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Plate convergence, crustal delamination, extrusion tectonics and minimization of shortening work as main controlling factors of the recent Mediterranean deformation pattern

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Abstract
It is argued that the time-space distribution of major post middle Miocene deformation events in the Central-Eastern Mediterranean region, deduced from the relevant literature, can be coherently explained as a consequence of the convergence between the Africa/Arabia and Eurasia blocks. This plate convergence has mainly been accommodated by the consumption of the thinnest parts of the Northern African (Ionian and Levantine basins) and peri-Adriatic margins. During each evolutionary phase the space distribution of trench zones is controlled by the basic physical requirement of minimizing the work of horizontal forces, induced by plate convergence, against the resisting forces, *i.e.*, the cohesion of the upper brittle crustal layer and the buoyancy forces at the consuming boundaries. The significant changes of tectonic styles which determined the transition from one phase to the next, like those which occurred around the Messinian and the late Pliocene-early Pleistocene, were determined by the suture of consuming boundaries. When such an event occurs, the system must activate alternative consuming processes to accommodate the convergence of the major confining blocks. The observed deformations in the study area suggest that this tectonic reorganization mostly developed by the lateral extrusion of crustal wedges away from the sutured borders. This mechanism allowed the translation of maximum horizontal stresses from the locked collisional fronts to the zones where consumable lithosphere was still present, in order to activate the next consuming processes. The extensional episodes which led to the formation of basins and troughs in the Tyrrhenian and Aegean zones are interpreted as secondary effects of the outward escape of crustal wedges, like those which occurred in response to longitudinal compressional regimes in the Apennines and Aegean regions.

Key words *Mediterranean – tectonics – driving mechanisms*

1. Introduction

The deformation pattern in the Mediterranean region has been rather complex, as usually occurs in continental collision zones. However, in this case there have occurred some peculiar features, such as crustal stretching in some intra-orogenic basins (*e.g.*, the Balearic, Tyrrhenian and Aegean regions) lying along the collisional boundary between the Africa and Eurasia continents. This evidence led some authors to suppose that the evolution of extensional zones was mainly controlled by local independent driving mechanisms, as, for example, the uprising of mantle material or the gravitational sinking of the subducted lithosphere (see Lavecchia *et al.*, 1988; Serri *et al.*, 1991; Van Dijk and Okkes, 1991 for extended

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reviews of the proposed models in the Central Mediterranean). In this work, instead, we try to demonstrate that the deformation pattern in the Central-Eastern Mediterranean region (fig. 1), including the extensional episodes, can be coherently explained as a consequence of the convergence between the confining blocks, *i.e.*, Africa, Arabia and Eurasia, as already proposed by a number of authors (see, *e.g.*, Caire, 1973; Brunn, 1976; Tapponier, 1977; Boccaletti *et al.*, 1982; Mantovani *et al.*, 1992, 1993a, 1994, 1996).

In order to facilitate understanding and the check of the proposed reconstruction, a detailed description is given of the evidence considered, the boundary conditions adopted and the main tectonophysical concepts underlying our approach.

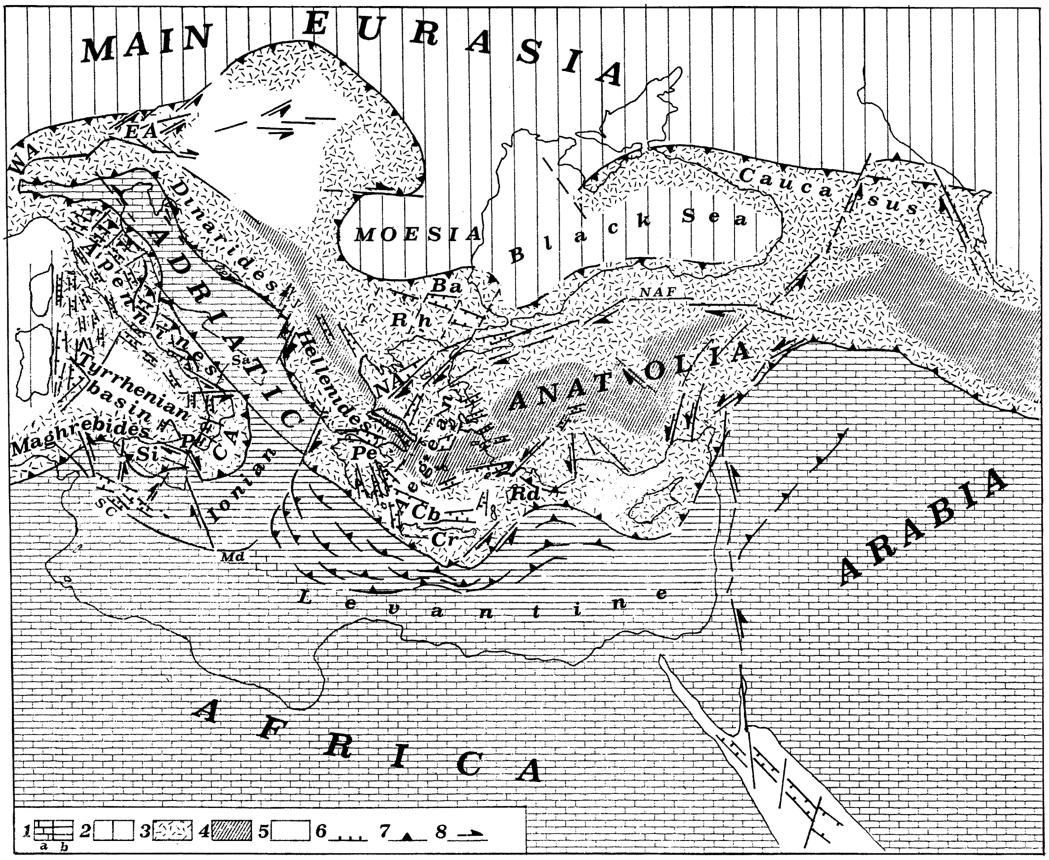


Fig. 1. Main structural-tectonic elements in the Central-Eastern Mediterranean area. 1) African/Adriatic domain with continental (a) and thinned crust (b); 2) Eurasia domain; 3) deformation belts; 4) internal massifs of the Tethyan belts; 5) main Neogene-Quaternary basins and troughs; 6, 7, 8) tensional, compressional and transcurrent features. Ba = Balkan belt; CA = Calabrian Arc; Cb = Cretan basin; Cr = Crete; EA = Eastern Alps; Md = Medina Seamounts; NA = Northern Aegean trough; NAF = North Anatolian fault system; P = Peloritani block; Pe = Peloponnesus; Rd = Rhodes; Rh = Rhodope massif; Sa = Salento peninsula; Si = Sicily; SC = Sicily Channel; WA = Western Alps. Other geographical and tectonic features cited in the text are reported in the maps of figs. 5a-d.

2. Evidence

A huge amount of information is available on the Mediterranean region from all branches of Earth Sciences. We tried to turn this framework of data into a limited number of major deformation events to use as constraints on the evolutionary model. An event is mainly defined by the kind of deformation observed (for instance compressional, tensional or transcurrent) in a given zone and by the period of time during which this deformation occurred. The recognition of each event is generally based on the analysis of many features possibly related to independent observations. The description of the evidence considered and the respective references are reported in a separate paper (Babbucci *et al.*, 1997) in this volume.

3. Boundary conditions: Africa and Arabia kinematics

3.1. *Africa*

The recent/present motion trend of this plate with respect to Eurasia is generally assumed as coherent with a pole located offshore Morocco (see, *e.g.*, Dewey *et al.*, 1989; DeMets *et al.*, 1990 and references therein) which implies a convergence trend ranging from NW to N in the Central-Eastern Mediterranean. These models are constrained by magnetic lineations and other kinematic indicators in the North Atlantic and are supported by the geological and seismic shortening pattern in the Ibero-Maghrebian region, which indicates a NW to NNWward plate convergence (see, *e.g.*, King and Vita-Finzi, 1981; Philip and Meghraoui, 1983; Buforn *et al.*, 1988; Rebai *et al.*, 1992). This approach is based on a plate mosaic which assumes Eurasia as a unique rigid plate from the North Atlantic ridges to the Pacific trenches.

However, this last assumption cannot easily account for the presence of tectonic and seismic activity in Western Europe, from the Rhine graben system to the Iberian zone and in the adjacent North Atlantic area (Ahorner, 1975; Biju-Duval *et al.*, 1977; Bousquet and Philip, 1986; Grimison and Chen, 1988; Buforn *et al.*, 1988). Furthermore, in our opinion, a NNW to NWward motion of Africa can hardly be reconciled with several major features of the post Tortonian deformation pattern in the Central Mediterranean area, which led us to believe that a NE to NNEward motion of Africa with respect to Eurasia during the last 10 My is more plausible (Mantovani *et al.*, 1992, 1996).

To demonstrate that this last kinematic hypothesis is not in contrast with any available evidence, Mantovani *et al.* (1992) and Albarello *et al.* (1993, 1995) have shown that it can be reconciled with all North Atlantic data, within their respective errors, and with the Maghrebian collisional pattern, if Western Europe is allowed a (even small) relative motion with respect to Main Eurasia. The resulting kinematic model is shown in fig. 2.

It must be pointed out that these computations are not sufficient to favour the Africa-Eurasia kinematic model here adopted, but they at least point out that if Eurasia was not a unique rigid plate, a NNE to NEward convergence between Africa and Main Eurasia would be an acceptable kinematic model. Thus, a possible discrimination between the different motion trends so far hypothesized for the Africa-Eurasia convergence depends on the resolving power of the evidence on the internal rigidity of the Eurasian system. In this regard, one must consider that the kinematic solution illustrated in fig. 2 implies a relative motion of a few millimetres/year between Western Europe and Main Eurasia. Since it seems hard to demonstrate that such a small rate could not be accommodated by the tectonic activity observed in the several active regions of Western Europe, we believe that, in the light of the North Atlantic and Maghrebian evidence both kinematic models discussed above must be considered possible solutions.

Our choice of Africa-Eurasia kinematics (fig. 2) has been determined by the analysis of the Mediterranean deformation pattern and by some other considerations reported below. Previous reconstructions of the Africa-Eurasia relative motion over time, based on North Atlantic magnetic lineations (see, *e.g.*, Dewey *et al.*, 1989) suggest that before 9 My the con-

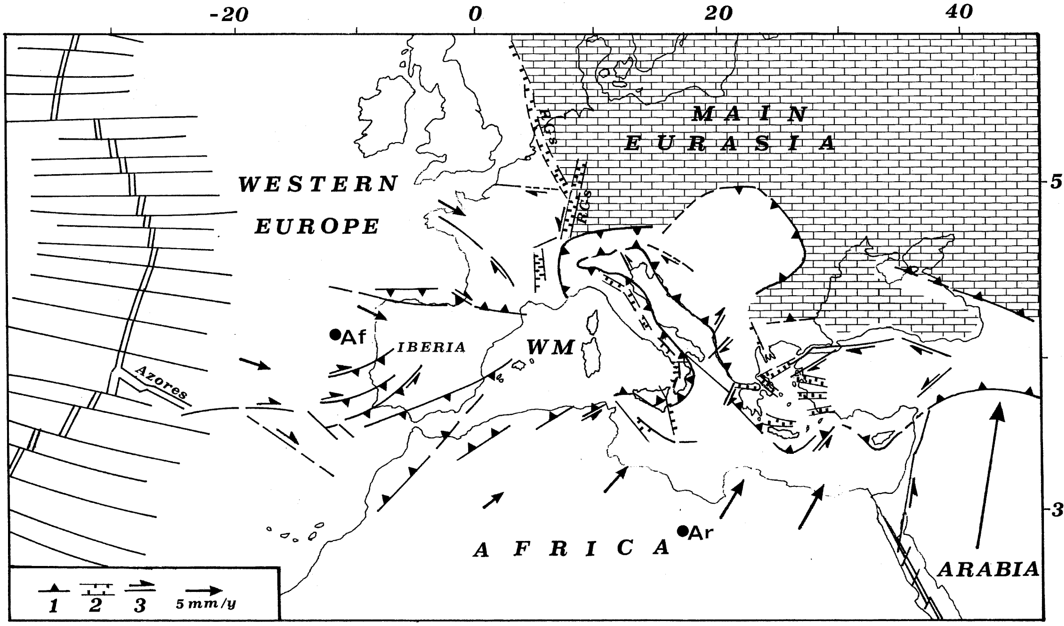


Fig. 2. Relative motions of Africa, Western Europe and Arabia with respect to Main Eurasia (bricked area), as proposed by Mantovani *et al.* (1992) and Albarello *et al.* (1993, 1995). The Western Europe block is bounded southward by the Azores-Gibraltar discontinuity and eastward by the Rhine Graben system (RGs) and the Western Mediterranean basin (WM). Arrows in Africa, Western Europe and Arabia indicate the proposed velocity fields with respect to Eurasia, in accord with the rotation poles respectively located offshore Portugal at 41.6°N, 11.8°W (Af) with an angular velocity of 0.12°/myr, at 49.9°N, 129.6°E outside fig. 2, with an angular velocity of 0.04°/myr and in North Africa, at 28.4°N, 17.0°E (Ar) with an angular velocity of 0.51°/myr. 1, 2, 3) Main compressional, tensional and transcurrent features.

vergence between Africa and Eurasia was directed roughly NEward and then it turned roughly NWward. However, such a considerable change of drifting trend (roughly 90°) of a large continental block, like Africa, would imply a radical reorganization of shortening processes, transform fault systems and extensional phenomena in all surrounding regions, which is not supported by any clear evidence. The solution shown in fig. 2, instead, implies a much simpler time pattern of the Africa-Eurasia kinematics, since it suggests that at 9 My no important change has occurred in the convergence trend between these two blocks.

