomechanical models (lithospheric stretching and subcrustal attenuation) have been suggested and applied to the Mediterranean back-arc basins (McKenzie 1978*a b*; Sclater *et al.* 1980).

Lithospheric stretching is a kinematic description of the process of lithospheric attenuation by listric faulting (brittle lithosphere) and plastic deformation (ductile lithosphere) accompanied by passive upwelling of hot asthenosphere. This change in lithospheric structure is accompanied by an immediate isostatic subsidence. It is followed by a slower subsidence controlled by the thermal decay of the updomed asthenosphere. This type of mechanism was applied by McKenzie (1978*b*) for the Aegean region. He suggested that asthenospheric convection produced by subduction along the Hellenic arc maintains the stretching. Le Pichon & Angelier (1979) argue, however, that because of the changing boundary conditions the convection pattern is variable and it cannot explain the strain pattern of Aegea, which seems to have been relatively stable through time. The principal difficulty of the stretching model is the unreasonably large values of extension required to explain the observed subsidence. McKenzie suggested a twofold stretching for the Aegean region which, in the light of crustal structure and fault patterns, appears to be an overestimate (Le Pichon & Angelier 1979). The average subsidence rate of the Pannonian Basin is so high (150–250 m/Ma) that fourfold or fivefold stretching would be required (Sclater *et al.* 1980). Evidently, pure stretching cannot explain the formation of the Pannonian Basin and another model must be looked for.

Another model is subcrustal attenuation. Here, attenuation of the subcrustal lithosphere occurs without significant extension of the crust. Because of asthenospheric updoming, this phase is accompanied by uplift. If sufficient thickness of the crust is involved in the attenuation process, initial isostatic subsidence rather than uplift may occur. These processes are then followed by long-term thermal subsidence due to thermal relaxation of the updomed asthenosphere. Subcrustal attenuation can be the result of either subcrustal extension or subcrustal melting due to a change in asthenospheric thermal conditions. It is an important feature of this model, which is very much like active mantle diapirism, that it can describe non-proportional thinning of the crust and subcrustal lithosphere and can therefore account for lithospheric uplift. Crough & Thompson (1976) have shown that the average topography of elevated physiographic provinces of North America can be explained in terms of lithospheric attenuation. In the Basin and Range province the asthenosphere either directly underlies the 25–35 km thick crust (Crough & Thompson 1976; Keller *et al.* 1979) or a thin (15–30 km) mantle lid separates them (Thompson & Zoback 1979). As 10–35% average extension is inferred for the region (Stewart 1978; Lachenbruch & Sass 1978), the very shallow asthenosphere apparently cannot be the result of pure stretching. Other highlands of the world associated with orogenic belts (the Turkish—Iranian plateau, Tibet, Altiplano) are also characterized by thinned-out lithosphere (Toksöz & Bird 1977; Sengör & Kidd 1979). The uplift of central Aegea towards the end of the Miocene can also be understood in terms of lithospheric attenuation (Makris 1977). The Pannonian region was also emergent before mid-Miocene times. The mid-Miocene was characterized by the formation of several horsts and grabens, and by eruptions of rhyolitic ignimbrites. In the Pannonian Basin, where fast linear subsistence has occurred since the late Miocene, all available data suggests that the model of subcrustal attenuation is applicable (Sclater *et al.* 1980).

4. Conclusion

We suggest the following evolutionary scheme for the back-arc basins of the Alpine–Mediterranean region discussed in this paper.

1. Deformation of the internal zones of an orogen culminates in continent–continent collision. Plate consumption can proceed significantly further by tectonic shortening of the external zones. The shortening is accompanied by thickening of the crust, which leads to strong crustal contrast to the foredeep and foreland characterized by thin to normal crust. Attenuation of the lithosphere leads to updoming of a large part of both the external and internal zones; it is accompanied by widespread faulting and volcanism. Lithospheric attenuation is initiated by an asthenospheric thermal event which may somehow be related to the former subduction process. This is the highland phase of back-arc basin evolution. Actual examples are the Turkish–Iranian plateau, the west Anatolian graben system and the Basin and Range Province of western U.S.A.

2. Gravity instability gives rise to radial extension of the uplifted and already faulted region. Detached slivers of the external zone override the foredeep lithosphere, forming the active arc and the conjugate zone of underthrusting. The extensional process involves the whole lithosphere and can be described as lithospheric stretching. Back-arc subsidence is controlled by the immediate isostatic response of the lithosphere to structural changes (attenuation, stretching) and by conductive decay of the heat anomaly (asthenospheric dome). The end-products of this evolutionary phase are the continental and subcontinental back-arc basins like the Pannonian and Aegean basins.

3. If further thrusting of the arc over the foreland is still possible, for example because the foredeep exhibits unusually thin crust, a back-arc basin strongly intruded with basaltic dikes and locally characterized by true oceanic crust may be formed. In this case the arc is strongly curved and the underthrust foreland lithosphere may exhibit features of a poorly developed Benioff zone. This is the suboceanic or oceanic phase of back-arc basin evolution, exemplified by the Tyrrhenian Basin.

As the principal force of back-arc basin formation is the gravity instability at the arc–foredeep boundary and that of the lithosphere overlying an asthenospheric dome, back-arc basin formation may be considered as a gravity runaway of a particular segment of an orogenic belt.

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Discussion

C. Coulon (*Laboratoire de Petrologie, Faculté des Sciences, Marseille, France*). Does Dr Horváth also interpret the North Balearic Basin, which is related to the Cainozoic drifting of Corsica and Sardinia, as a marginal basin?

F. Horváth. I do not think that the north Balearic–Ligurian basin is the same kind of marginal basin as those discussed. A major difference is that this basin opened up at the edge of stable Europe, i.e. it cannot be considered as a post-collisional episutural basin. Moreover, the calc-alkaline volcanism can well be related to the subduction of oceanic lithosphere. In other words, the north Balearic–Ligurian basin appears to be much more similar to the marginal seas of the western Pacific than the other four Mediterranean basins.

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