Geological and paleomagnetic evidence in Central Europe and the formation of an important tectonic feature, like the Rhine graben system, which crosses Central Europe from the North Sea to the Mediterranean basin, led some authors (Le Pichon *et al.*, 1988; Bergerat, 1987) to suggest that an independent motion between Western Europe and Main Eurasia occurred in the past. Thus, the idea that some small relative motion, or at least some reactivation of such an important decoupling, may still go on between these two parts of the original Eurasia plate can hardly be ruled out.

A more definitive answer to this problem will most probably be given by geodetic space measurements (VLBI, SLR and GPS) in the near future. At present, in our opinion, no significant information can be derived from this kind of data about the Africa-Eurasia kinematics (see Mantovani *et al.*, 1995).

3.2. *Arabia*

Since the available evidence in the Red Sea/Gulf of Aden rift zones and in the Dead Sea fault system indicates that no activity has occurred from roughly the middle Miocene to the early Pliocene (see, *e.g.*, Girdler, 1985; Hempton, 1987 and references), we have assumed that the Arabian promontory moved as part of Africa until the early Pliocene. The reactivation of extensional activity in the Red Sea/Gulf of Aden zones and of transcurrent motion along the Dead Sea fault system in the lower Pliocene (and the successive time patterns of these deformations) indicate that during the last 4-5 My Arabia moved faster than Africa with respect to Eurasia (Girdler, 1985; Hempton, 1987).

The estimates of the recent/present migrating trend and rate so far carried out for Northern Arabia (Chase, 1978; Minster and Jordan, 1978; DeMets *et al.*, 1990; Taymaz *et al.*, 1991; Lyberis *et al.*, 1992; Oral *et al.*, 1993; Westaway, 1993) suggest that the most probable motion rate is comprised between 2.0 and 2.5 cm/yr and that the motion trend is roughly northward. We have assumed a rotation pole located at 28.4°N, 17.0°E and an angular velocity of 0.51°/My for the Plio-Quaternary motion of Arabia (fig. 2), which is compatible with the amount of shortening estimated along the North-Eastern Arabian border (see, *e.g.*, Dewey and Burke, 1973; Dercourt *et al.*, 1986; Zonenshain and Le Pichon, 1986; Klitgord and Shouten, 1986; Chaimov *et al.*, 1990; Lyberis *et al.*, 1992; Jackson, 1993; Chorowicz *et al.*, 1994; Hempton, 1987) and with the kinematic indicators in the Red Sea-Gulf of Aden (DeMets *et al.*, 1990). This rotation vector implies Arabia-Eurasia motion rates in the range mentioned above.

4. Main concepts on the mechanics of continental collision

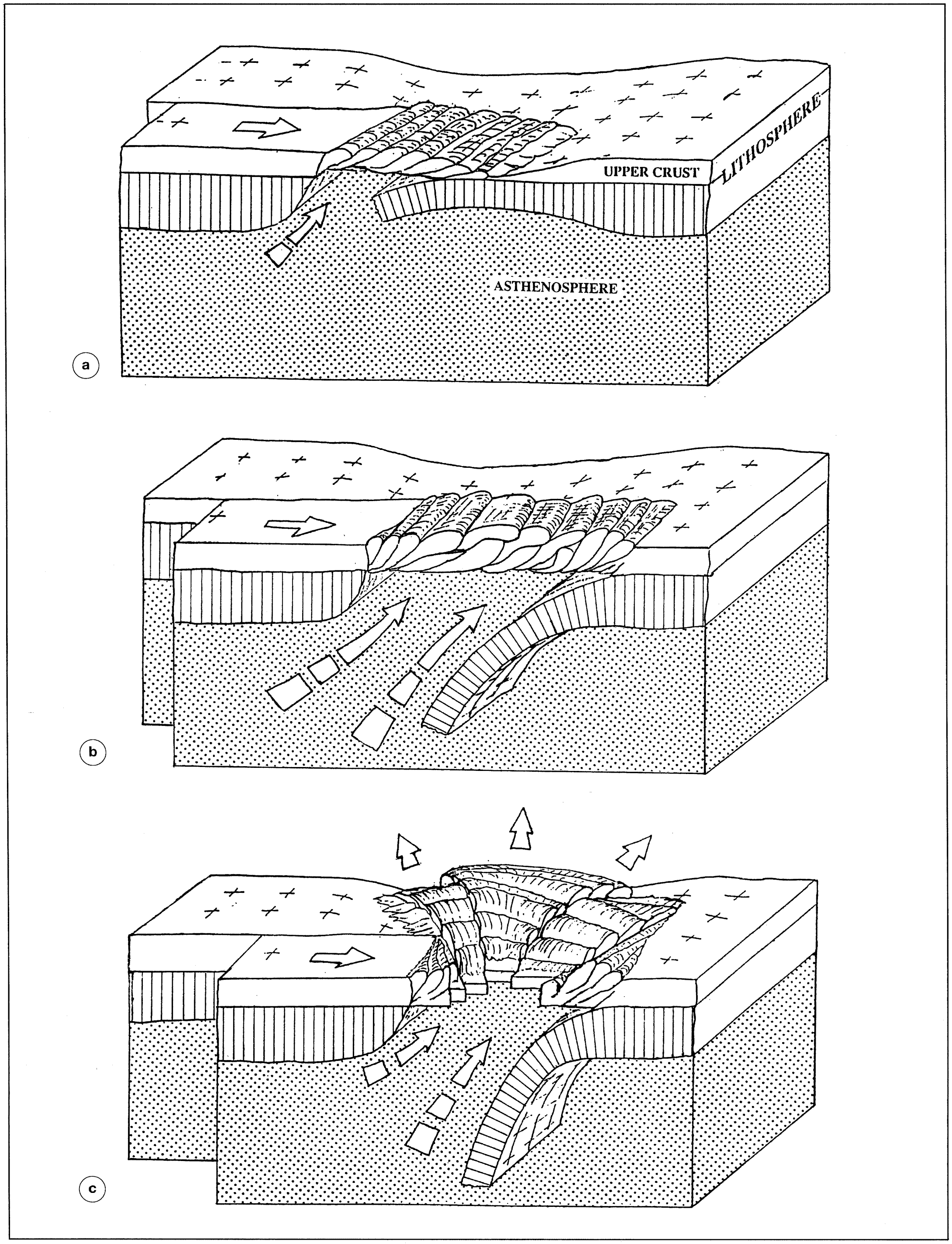
Tapponier and Molnar (1976) have demonstrated that the slip-line field theory, based on plastic rheology, can account for several features of the India-Eurasia collision. This model, which is also supported by plasticine experiments (Tapponier *et al.*, 1982; Peltzer and Tapponier, 1988) and by the comparison with estimated strike-slip and convergence rates (Avouac and Tapponier, 1993), implies that a limited number of faults with large offsets and an unconstrained margin can accommodate the lateral motion of crustal wedges in response to convergence and indentation. In this model, plate convergence is completely compensated by the lateral escape of crustal wedges towards weak lateral constraints.

Other authors (*e.g.*, England and McKenzie, 1982; Vilotte *et al.*, 1986; Cohen and Morgan, 1987; Houseman and England, 1986, 1993) emphasized, by numerical calculations, the role of gravity on the displacement field of a continuum viscous medium. In this kind of model, crustal thickening in front of an indenter is relaxed by lateral flow towards a thinner crust.

Both numerical and experimental investigations suggest that lithosphere rheology, gravity and the boundary conditions are the factors controlling continental collision processes. In particular, some authors (see, *e.g.*, Ratschbacher *et al.*, 1991; Faccenna *et al.*, 1996) demonstrated by experimental modeling that crustal thickening, lateral escape and gravitational spreading may all contribute to the overall deformation, depending on structural and dynamic conditions, such as the degree of lateral confinement, the size of the deformable area, the indenter shape and strength, the foreland strength and the velocity of indentation.

On the basis of these results and the analysis of the deformation pattern in the study area we developed some basic ideas on the main controlling factors of Mediterranean evolution, which are synthetically described below:

a) The rheological behaviour of the blocks which interact in the Mediterranean region can be represented, in general, by a vertical profile which encompasses a brittle upper crust, a ductile lower crust, a hard mantle layer and a low viscosity asthenosphere (see, *e.g.*, Lobkovski and Kerchmann, 1991; Ranalli, 1995). A tentative estimate of rheological profiles in a number of structural provinces of the area here considered is reported by Viti *et al.* (1997). The results of this study point out some basic


UPPER CRUST
LITHOSPHERE
ASTHENOSPHERE
a
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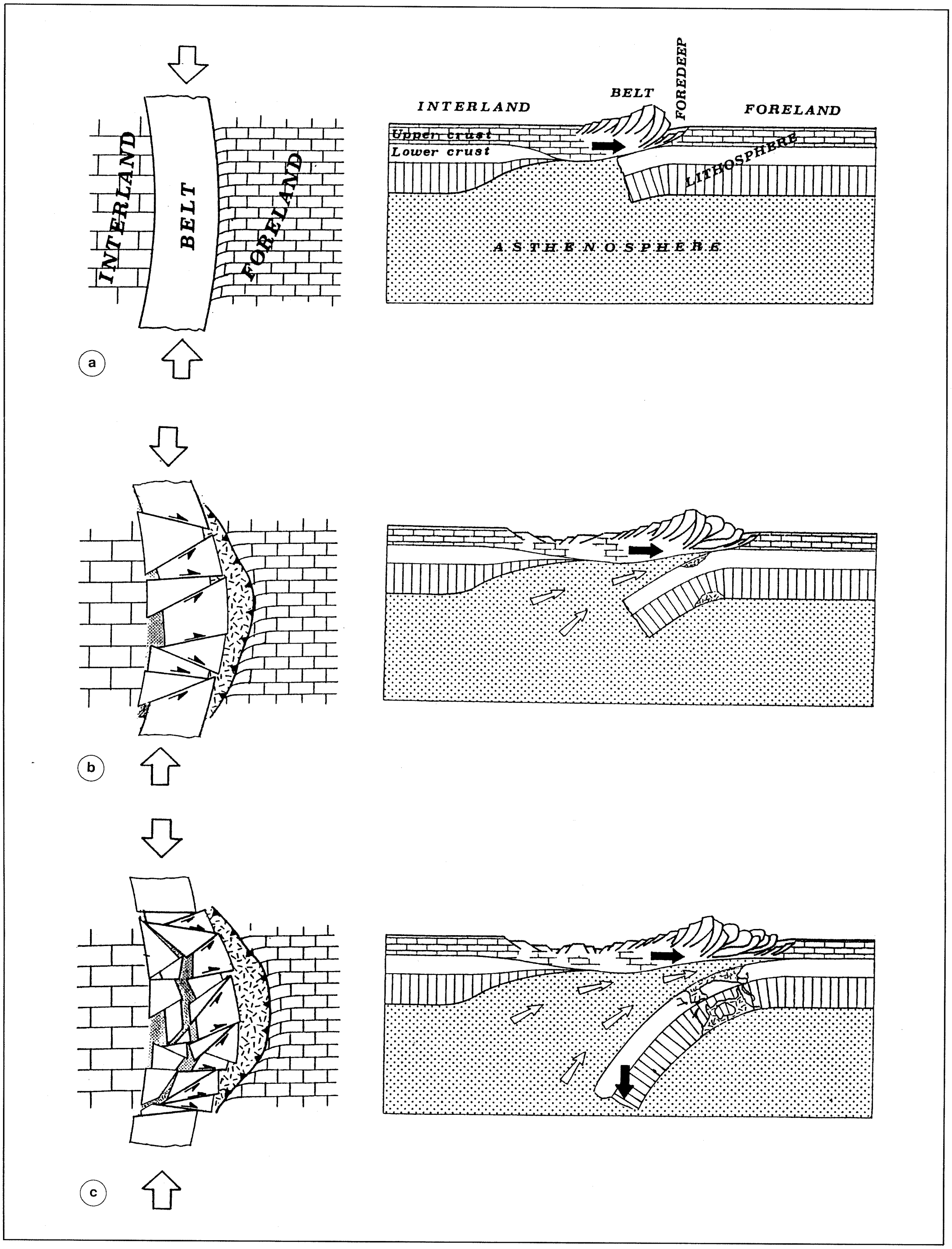
features which can give important insights on the geodynamic behaviour of the various Mediterranean regions: a) in the zones affected by extensional tectonics, as the Tyrrhenian, the Sicily channel and the North Aegean, the mechanical strength is confined to the upper brittle crust; b) in most of the other zones (continental fragments and orogenic belts) high mechanical strength characterizes the upper brittle crust and the uppermost part of the mantle, but these strong layers are separated by a ductile lower crust. The only zones characterized by a thick lithospheric layer with no ductile zones inside are the «cold» Mesozoic oceanic areas of the Ionian and Levantine basins. One could expect that in these last zones delamination of the crust cannot occur. Thus, when this kind of structures are involved in subduction processes most of the lithospheric layer sinks into the underlying mantle, only the sedimentary cover is decoupled and accumulated in the trench zones. This might explain why the consumption of large sectors of the Ionian and Levantine lithosphere beneath the Calabrian and Hellenic-Cyprus arcs has only produced very limited accretionary belts, like the external Calabrian arc and the so called «Mediterranean Ridge» (Finetti, 1976).

b) The collision between converging blocks characterized by a normal continental or thinner crust is accommodated by a delamination mechanism (fig. 3a-c), which develops through two main phases: first, the horizontal compression causes the shortening of the upper buoyant brittle part of the crust, by faulting and imbrication of the resulting slivers (see, *e.g.*, Burchfiel, 1980; Bally, 1981; Meissner, 1986; Gillet *et al.*, 1986), then the underlying lithosphere (lower crust and strong mantle) sinks into the asthenosphere driven by its negative density contrast. In the wake of the sinking slab, asthenospheric material flows laterally and upwards up to the bottom of the delaminated crustal units. This contact of the upper crust with asthenospheric material could explain the fact that most orogenic belts are characterized by higher than normal heat flow (Morgan and Sass, 1984). In fig. 3a-c the detachment level is arbitrarily assumed at the base of the upper crust. In reality, this level could be located anywhere in the lower crust, depending on the thickness and rheological properties of this layer (see, *e.g.*, Kirby, 1983; Ranalli and Murphy, 1987; Van den Beukel, 1992; Burov and Diament, 1995). The occurrence of a delamination-like process has been also suggested, under the name «ensialic subduction», to explain the deformation pattern in the Apennines (Castellarin *et al.*, 1978; Boccaletti *et al.*, 1980).

c) The shortening of the upper crust, involving faulting and spatial reorganization (with uplift) of the resulting fragments, is contrasted by the mechanical strength of the crust and by gravity and thus its occurrence requires a given work. The basic physical requirement of minimizing the sum of all contemporaneous «shortening works» in the whole Mediterranean collision zone is a major factor controlling the occurrence and the distribution of shortening processes during each evolutionary phase. This implies that the system, in re-

Fig. 3a-c. Tentative sketch of the tectonic processes which are believed to have accommodated continental plate convergence in the Mediterranean area. a) Plate convergence causes shortening (by imbrication of crustal slivers) of the upper brittle buoyant crust, determining its detachment from the underlying lithosphere. b) Once decoupled from the shallow buoyant layer, the lower lithosphere sinks into the mantle, due to its negative density contrast. In the wake of the sinking slab, asthenospheric material flows to reach the bottom of the delaminated upper crust. c) In response to significant increases of the resisting forces in the consuming boundary (possibly connected with thickening or rheological changes in the crust entering subduction) the tectonic pattern may be considerably modified. The consumption of the lithosphere comes to an end at that boundary, and lateral escape of crustal wedges develops, at the expense of a weak constraint (constituted, for example, by a thinner crust in some other zone of the system). The lateral mobility of the orogenic wedges is favoured by their weak coupling with the underlying, mostly asthenospheric, material. See text for further comments.


INTERLAND
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INTERLAND
FORELAND
Upper crust
Lower crust
LITHOSPHERE
ASTHENOSPHERE
a
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sponse to significant increases of the resisting forces at one or more active zones, can find it more «convenient» to interrupt a given shortening pattern and activate new processes at other «weaker» boundaries.

d) The shift of maximum compressional stresses from the suturing boundary to less constrained zones is generally achieved by the lateral escape of crustal wedges (fig. 3c). Depending on the degree of confinement and the mechanical properties of the collision zone, the rate of indentation and the presence of pre-existing discontinuities, the extrusion process may be preceded by a more or less pronounced crustal thickening and consequent uplift. Gravitational collapse from the uplifted zones may then contribute to the lateral escape of crustal wedges (see, *e.g.*, Ratschbacher *et al.*, 1991).

e) Extrusion processes generally involve orogenic belts, both in the Mediterranean (see, *e.g.*, Mantovani *et al.*, 1996) and the India-Eurasia collision zones (see, *e.g.*, Tapponier and Molnar, 1976; Tapponier *et al.*, 1986). This evidence could be explained by the fact that orogenic belts, due to their genetic mechanism (fig. 3a-c), are presumably characterized by a low internal coherence, high buoyancy and scarce connection with the underlying, mostly asthenospheric, mantle. These features facilitate those processes, such as internal displacements of masses and uplift, which allow thickening of the collision zone and also imply a little work in the lateral displacement of upper crustal wedges, due to the presence of a shear stress free boundary at their bottom (see fig. 3a-c).

f) The Mediterranean system is affected by a number of major discontinuities, where most deformation is accommodated. The most evident and large scale example is the North Anatolian fault system, but several other significant discontinuities at regional and local scale are pointed out by geological and seismotectonic studies in the Apenninic belt, Calabrian arc, Sicily, Peloponnesus, Crete-Rhodes and Anatolia (see, *e.g.*, Finetti and Del Ben, 1986; Reuther *et al.*, 1993; Ben-Avraham and Grasso, 1990). The importance of large faults in controlling tectonic processes has also been stressed by other authors (Tapponier and Molnar, 1976; Tapponier, 1977; Scotti and Nur, 1990; Jackson, 1993; Twiss *et al.*, 1993; Avouac and Tapponier, 1993).

g) Tensional/transtensional deformation may develop in the wake of extruding wedges or in the internal sides of bowing arcs (see fig. 4a-c). A discussion on the tectonic processes which can lead to the development of tensional features in the framework of large scale compressional regimes, with examples in the Mediterranean and Asian regions, is reported by Tapponier (1977). In particular, the occurrence of

Fig. 4a-c. Proposed genetic mechanism of tensional tectonics in some Mediterranean zones, such as the Tyrrhenian. a) The driving mechanism is constituted by the compressional regime (biggest arrows) which causes the bending and outward extrusion of the belt and forces it to collide with the adjacent foreland, as indicated by black arrows in the sections. See figs. 5a-d to understand how the peculiar geometry kind of kinematic and dynamic patterns implied by this mechanism may have developed in the Tyrrhenian-Apennines and Aegean systems. b) The collision between the extruding orogenic wedges and the foreland causes the imbrication of upper crustal slivers forming a new accretionary belt (dashed area). Once decoupled from the upper buoyant layer, the lower lithosphere sinks into the mantle, due to its negative density contrast. Extensional tectonics, associated to strike-slip movement, may develop in the wake of the migrating thrust belt, as tentatively sketched in the figure. Asthenospheric material intrudes in between the sinking slab and the upper crust (little empty arrows). The detachment between the upper and lower lithosphere has been placed for simplicity at the basis of the upper crust, but in reality it can occur everywhere inside the lower crust. c) At a certain stage of subduction, the downward bending of the sinking lithosphere induces intense stresses (tensional in the upper part and compressional in the lower part) in the most flexured sector of the slab. Stresses induced by the downward pull of the deepest subducted body might cause the detachment of the weakened slab, as discussed in the last section of the paper.

secondary tensional zones in front of an indentation mechanism can be quantitatively predicted by slip-line fields computed for plastic bodies deforming in plane strain (Tapponier and Molnar, 1976; Ranalli, 1995). The possibility that secondary tensional features may develop within a collisional pattern has also been demonstrated by numerical and experimental investigations (see, *e.g.*, Peltzer and Tapponier, 1988; Ratschbacher *et al.*, 1991; Faccenna *et al.*, 1996; Albarello *et al.*, 1997). This kind of mechanism may generate relatively large basins, as occurred in the Tyrrhenian and Aegean zones. The conditions which led to the formation of the above basins suggest that this kind of mechanism tends to develop in the zones where an orogenic belt is stressed more or less parallel to its main trend (see fig. 1), as already suggested by Brunn (1976) and Mantovani *et al.* (1996).

5. Evolutionary reconstruction

The proposed evolution is illustrated by four paleogeographic maps (fig. 5a-d) which show the paleopositions of major structural/tectonic elements in the middle-upper Miocene, the late Miocene-early Pliocene, the late Pliocene-early Pleistocene and the Present time. Each of these maps illustrates the structural/tectonic setting at the end of an evolutionary phase, just before the cessations of important lithosphere consuming processes, which implied significant reorganizations of shortening patterns and plate/microplate kinematics.

The proposed scheme is the result of a long series of attempts (see Mantovani *et al.*, 1981, 1982, 1985, 1990, 1992, 1993a,b, 1994, 1996) through which we tried to find physically plausible and coherent explanations for the major deformations observed in the study area. In our approach, we first try to identify the kinematic reconstructions which can best account for the peculiar space-time distribution of major compressional, tensional, transcurrent deformations, then we choose among the possible kinematic solutions the ones for which the most coherent and physically plausible geodynamic interpretation, even though qualitative, can be found. Due to the complexity of the problem, the procedure followed cannot always be framed in the relatively simple scheme mentioned above.

The reconstruction of local tectonic patterns is not an aim of this paper; the proposed evolutionary scheme must only be taken as a tentative recognition of the general evolutionary features and of large and regional scale tectonic processes.

Plate motions mentioned in the text are referred to a fixed Eurasian frame unless different indications are given.

Most references concerning the deformations mentioned in the text are given in the work of Babbucci *et al.* (1997), which also provides a more detailed description of the supporting evidence.

The arguments reported in support of the proposed evolution are only based on qualitative comparison between expected and observed deformation. However, in spite of its qualitative nature, the constraining power of this approach may be significant, since the possibility to provide plausible and coherent explanations for a fairly complex time-space distribution of compressional, tensional and shear deformation in the whole region considered may constitute an important discriminating factor among different geodynamic interpretations. Thus, we believe that the proposed interpretation may represent a useful working hypothesis, for future attempts at geodynamic reconstructions and for planning the most resolving quantitative analyses. A preliminary attempt to carry out numerical simulations of the proposed tectonic processes in a zone of the study area is given by Albarello *et al.* (1997).

5.1. *Middle Miocene tectonic setting*

Figure 5a shows a tentative reconstruction of the geometry of major deformation belts and structural domains around the middle Miocene (see, *e.g.*, Boccaletti and Guazzone, 1974; Channell and Horvath, 1976; Biju-Duval *et al.*, 1977; Dewey and Sengor, 1979; Le Pichon and Angelier, 1979, 1981; Kissel *et al.*, 1985; Dercourt *et al.*, 1986; Sorel and Mercier, 1988; Royden, 1988). The choice of this particular time is due to the fact that it preceded a drastic change of tectonic pattern in the whole Central-Eastern Mediterranean region.

An important structural element which, in our opinion, has strongly conditioned the successive evolutionary pattern is the presence of the Tethyan belts (pink and violet zones in the maps), *i.e.* the intermediate massifs and the main remnants left by the closure of the Tethyan oceanic domain (see, *e.g.*, Boccaletti and Dainelli, 1982). The proposed shape of these belts is only indicative, as is their lithological meaning, or affinity with other zones reported in the maps in the same colour. We tentatively propose an almost E-W alignment for these belts, in line with evolutionary reconstructions of some sectors of them given by other authors (see, *e.g.*, Le Pichon and Angelier, 1979 and references). The African/Adriatic domain is completely surrounded by orogenic belts (Maghrebides, Apennines, Southern Alps, Dinarides, Hellenides, Hellenic arc, Cyprus arc) which represent the African margin deformed by the previous collisional processes with Eurasia. We assume that, at the middle Miocene, the Carpatho-Balkan-Pannonian region had already undergone the most intense deformations (see, *e.g.*, Channell and Horvath, 1976; Horvath, 1984; Royden *et al.*, 1983; Royden, 1988). Anyway, we make reference to the above authors and the relevant literature for the evolutionary reconstruction of this area.

The middle Miocene tectonic/structural setting of the Central Mediterranean region mainly resulted from the opening of the Western Mediterranean basin and the related thrust belt-foredeep migration and magmatic activity (see, *e.g.*, Biju-Duval *et al.*, 1977; Scandone, 1979; Réhault *et al.*, 1987; Beccaluva *et al.*, 1994).

The lateral extent of the Adriatic promontory was much greater than the present one (identified by the darker colour). The restoration of this foreland domain back to the Tortonian is tentatively based on shortening estimates across the Alps, Dinarides, Hellenides, Apennines and Maghrebides (see, *e.g.*, Boccaletti *et al.*, 1976; Mercier *et al.*, 1979, 1989; Burchfiel, 1980; Laubscher, 1983; Ghisetti and Vezzani, 1984; Horvath, 1984; Castellarin and Vai, 1986; Philip, 1987; Catalano *et al.*, 1989; Sartori, 1989; Schmid *et al.*, 1989; Patacca *et al.*, 1990). The north-western protuberance of the Adriatic promontory was deeply indented into the Eurasian domain, after its previous continental collision in the Western Alps (see, *e.g.*, Semenza, 1974; Channell and Horvath, 1976; Laubscher, 1983, 1988; Dal Piaz, 1995). Due to this embedding, the North-Western Adriatic edge was most probably characterized by a very low mobility and, consequently, it could have represented a sort of hinge zone for the Adriatic block.

The Corsica-Sardinia microplate was already in its present position and was separated from the African-Adriatic foreland by an orogenic belt, constituted by Alpine and pre late Miocene Apenninic units (see, *e.g.*, Biju-Duval *et al.*, 1977; Dercourt *et al.*, 1986; Réhault *et al.*, 1987; Patacca and Scandone, 1989; Vigliotti and Langenheim, 1995).

A feature of the Tortonian structural setting which significantly influenced the successive evolution of the Central Mediterranean was the presence of thinned lithosphere along the Northern African margin, from the Ionian to the Levantine basin (see, *e.g.*, Biju-Duval *et al.*, 1977; Rossi and Sartori, 1981; Scandone *et al.*, 1981; Dercourt *et al.*, 1986; Malinverno and Ryan, 1986; Patacca *et al.*, 1993).

The Iblean zone was still connected with Africa. The region corresponding to the present Sicily channel and surroundings was affected by SW-NE compressional stresses (see, *e.g.*, Illies, 1981; Reuther, 1987; Boccaletti *et al.*, 1990; Barrier, 1992).

The distribution of Oligo-Miocenic magmatic activity in Sardinia and North Africa, which ended about 10-15 My ago (see, *e.g.*, Savelli *et al.*, 1979; Bellon, 1981) suggests that after the opening of the Balearic basin an extended arcuate edifice of subducted lithosphere was lying beneath the Apenninic and Maghrebian belts (see Mantovani *et al.*, 1992, 1996 for a tentative perspective view).

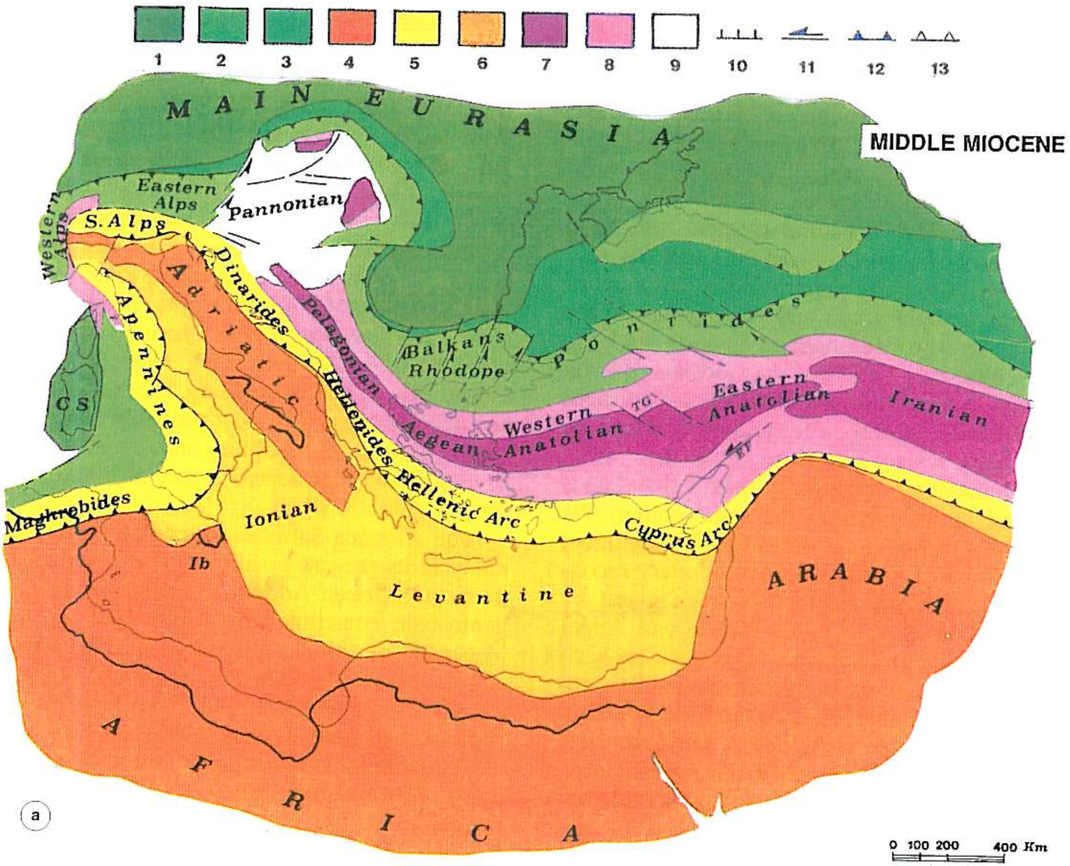
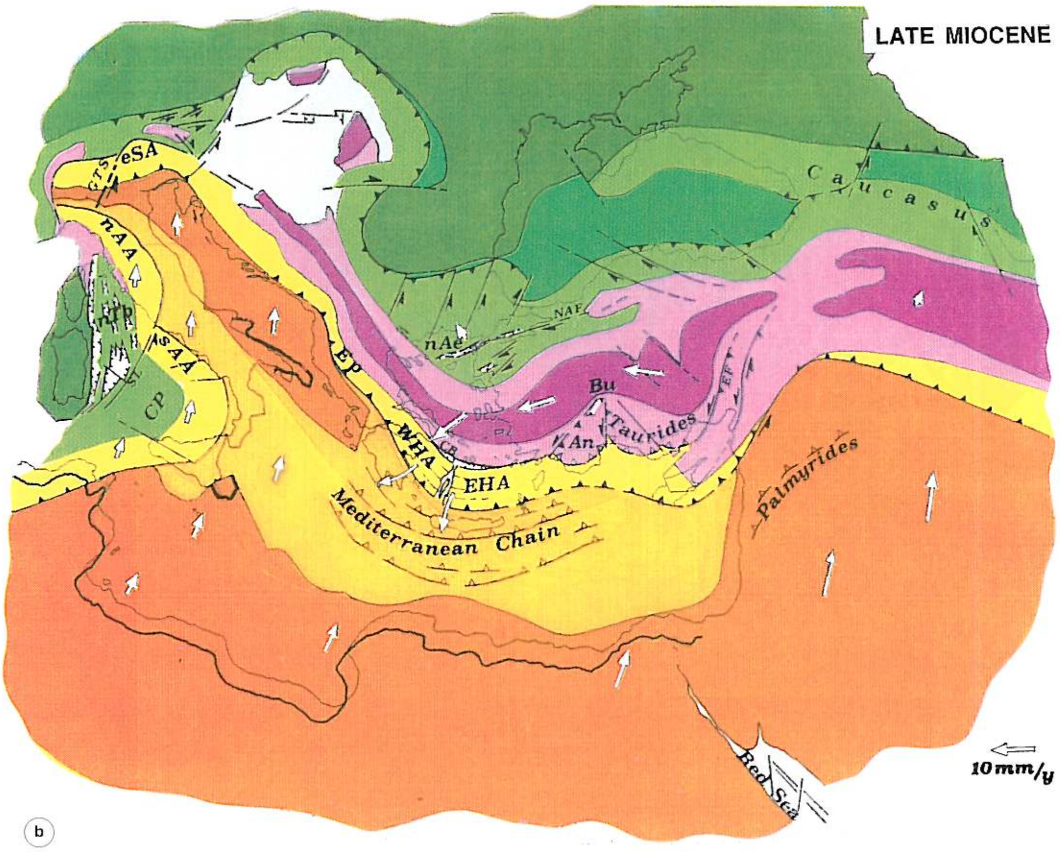


Fig. 5a,b. a) Tentative reconstruction of the middle-upper Miocene paleogeographic setting in the Central-Eastern Mediterranean region (see text for comments). Eurasia domain: 1) undeformed foreland; 2) deformed margin; 3) thinned margin. Africa/Arabia domain: 4) undeformed foreland; 5) deformed margin; 6) thinned margin. Tethyan belts: 7) internal massifs; 8) outcropping oceanic remnants (flysch, ophiolites, etc.) and metamorphic zones. 9) Main basins and troughs formed since the middle Miocene. 10, 11, 12) Main tensional, transcurrent and compressional features. 13) Intraplate deformations. CS = Corsica Sardinia block; EF = Ecemis fault system; Ib = Iblean plateau; TG = Tuz Golu fault system. The present shape of the Adriatic plate is reported, for reference, by the darkest colour. b) Tentative reconstruction of the paleogeographic setting in the

5.2. *First phase: from the middle-upper to the late Miocene*

The key events which caused a profound change of tectonic pattern around the middle-upper Miocene were the continental collision of Arabia against Eurasia and the decoupling of the Adriatic plate from the Eurasia domain, through the Giudicarie transpressional system. Most authors agree on the middle-upper Miocene as the timing of the Arabia-Eurasia continental collision (see, *e.g.*, Dewey *et al.*, 1973; Dewey and Sengor, 1979; Sengor and Yilmaz, 1981). Important evidence in this sense is the end of spreading activity in the Red Sea and the transcurrent motion in the Dead Sea fault system, which is interpreted as an effect of the Arabian continental collision (see, *e.g.*, Hempton, 1987 and references therein).



late Miocene-early Pliocene (see text for comments). An = Antalya region; Bu = Burdur zone; CB = Cretan basin; CP = Calabria-Peloritani block; EF = Ecemis fault system; EHA = Eastern Hellenic Arc; Ep = Epirus zone; GTS = Giudicarie transpressional fault system; eSA = eastern Southern Alps; nAA = north Apenninic Arc; nAe = north Aegean basin; nTb = north Tyrrhenian basin; NAF = North Anatolian fault system; SL = Selli line; sAA = southern Apenninic Arc; WHA = Western Hellenic Arc. Empty arrows indicate motion rates (scale in the figure). Some present geographical references, as the African coast, the South-Eastern Sicily coast and the Gargano-Apulia zone, are reported in the respective blocks in order to facilitate the recognition of the proposed kinematics.

Due to the almost triangular shape of the northern edge of the Arabian promontory, the indentation caused a progressive separation between the Anatolian and Iranian segments of the Tethyan belts, which were extruded in almost opposite directions. In this context, Eastern Anatolia underwent a strong shortening, by thrusting and uplift, and a general translation roughly towards NW, favoured by a series of SE-NW dextral transcurrent faults (fig. 5a). This kinematics is suggested by the pattern of compressional and strike-slip deformations which affected the Pontides, at the border with the Black Sea Mesozoic basin.

The Iranian segment underwent a roughly NEward displacement, at the expense of the thinned margin which still existed NE of it. The consequent subduction process which consumed this zone is evidenced by the occurrence of calc-alkaline volcanism in the Great

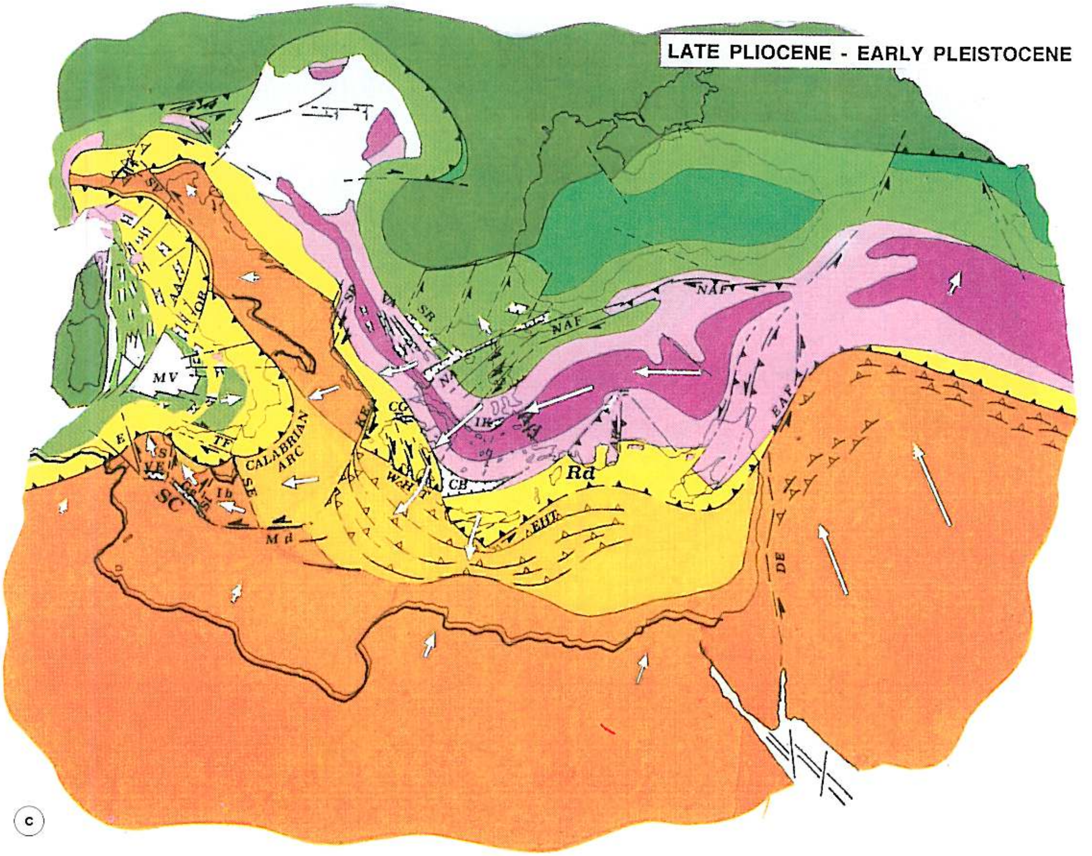
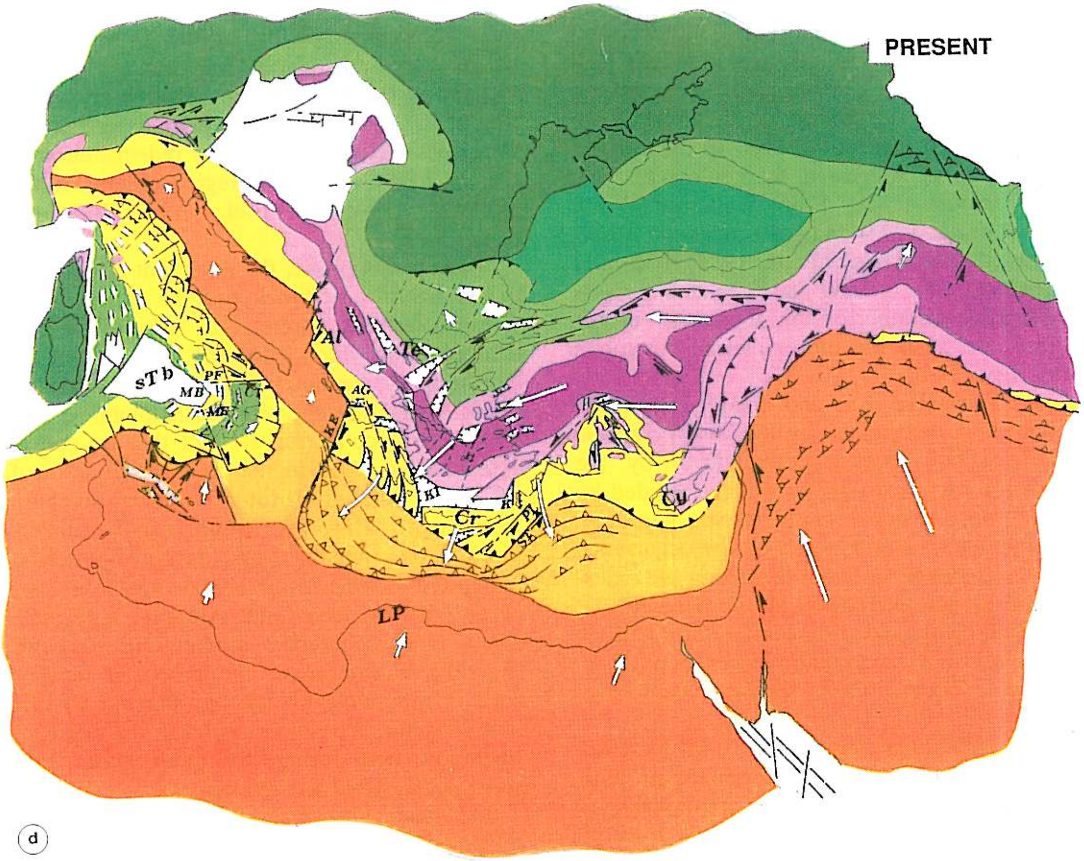


Fig. 5c,d. c) Tentative reconstruction of the paleogeographic setting in the late Pliocene-early Pleistocene (see text for comments). AA = Anzio-Ancona (Olevano-Antrodoco) discontinuity; AK = Aksu lineament; CB = Cretan basin; CG = Corinth Gulf; DE = Dead Sea fault system; E = Egadi fault system; EAF = Eastern Anatolian fault system; EHT = Eastern Hellenic Trench; Ge = Gela basin; Ib = Iblean plateau; IK = Ikaria basin; KE = Kefallinia discontinuity; Md = Medina shear zone; MV = Magnaghi-Vavilov basin; NA = North Aegean trough; NAF = North Anatolian fault system; OR = Ortona-Roccamonfina discontinuity; Rd = Rhodes Island; S = Sciacca fault; SC = Sicily Channel; SE = Syracuse escarpment; SI = Scicli-Comiso fault system; SP = Scutari-Pec transpressional system; SR = Struma fault system; SV = Schio-Vicenza fault system;

Caucasus (see Babbucci *et al.*, 1997 for references about the above deformation events). Around the late Miocene-early Pliocene sets of conjugate transcurrent fault systems activated in Eastern Anatolia, in response to the Arabia indentation, favouring the westward extrusion of crustal wedges. To explain this phenomenon, McKenzie (1972) and Dewey and Sengor (1979) suggested that after the uplift of the Arabia-Eurasia collision zone by 1-2 km, it became more efficient to accommodate the convergence by moving Anatolian wedges to the west, over the subducting African lithosphere at the Hellenic trenches.

The westward drifting of the Eastern Anatolian wedges was mainly accommodated by a S-SW bowing of the Tethyan belts, especially in the Aegean region (see, *e.g.*, Le Pichon and



TF = Taormina fault system; VA = Vardar fault system; VE = Ventura plateau; WHT = Western Hellenic Trench. d) Tentative reconstruction of the present paleogeographic setting (see text for comments). AG = Ambracique Gulf; Al = Albanides belt; Cr = Crete Island; CT = Crati trough; Cy = Cyprus Island; K = Karpathos trough; KE = Kefallinia discontinuity; KI = Kithira trough; LP = Libyan Promontory; MB = Marsili basin; ME = Mesima trough; PF = Palinuro fault; PL = Pliny fault system; ST = Strabo fault system; sTb = southern Tyrrhenian basin; Te = Thessaloniki zone; VF = Vulcano fault system. Some present geographical references, as the African coast, the South-Eastern Sicily coast and the Gargano-Apulia zone, are reported in the respective blocks in order to facilitate the recognition of the proposed kinematics.

Angelier, 1979). The formation of the Aegean arc had significant consequences in the zones lying north and south of it. The North Aegean region was contemporaneously affected by E-W compression, due to the convergence between Eastern Anatolia and the Adriatic, and roughly N-S extension, in the wake of the Aegean arc, bowing southward. This combination of stresses led to the formation of a SW-NE trending fault system (see, *e.g.*, Mercier *et al.*, 1989).

The outward migration of the Aegean arc was accommodated by the consumption of the Northern African margin (Ionian and Levantine forelands), with a mechanism similar to that illustrated in fig. 4a-c. Due to bowing, the Hellenic arc began to decouple from the internal belts, (fig. 5b) triggering S-N extensional activity in the Cretan basin and the incipient separation between the eastern (Crete-Rhodes) and western (Peloponnesus) sectors of the Hellenic arc (figs. 5b,c), although the most intense development of the bowing of the Hellenic arc and of its side effects occurred in the successive phase, as suggested by geological (see, *e.g.*, Dercourt *et al.*, 1986; Underhill, 1989; Reuther *et al.*, 1993) and paleomagnetic observations (see, *e.g.*, Kissel *et al.*, 1985). A similar mechanism of bowing and related effects might also account for the separation of the Cyprus arc from the more internal Anatolian belts (figs. 5b,c).

In the Antalya zone and Taurides, the E-W compression determined by the bending of the Anatolian belts was accommodated by the southward escape of crustal wedges, guided by lateral transpressional fault systems (fig. 5b). This hypothesis is suggested by the occurrence of compressional deformations along the external front of the wedges, at the border with the Levantine basin, and of tensional deformations in the wake of the extruding wedges, in the Burdur zone (Barka, 1992; Taymaz and Price, 1992).

In the Rhodope region, the E-W to ESE-WNW compressional regime connected with the Arabian indentation was accommodated by the lateral escape of crustal wedges towards the Eurasian foreland, guided by a system of NNE transcurrent faults (fig. 5b). This hypothesis might account for the occurrence of compressional tectonics at the border between the extruding wedges and the Balkan belt and for the development of tensional tectonics in the wake of the extruding wedges, in Westernmost Anatolia and in the North Aegean region. This last tensional event might be responsible for the detachment of a number of fragments, now corresponding to North-Eastern Aegean islands, from the Rhodope massif.

In the Central Mediterranean, the activation of the left-lateral Giudicarie fault system and the lateral escape of crustal wedges in the Eastern Alps (see, *e.g.*, Castellarin and Vai, 1986) allowed the Adriatic plate to decouple from its north-western termination, sutured to the Eurasia domain, and to initiate a roughly NNE to NEward movement (Semenza, 1974; Mantovani *et al.*, 1992, 1996). This motion was then accommodated by further shortening in the Southern Alps, by lateral escape of crustal wedges in the Eastern Alps, by the consumption of the Eastern Adriatic margin beneath the Dinarides-Hellenides belt and redistribution of masses in the Pannonian region (Channell and Horvath, 1976; Burchfiel, 1980; Castellarin *et al.*, 1992).

During this phase, the divergent motion between the Adriatic block, and the almost stable Corsica-Sardinia microplate induced a tensional regime which led to the formation of the Northern Tyrrhenian basin (see, *e.g.*, Sartori, 1990). This interpretation (Mantovani *et al.*, 1996) can account for the fact that no coeval extension occurred in the zone lying south of the Selli line (the future Central-Southern Tyrrhenian). This last feature can be understood if one considers that during this period the future Southern Tyrrhenian basin lay between the African and Adriatic forelands, which were coherently moving NE to NNE (fig. 5b). Furthermore, the proposed mechanism is coherent with the fact that no significant orogenic activity occurred in the adjacent Apenninic belt during this phase (see, Babbucci *et al.*, 1997 for references).

It is reasonable to think that the lateral displacement of the Adriatic plate had significant consequences on its western subducted margin, lying beneath the Corsica-Sardinia block. In particular, Mantovani *et al.* (1992) advanced the hypothesis that in response to E-W divergence the slab underwent stretching and then break-off. This last hypothesis might explain the time and space location of a Pliocenic calc-alkaline magmatic arc (see fig. 4 in Babbucci *et al.*, 1997) in the Tyrrhenian area (Savelli *et al.*, 1979; Bigi *et al.*, 1989; Sartori *et al.*, 1989; Francalanci and Manetti, 1994).

5.3. *Second phase: from the late Miocene-early Pliocene to the upper Pliocene*

The structural-tectonic conditions which progressively led to the end of the first evolutionary phase and to the beginning of the second one (fig. 5c) developed along the collisional boundary between the Adriatic plate and the Pelagonian-Aegean segments of the Tethyan belts. Along this consuming border the resisting forces underwent a progressive strengthening, perhaps due to the increase in the crustal thickness of the Adriatic foreland (pre-Apulian zone) entering the trench zone or to the considerable accumulation of buoyant material at this consuming boundary. Around the late Messinian, these forces probably reached such a high value as to trigger the end of subduction beneath the Hellenides and the activation of a new consuming process on the other side of the Adriatic plate, at the border with the Southern Apenninic arc, where thinner crust was presumably present (*e.g.*, Patacca *et al.*, 1993). This result was achieved by a relatively complex reorganization of plates/microplates kinematics in the Central Mediterranean region. The Adriatic plate, stressed by the Aegean/Balkan system after the suture of the Hellenides boundary, underwent a clockwise rotation, characterized by a roughly westward motion in the south and roughly NWward movements in the north (fig. 5c). This new kinematics could explain the activation around the Messinian of the Schio-Vicenza shear zone (Cantelli and Castellarin, 1994). This allowed a new decoupling, with a different geometry, between the main Adriatic plate and its Padanian fragment (fig. 5c). The westward motion of the Southern Adriatic plate, along with the Northern Ionian and Iblean wedge, was allowed by the activation of a left lateral shear zone, the Medina fault system, which decoupled the above structures from Africa, involved in a roughly NNEward drifting. The resulting E-W convergence between the Northern Ionian-Iblean foreland zone and the African Tunisian promontory was accommodated by the lateral escapes of the Gela and Ventura crustal wedges. In the wake of these wedges extensional tectonics developed in the Sicily channel as evidenced by the formation of some troughs (*e.g.*, Finetti and Del Ben, 1986; Reuther, 1987). The lateral escapes of the Ventura and Gela wedges were allowed by transcurrent guides, as the Egadi, Sciacca and Scicli fault systems (see Ben-Avraham and Grasso, 1990; Babbucci *et al.*, 1997 for references). On its turn, the roughly NWward indentation of the Sicilian fragments (Gela and Ventura) was accommodated by the eastward escape of crustal wedges in the adjacent Apenninic-Alpine belt (the future Central Tyrrhenian zone), at the expense of the western thinned Adriatic margin, whose consumption represented in effect the final objective of the complex mobilization of crustal wedges described above.

In the proposed kinematic scheme, the considerable shortening rate recorded in the Southern Apennines during the Pliocene resulted from the relative motion between the Southern Adriatic plate, moving roughly westward, and the Alpine-Apenninic wedges escaping eastward, in response to the indentation of the Sicilian fragments. The consumption of the Adriatic foreland in the Southern Apenninic arc developed by a delamination mechanism, like the one shown in fig. 4a-c, implying imbrication of upper crustal units and gravitational sinking of the lower lithosphere. The time pattern of this last process has been accurately reconstructed by geological analyses (Casnedi *et al.*, 1982; Royden *et al.*, 1987; Patacca and Scandone, 1989).

We interpret the coeval E-W crustal stretching in the Central Tyrrhenian region (Magnaghi-Vavilov basin) as an effect of the tensional regime which occurred in the wake of the escaping crustal wedges in the Southern Apennines. The plausibility of this kind of mechanism, as responsible for the formation of tensional features, is supported by physical considerations and experimental and numerical modeling (*e.g.*, Tapponier, 1977; Peltzer and Tapponier, 1988; Ratschbacher *et al.*, 1991; Faccenna *et al.*, 1996; Albarello *et al.*, 1997).

During this phase, the Ionian foreland underwent downward flexure in response to the outward migration of the Calabria-Peloritani block. This phenomenon might explain the Pliocene reactivation of vertical movements along the borders (the Syracuse and Apulian escarpments) between the subsiding Ionian area and the adjacent continental margins, *i.e.* the Iblean block to the west and the Adriatic plate to the east (Carbone *et al.*, 1982; Auroux *et al.*, 1984; Finetti and Del Ben, 1986) which did not undergo significant vertical movements. This interpretation is coherent with the fact that the vertical offset along the Syracuse escarpment decreases from north to south (Reuther *et al.*, 1993). The vertical movements at the Apulian escarpment might also have been favoured by the downward flexure of the adjacent Levantine foreland involved in the subduction process beneath the Western Hellenic arc.

It seems opportune to discuss the compatibility of the proposed Pliocenic kinematic pattern of the Adriatic plate with the relevant paleomagnetic data. A review and analysis of the paleomagnetic observations so far carried out in the Apulian foreland is given by Meloni *et al.* (1997). This analysis suggests that the existence of a (slight) post-Cretaceous counterclockwise rotation of the Adriatic plate is still controversial. Paleomagnetic data from the Oligocene sediments in the Salento peninsula (fig. 1) were interpreted as evidence of a clockwise rotation (Tozzi *et al.*, 1988), but a reanalysis of these data discouraged their use for geodynamic reconstruction (Bazhenov and Shipunov, 1991). Data from Plio-Pleistocene clayey units in the Apulian foreland (Scheepers, 1992) show that no rotation has occurred since the late Pliocene, in line with the kinematic pattern proposed in fig. 5d. Another deformation which seems to be compatible with the proposed Adriatic kinematics is the occurrence of extensional tectonics, with formation of troughs and magmatic activity (*e.g.*, Stanishkova and Slejko, 1991) in the Struma and Vardar zones (fig. 5c). These features clearly imply a divergence between the Rhodope massif (Eurasia) and the Pelagonian units, which could be imputed to the roughly westward displacement of these belts, in close connection with the Aegean segment of the Tethyan belts. If the Adriatic plate had not rotated clockwise, as we suggest, it would not have allowed any westward displacement of the Pelagonian belts and thus no extension would have occurred in the Vardar-Struma zone. It is interesting to note that the cited extension is limited to the Vardar-Struma sector, in line with the fact that the mobility of the Pelagonian belts is mostly confined to the segment lying south of the Scutari-Pec fault system, which decouples the Hellenides from the Dinarides. The differentiated kinematic behaviour of the Hellenides and the Dinarides sectors is also indicated by paleomagnetic observations (Kissel *et al.*, 1985; Speranza *et al.*, 1995) which suggest a Plio-Quaternary clockwise rotation of about 25° for the sector lying south of the Scutari-Pec fault system and almost no rotation for the Dinarides.

The important change of plate kinematics implied by our evolutionary reconstruction around the late Miocene may explain why orogenic activity in the Apenninic belt underwent a relatively sudden and intense reactivation in the Messinian, after a relatively long (2-3 My) quiescence (see Babbucci *et al.*, 1997 for references).

Since the late Messinian-early Pliocene, the Peloponnesus and the Epirus region, that is the Aegean zones which were most directly involved in the collision with the Southern Adriatic plate underwent crustal thickening and uplift (see Babbucci *et al.*, 1997 for references). In the Peloponnesus, E-W shortening was also accommodated by the lateral escape (roughly southward) of narrow crustal wedges, at the expense of the adjacent African margin along the Western Hellenic trench (fig. 5c). In the wake of these escaping wedges, extensional features developed in Central-Northern Greece (figs. 5c,d), causing the formation of troughs such as the Corinth and Ambracique gulfs (Berckhemer and Kowalczyk, 1978; Stiros, 1988). During this phase, the onland segment of the Kefallinia discontinuity behaved as a transpressional dextral fault between the Adriatic plate and the Peloponnesus wedges (see, *e.g.*, Sorel, 1989; Mercier *et al.*, 1989).

In response to the suture in the Hellenides, the SWward bowing of the Aegean Tethyan belts underwent a significant acceleration (fig. 4c). This hypothesis is suggested by a number of observations:

– The most intense extensional activity in the Northern Aegean zone, interpreted as an effect of the above bowing, mainly occurred in the Pliocene, with a SW trend.

– S-N extensional activity occurred in the Central-Western Cretan basin which during this phase almost reached its present configuration.

– The consumption of the North African lithosphere beneath the outward migrating Hellenic arc (Peloponnesus and Crete-Rhodes) involved the accumulation of offscraped buoyant crustal material along the trench zone, which led to the formation of the long deformation belt (the so called «Mediterranean Ridge») which now runs from the Kefallinia line to the Cyprus arc (see, *e.g.*, Finetti, 1976).

– Mercier *et al.* (1987) suggested that the internal Aegean zone underwent a compressional phase during the lower Pliocene; the effects of this phenomenon, in terms of intense fracturation and block tilting, are most evident in the Cyclades arc (fig. 6), *i.e.*, in the internal massifs (violet belt in the maps of figs. 5a-d) and are instead moderate in the external arc (Peloponnesus, Crete, Rhodes) and in the Northern Aegean zone.

– The major morphological features of the Aegean zone (fig. 6) suggest that the strong E-W compression which affected the Cyclades arc, just after the suture in the Hellenides, might have caused the formation of SW-NE fractures and the consequent SWward extrusion of crustal wedges in the most arcuate segment of the Cy-

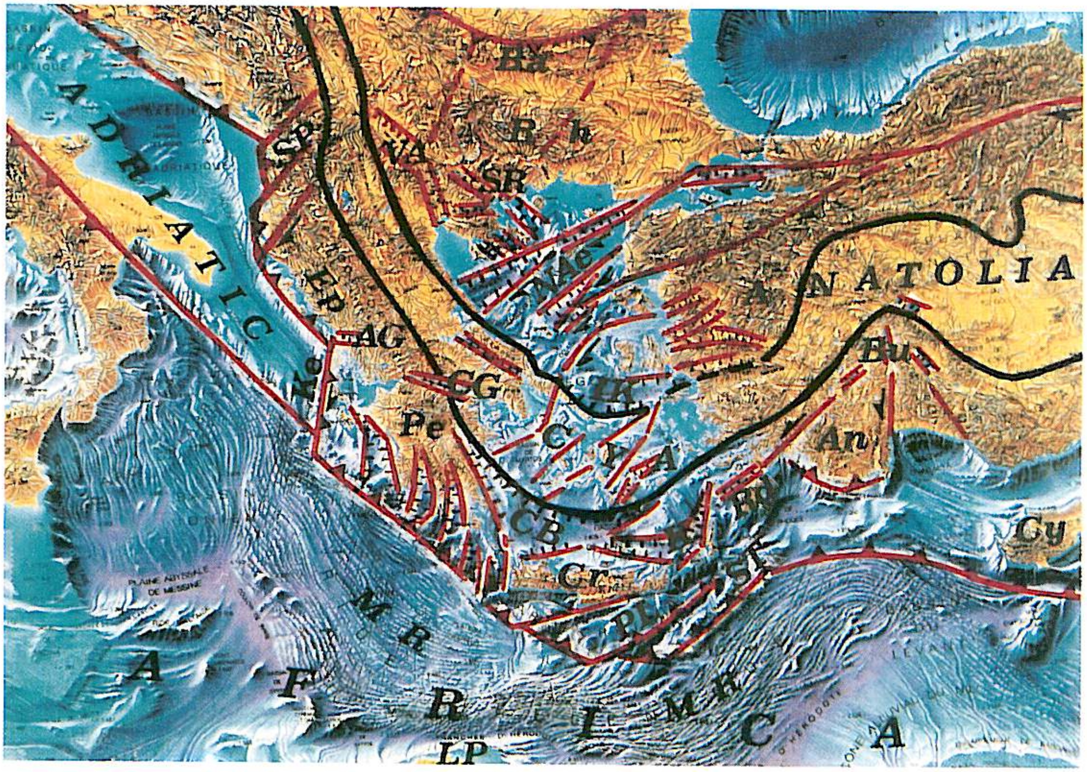


Fig. 6. Main morphological features of the Aegean and surrounding zones (from Le Pichon and Biju-Duval, 1990) and major tectonic lineaments (red lines). To help the correspondence with the evolutionary maps (figs. 5a-d) a rough contouring of the internal Tethyan massifs is reported (black solid lines). AG = Ambracique Gulf; An = Antalya zone; Ba = Balkan belt; Bu = Burdur zone; CB = Cretan Basin; CG = Corinth Gulf; Cr = Cretan segment; Cy = Cyprus arc; CYA = Cyclades Arc; Ep = Epirus zone; Ik = Ikaria basin; K = Karpathos trough; Ke = Kefallinia discontinuity; LP = Libyan Promontory; MR = Mediterranean «Ridge»; Nae = North Aegean region; Pe = Peloponnesus; PL = Pliny trench; Rd = Rhodes segment; Rh = Rhodope massif; SP = Scutari-Pec fault system; SR = Struma fault system; ST = Strabo Trench; VA = Vardar fault system.

clades arc. This mechanism might explain the formation of the Ikaria basin, in the wake of escaping wedges (fig. 6). The weak lateral constraint which could have favoured this SWward expulsion of wedges in the Cyclades arc is most probably given by the stretched crust of the Central-Western Cretan basin.

The fact that the most intense effects of the E-W compressional regime occurred in the internal massifs (Cyclades arc) might suggest that this structure mostly sustained the E-W compression between the Eastern Anatolian system and the Adriatic plate during this phase. The relatively short duration of this compressional phase (Mercier *et al.*, 1987) might be due to the fact that the «Hellenides suture» was soon followed by the activation of the consuming process beneath the Southern Apennines and by the consequent clockwise rotation of the Adriatic plate, which then allowed a considerable attenuation of E-W compressional stresses in the Aegean Tethyan belts.

Since the early Pliocene, the North Anatolian fault was almost completely active from east to west. The greater freedom that this fault system allowed the Anatolian wedge for a westward escaping might explain the observed slowdown of compressional deformations in the Pontides. Other important faults favoured the westward escape of the Anatolian wedges, like the East Anatolian, Ecemis and Tuz Golu fault systems (see Dewey and Sengor, 1979; Barka, 1992).

Around the early Pliocene, a renewal of extensional activity in the Red Sea/Gulf of Aden and of transcurrent activity in the Dead Sea fault system took place (see, *e.g.*, Hempton, 1987 and references therein). This phenomenon might be a consequence of the lateral escapes of Anatolian wedges on one side and of Iranian wedges on the other side, which allowed an acceleration of the Arabian block with respect to Africa.

5.4. *Third phase: from the late Pliocene to the Present*

The two key events that around the late Pliocene-early Pleistocene determined a significant change of tectonic pattern in the Central and Eastern Mediterranean region and the starting of the present evolutionary phase (fig. 5d), were the suture of the consuming boundary in the Southern Apennines (see, *e.g.*, Patacca and Scandone, 1989) and the continental collision, with the consequent slowdown of convergence rate, in the Hellenic trench located just south of Crete (see, *e.g.*, Armijo *et al.*, 1992). As in previous cases, the increase in resisting forces at the above consuming boundaries made the activation of other shortening processes in the surrounding regions more efficient.

In the Central Mediterranean, one of these new processes was crustal thickening and consequent rapid uplift in the whole Africa-Adriatic collision zone, *i.e.*, the Calabrian arc and Southern Apennines, as clearly indicated by neotectonic data. Another important effect of the suture in the Southern Apennines was a new change of Adriatic kinematics. After this suture the mobility of the Adriatic plate in the roughly E-W direction was strongly limited. The occurrence of SE-NW Quaternary shortenings in the eastern Southern Alps and the renewal of E-W to WSW-ENE thrust deformations in the outer Hellenides suggest that the Quaternary kinematics of the Adriatic plate was mainly characterized by a minor counterclockwise rotation (fig. 5d). The present motion could be tentatively described by a pole located in the Western Mediterranean basin (41°N, 6°E), with an angular velocity of roughly 0.15°/My. The Africa-Adriatic convergence was accommodated by lateral escapes of crustal wedges towards the foreland zones where consumable lithosphere was still present, *i.e.*, the Ionian and the Northern Adriatic areas. The consumption of the Ionian foreland was achieved by the lateral escape of the Calabrian wedge (fig. 5d), with a mechanism like the one illustrated in fig. 4a-c. This last extrusion process is suggested by some major deformations which have occurred since the late Pliocene-early Pleistocene:

– Intense compressional deformations along the collisional front of the Calabrian arc with the Ionian foreland (see, *e.g.*, Rossi and Sartori, 1981).

– NW-SE extensional tectonics in the Southernmost Tyrrhenian (Marsili basin), and the internal part of Calabria (Sartori, 1990).

– Emphasis of the SEward arcuation of the Calabrian arc-Southernmost Apennines, as indicated by paleomagnetic data (see, *e.g.*, Sagnotti, 1992; Scheepers *et al.*, 1994).

– Acceleration of tectonic activity in the Calabrian arc, as indicated by the formation of a number of important discontinuities, as the Palinuro and Catanzaro transcurrent fault systems, the decoupling of the Peloritani block from the Southern Calabria, the acceleration of the extensional rate in the Mesima and Crati troughs (see Babbucci *et al.*, 1997 for references).

The consumption of the Northern Adriatic foreland was achieved by the outward migration of crustal wedges in the Central-Northern Apennines (fig. 5d), through a mechanism of the type shown in fig. 4a-c. The most intense phase of this process was driven by the longitudinal stresses (N to NNW) which affected this part of the belt after the suture of the Southern Apennines. This hypothesis is suggested by the fact that the Southern Apennines after their suture moved in close connection with the Adriatic plate, roughly towards N to NNW (see fig. 5d) and that this movement was accommodated in different ways by the Adriatic plate and the Apenninic belt. The drifting of the Adriatic block, characterized by scarce or null internal deformation, was mainly absorbed by underthrusting processes along its northern border, in the eastern Southern Alps, whereas the coherent drifting of the Southern Apennines was mostly accommodated by the deformation of Central-Northern Apennines, mainly through oroclinal activity and the consequent outward migration of crustal wedges, at the expense of the adjacent Adriatic foreland. This consuming process was probably responsible for the formation of the lithospheric slab beneath the Northern Apennines, which is now evidenced by subcrustal earthquakes (Amato *et al.*, 1993).

The Pliocenic SEward displacement of Calabria produced a significant change in the boundary conditions of the other extrusion process, involving the Iblean wedge. After the removal of the Calabrian «obstacle», the Iblean block started to move roughly northward, guided by the Vulcano fault system. As argued by Mantovani *et al.* (1996), this hypothesis can explain the major variations of strain regimes in the Sicily channel and in the Northern Sicilian belts since the late Pliocene-early Pleistocene.

The suture of the Southern Apennines and the consequent stop of the Adriatic clockwise rotation caused a strengthening of the E-W compressional stresses between the Tethyan belts and the Adriatic block, which determined a renewal of thrusting activity and uplift in the collision zones (Epirus and Peloponnesus). In the Peloponnesus the E-W compression emphasized the roughly southward expulsion of crustal wedges, at the expense of the Ionian foreland in the Western Hellenic trench. This hypothesis is suggested by the contemporaneous occurrence of crustal thickening-uplift in the bulk of the Peloponnesus and of S-N extensional tectonics in Central Greece and Southern Albania, with particular regard to the Corinth and Ambracique troughs. The fact that the observed S-N extensional rate was considerably more intense in the eastern segment of the Corinth trough (50%) than in the western segment (10%) suggests that the southward escape of the Eastern Peloponnesus slices was faster than that of western slices.

The other incipient suture mentioned at the beginning of this section, *i.e.*, the one between the African margin (Libyan promontory) and Crete had opposite consequences in the eastern and western sectors of the Hellenic arc. The Crete-Rhodes sector, squeezed in between the African continental margin and the advancing Anatolian block, underwent a SEward bowing (fig. 5d). This hypothesis is indicated by the present morphology of this sector (see fig. 6) and by the occurrence of extensional tectonics in the internal zones (Eastern Cretan basin and Karpathos trough) and of transpressional deformations along the oblique collisional border with the adjacent African margin, *i.e.*, the Pliny and Strabo trenches (see, *e.g.*, Le Pichon and Angelier, 1979). This deformation mechanism is still active, as indicated by focal mechanisms in the Eastern Hellenic arc (Papazachos and Kiratzi, 1996; Viti, 1996).

Since the late Pliocene-early Pleistocene, the zone comprised between Crete and the Peloponnesus underwent roughly E-W extensional deformations, as clearly testified by the formation of a number of S-N trending normal faults (see, *e.g.*, Armijo *et al.*, 1992). This strain regime might be explained by the slowdown of Crete with respect to the other parts of the Aegean system, and in particular with respect to the Peloponnesus, which had not changed its fast SWward drifting at the expense of the thinned Ionian foreland. This interpretation is also supported by the fact that the Crete-Libyan continental collision was coeval with the transition from a dominant S-N extensional regime to a dominant E-W tensional field in the Southern Aegean zone (Mercier *et al.*, 1989).

The present kinematic pattern proposed in fig. 5d represents a natural development of the previous evolutionary history and is also consistent with the available constraints on the present displacement and strain fields, provided by seismological, neotectonic and geodetic observations (see, Mantovani *et al.*, 1992, 1994, 1995, 1996 and references therein; Viti, 1996; Babbucci *et al.*, 1997).

6. Discussion

In the recent literature a number of authors have proposed that the downward pull of the subducted lithosphere, due to gravitational sinking, played a significant role in the Mediterranean evolution and, in particular, in the formation of major Neogenic basins such as the Tyrrhenian (see, *e.g.*, Malinverno and Ryan, 1986; Patacca and Scandone, 1989) and the Aegean (see, *e.g.*, McKenzie, 1978; Le Pichon, 1982; Taymaz *et al.*, 1991). Here we think it is opportune to clarify the differences between this point of view and our interpretation.

First of all, the divergence between the two hypotheses does not concern the occurrence of trench retreat (roll-back), thrust belt-foredeep migration and crustal stretching in its wake, which is supported by convincing evidence, both in the Tyrrhenian and Aegean regions. The discrepancy relates to the dynamic framework which is supposed to be responsible for the above processes.

For this reason, in the following discussion we will identify the first hypothesis with the term «slab-pull», which clearly specifies the underlying dynamics rather than with the term «roll-back» (often used in literature) which only recalls a kinematic feature which is recognized by both interpretations.

The «slab-pull» hypothesis, summarized in fig. 7, assumes that in the zones where a well developed subducted lithosphere exists, the sinking of the slab, driven only by gravity, becomes the main controlling factor of the observed surface deformations, since the retreat of the trench occurs faster than the convergence between the overthrusting block and the foreland. This scheme interprets the outward migration of the belt as an effect of «trench suction» induced by the sinking slab; crustal stretching in the hinterland is supposed to occur in the wake of the migrating belt (see, Malinverno and Ryan, 1986 and references therein).

This type of interpretation, at the light of the present knowledge, cannot be ruled out by theoretical or experimental considerations, but we think that it can hardly be applied in the region here considered, since its main implications do not seem to be coherent with the observed deformation patterns in the Tyrrhenian and Aegean zones.

Since the opening of the northern, central and southern parts of the Tyrrhenian basin occurred in different periods and with different tectonic styles both in the sea area and in the adjacent belt (see Babbucci *et al.*, 1997) each sector will be discussed separately.

6.1. *North-Western Tyrrhenian*

A basic feature of this extensional episode, occurred from the late Tortonian to an intra Messinian time in the zone lying north-west of the Selli line (Sartori, 1990), is the lack of significant compressional deformation in the adja-

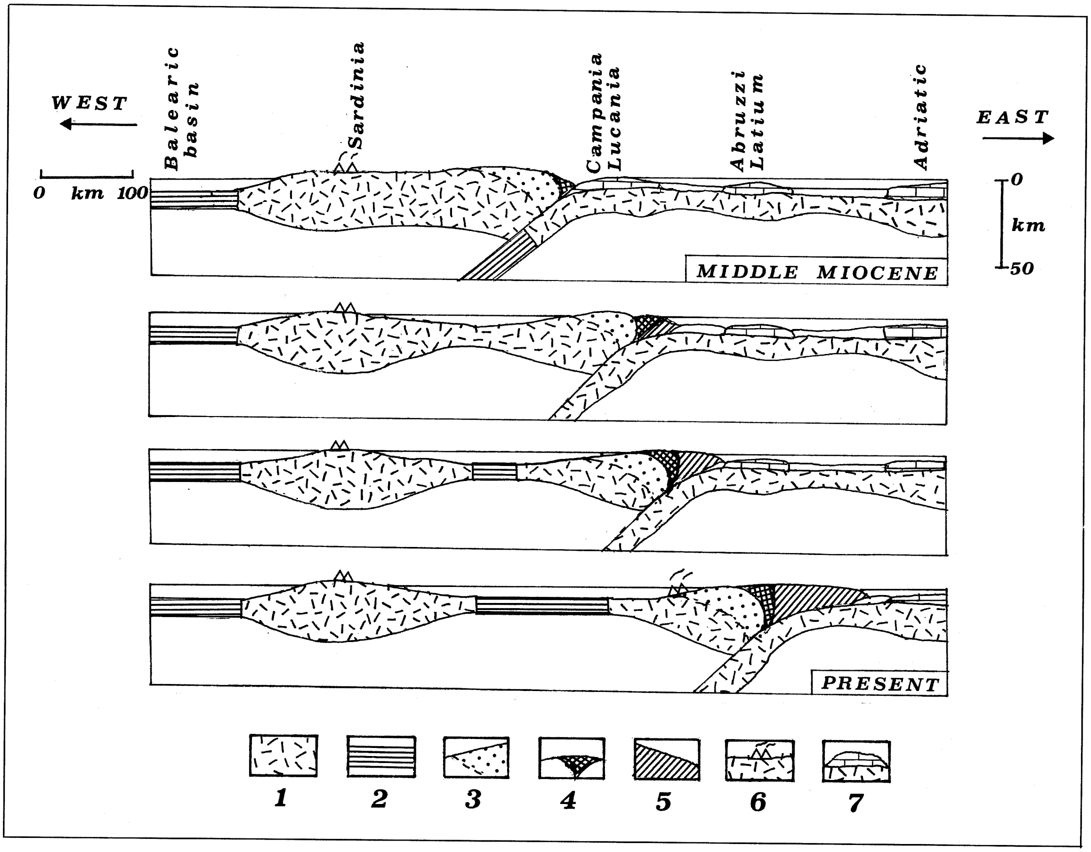


Fig. 7. Scheme of a slab-pull mechanism applied to the interpretation of the Tyrrhenian-Apennines evolutionary pattern, along an E-W cross-section, as proposed by Malinverno and Ryan (1986, modified). 1 = Continental crust; 2 = oceanic crust; 3 = Alpine units of Calabria; 4 = innermost oceanic units of Apennines; 5 = Apenninic accretionary wedge; 6 = active calc-alkaline volcanos; 7 = carbonatic platforms. The downward pull of the deepest slab causes trench retreat (roll-back). This mechanism is supposed to induce a sort of «suction» on the overthrusting plate, which produces thrust belt migration and crustal stretching in its wake. See text for comments.

cent Northern Apennines. This feature clearly contrasts with one major implication of the slab-pull model shown in fig. 7.

Furthermore, it is difficult to understand why the gravitational instability (slab-pull) of the Adriatic subducted lithosphere (taken as the driving mechanism of crustal stretching in the Tyrrhenian) would have started in the Northern Tyrrhenian area and why it occurred only there for a relatively long period (roughly 3 My), up to the middle Messinian. The distribution of calc-alkaline magmatic activity and the features of the accretionary belt connected with the Oligo-Miocene formation of the Balearic basin (see, *e.g.*, Biju-Duval *et al.*, 1977; Dercourt *et al.*, 1986; Bellon, 1981; Scandone, 1979) suggest that the consumption of lithosphere which accompanied the opening of this basin built up an extended subducted lithospheric edifice from North Africa to the Corsica-Sardinia zone (see Mantovani *et al.*, 1992, 1996 with references; Babbucci *et al.*, 1997). Starting from this deep structural context, one could expect that gravitational instability of the subducted lithosphere and the associated surface deformations had also developed along the Northern African margin and in the Southern Apenninic arc, where, instead, no clear evidence exists about this kind of process. A plausible explanation for such different behaviours of subducted lithospheric bodies in the various sectors should be tentatively found, if one wants to adopt the slab-pull hypothesis.

6.2. *Central Tyrrhenian (Magnaghi-Vavilov basin)*

Around the middle Messinian, the deformation pattern in the Tyrrhenian-Apennines system underwent important changes (see Babbucci *et al.*, 1997 for references):

– Crustal stretching ceased in the Northern Tyrrhenian and began in the Central Tyrrhenian with a roughly E-W extensional trend.

– A well documented orogenic pulse began in the whole Apenninic belt; the most intense migration and accretion occurred in the Southern Apenninic arc, which also underwent a counter-clockwise rotation with respect to the Northern Apenninic arc.

– In the Northern Apennines, thrusting activity along the external fronts was accompanied by minor extensional tectonics in the internal side.

– The Central Apennines (Latium-Abruzzi platform) underwent minor deformation and block rotation with respect to the Northern and Southern Apennines.

– A compressional regime parallel to the main trend of the Apennines (S-N to SSE-NNW) affected the whole belt, as indicated by the formation of major arcs from the Northern Apennines to the Calabrian arc, characterized by compressional deformation along the external fronts and transtensional tectonics in the internal sides.

We find great difficulty, in general, in providing physically plausible explanations for the time-space distribution of the above events as effects of a slab-pull mechanism. It is not clear why during this phase the slab-pull mechanism (and related surface extension) only affected a limited sector of the African-Adriatic subducted edifice, and why, in contrast with what happened in the Northern Apennines, the adjacent belt was affected by very intense accretion and outward migration. It also seems hard to explain the different rotations of the Northern and Southern Apennines with respect to the central part of the belt. To explain such a peculiar geometry of surface effects one should assume a very complex flexuring pattern of the Adriatic subducted lithosphere, providing a plausible justification for it.

A major problem in explaining the opening of the Central Tyrrhenian as an effect of slab-pull is, in our opinion, the presence of shortening parallel to the main trend of the belt. This feature, in fact, clearly contrasts with the extensional and transtensional deformations which one would expect from the eastward to SEward pulling of the belt, as an effect of a simple trench retreat driven by slab-pull. To better understand this difficulty one can note in fig. 10 of Malinverno and Ryan (1986) that the total length of the tentatively reconstructed Apenninic belt prior to the Tyrrhenian opening can be roughly estimated in 1000 km and that this length is presumed to have approximately doubled during the subsequent evolution. One might expect that such a lengthening had produced large interruptions in the older segments (pre-Messinian) of the belt, which are not observed. On the contrary, the Apenninic belt seems to have undergone a longitudinal shortening (see, *e.g.*, Castellarin and Vai, 1986).

6.3. *Southern Tyrrhenian*

As concerns this last extensional episode (late Pliocene-Pleistocene), a major problem is explaining why intense crustal stretching only occurred in a narrow and elongated NW-SE corridor, the Marsili basin and with an extensional trend (NW-SE) so different from that contemporaneously affecting the central-northern sectors of the Apennines (Sartori and Capozzi, 1997). To interpret the formation of such a basin as a consequence of slab-pull one should suppose that the sector of lithosphere affected by gravitational sinking was confined to a very narrow slice of the Ionian foreland, which should have undergone a very different kinematic behaviour with respect to other adjacent parts of the Ionian area. However no clear evidence supports this required decoupling.

Around the early-middle Pleistocene, the Calabria-Peloritani block started being affected by a fast uplift, a significant acceleration of its SEward bowing and the activation of several significant transversal discontinuities, such as the Messina, Catanzaro and Sibari sphenocasms and of longitudinal troughs, like the Mesima and Crati grabens. This deformation pattern can hardly be reconciled with the slab-pull of the Ionian slab roughly towards SE, which should not imply a strong disruption of the overthrusting Calabrian block. Even if one manages to find plausible answers to these problems, it remains to be explained why a comparable uplifting rate did not also occur earlier (late Messinian-middle Pliocene) in the Southern Apennines, during the formation of the Magnaghi-Vavilov basin, which is presumed to be an effect of slab-pull too.

We encounter further difficulties when we try to explain the Plio-Quaternary deformation pattern in Sicily and surrounding zones. For example, the extensional episode in the Sicily channel can hardly be explained as a consequence of slab-pull for a number of reasons concerning its particular location, geometry and timing. Clear geological and geophysical evidence indicates that the retreat of the Adriatic/Ionian foreland has occurred since the late Miocene in a wide arc, running from the Southern Apennines to the Calabrian arc. It seems, thus, very difficult to explain why this mechanism would have provoked extensional effects, such as those observed in the Sicily channel, only in very limited sectors of the foreland area and for that time interval.

Some authors (Westaway, 1993; Cinque *et al.*, 1993) suggested that the Quaternary uplift in Southern Italy might represent the isostatic response of the shallow structure to the detachment of the underlying slab. However, this hypothesis does not take into account the presumable significant difference between the deep structural setting beneath the Southern Apennines, where no deep/intermediate earthquakes are recorded and no cold lithospheric material is evidenced by tomographic studies, and the one beneath the Calabrian arc, where deep/intermediate shocks and a cold lithospheric body are clearly indicated by seismological investigations (see, *e.g.*, Anderson and Jackson, 1987; Spakman, 1990; Giardini and Velonà, 1991; Amato *et al.*, 1993; Frepoli, 1995). This evidence seems to indicate that the slab detachment is well developed beneath the Southern Apennines, but the same hypothesis cannot be advanced for the slab beneath the Calabrian arc, where the distribution of deep foci seems to rule out large gaps in the subducted lithosphere. Thus it is difficult to believe that such different deep tectonic behaviours in the above two regions are responsible for the observed uplift, which instead presents very similar features and timing in Calabria and the Southern Apennines (Ghisetti and Vezzani, 1982; Ciaranfi *et al.*, 1983; Westaway, 1993). Furthermore, this hypothesis does not take into account the downward «suction» that the sinking of the detached slab would exert on the overlying crustal structures. A numerical simulation of this phenomenon, carried out by finite element modeling (Giunchi *et al.*, 1996) indicates that in the zone overlying the detached slab no uplift is expected, at least until the lithospheric body has sunk into the mantle at a considerable depth.

6.4. *Aegean system*

Geological and seismological evidence led several authors to suppose that the orientation of the subduction process beneath the Aegean region is roughly SW-NE and that this consuming pattern developed at least since the late Miocene (see, *e.g.*, Le Pichon and Angelier, 1979).

Given this geometry and assuming that the driving force of surface deformation in the Aegean zone was the slab-pull of the subducted lithosphere, with a mechanism like the one shown in fig. 7, one would expect a SW-NE oriented tensional regime in the inner Aegean region, with a time pattern compatible with the thrust belt-foredeep outward migration in the Hellenic arc. Instead, the space-time distribution of extensional tectonics in the Aegean region was much more complex and seems to be hardly compatible with the above prediction. First of all, crustal stretching mainly occurred in two limited zones only, the Cretan basin and the North Aegean trough (figs. 1 and 6), relatively far one from the other (Le Pichon and Angelier, 1979; Angelier *et al.*, 1982). Explaining this peculiar discontinuous distribution of extension as an effect of slab-pull seems to be too hard a task and thus we only try to use this mechanism to account for the extensional behaviour in the Southern Aegean area. This attempt, however, faces a major difficulty, given by the fact that location, rate and trend of extension in the Southern Aegean underwent a considerable change around the late Pliocene-early Pleistocene (see Mercier *et al.*, 1989). From the late Miocene to the late Pliocene the most intense crustal stretching with a roughly S-N orientation, occurred in the Central-Western Cretan basin, whereas since the late Pliocene-early Pleistocene, the most intense extension occurred more to the east with a NW-SE trend, leading to the formation of the Eastern Cretan basin and the Karpathos trough (Buttner and Kowalczyk, 1978; Meulenkamp *et al.*, 1994). It is difficult to interpret this peculiar time-space distribution of extensional features as an effect of a phenomenon, *i.e.*, the retreat of the Western Hellenic trench, which does not seem to have undergone important changes during the last 5-6 My. As concerns the recent/present tectonic setting, the lack of significant recent deformation and the very low seismicity of the Southern Aegean area (Angelier *et al.*, 1982; Jackson *et al.*, 1992) would imply a very low rate of trench retreat, whereas the relatively high rate of outward migration (3-4 cm/yr) estimated in the Hellenic arc by geodetic observations (see, *e.g.*, Smith *et al.*, 1994; Noomen *et al.*, 1996; Le Pichon *et al.*, 1995) seems to contradict this prediction. One could also ask why the most intense recent extensional activity has occurred in the inner side of the Eastern Hellenic trench (Pliny and Strabo), where transpressional movements are recognized by geological and seismological observations (see, *e.g.*, Jackson and McKenzie, 1988; Le Pichon and Angelier, 1979). In this case, too, it does not seem easy to explain the connection between the presumed slab-pull mechanism and the geometry of surface extensional and compressional deformations (see Papazachos and Kiratzi, 1996; Viti, 1996). Another feature which can hardly be explained as a consequence of the SWward roll-back of the Ionian slab is the occurrence, since the early Pleistocene, of E-W extensional tectonics in the Kithira trough, between the Peloponnesus and Crete (fig. 6).

As argued above, the analysis of the Tyrrhenian and Aegean deformation patterns led us to think that the slab-pull of subducted lithosphere can hardly be considered as the main mechanism responsible for the evolution of the above basins and the surrounding zones. However, in this case, one should try to understand why the gravitational sinking of the subducted lithosphere, which is believed to have contributed to the start of consuming process in the above zones, as proposed in figs. 3a-c and 5a-d, does not seem to have evolved in the way suggested by fig. 7, *i.e.*, becoming a self-sustaining process controlling surface deformations.

A possible solution to this problem might be found by taking into account the weakening that the subducted lithosphere is expected to undergo, in the zone of maximum bending (fig. 4a-c) where the concentration of very high tensional and compressional stresses (greater than 1000 MPa, Turcotte *et al.*, 1978; Chapple and Forsyth, 1979) may cause non-elastic deformations in the upper and lower parts of the slab's knee (see also Burov and Diament, 1995; Ranalli, 1994). Since the amount of plastic deformations in the slab's knee depends on its curvature, one could expect that the weakening of the slab is more pronounced in passive subduction processes, like the Tyrrhenian and the Aegean, where this effect is more pronounced. It is also reasonable to suppose that the bending moment, and thus the weakening of the slab, increases as its length becomes greater and greater. Given these premises, one could suppose that in the Mediterranean area the intensity of slab-pull forces has never reached the value necessary to produce significant effects on the kinematics and deformation of the shallow structures, since this was always preceded by the break-off of the slab in its weakened knee zone. The slab break-off may be favoured, in particular, in the trench zones where the consumption of thinned lithosphere is followed by the arrival of thick light continental crust. This situation of opposing buoyancy forces at depth creates strong extensional forces within the slab, resulting in a narrow mode of rifting and ultimately tearing off and falling away of the subducted lithosphere (von Blanckenburg and Davies, 1995). The above interpretation could help to explain, for example, why beneath the Southern Apennines no evidence of subcrustal earthquakes or of cold lithosphere at shallow subcrustal depths is found, in spite of the fact that less than 2 My ago the same region was involved in a roll-back process (see, *e.g.*, Patacca *et al.*, 1993).

7. Conclusions

It is argued that horizontal forces, induced by the convergence between Africa-Arabia and Eurasia, mainly controlled the observed deformations in the Mediterranean area. Gravitational sinking of subducted lithosphere (roll-back) only occurs after the collision between two plates (controlled by horizontal forces) has caused delamination, *i.e.*, the decoupling between the upper buoyant crust and the gravitationally instable lower lithosphere.

This latter process creates the so called «orogenic zones», *i.e.*, low coherent, buoyant and termally weakened crustal edifices which reach a more or less direct contact with asthenospheric material (fig. 3a-c). This kind of zones, when involved in compressional contexts do not oppose strong resistance to flowing laterally, by the escape of crustal wedges, allowing the maximum compressional stresses to shift from one zone to another of the deforming system.

It is argued that the space-time distribution of consuming processes and lateral escapes in the study area were controlled by the basic physical requirement of minimizing the work of horizontal forces against the forces which resist shortening, like crustal strength and gravity. This interpretation can provide plausible and coherent explanations for the time-space distribution of major events and, in particular, for the relatively fast reorganizations of tectonic processes which occurred in the study area around the middle-upper Miocene, the late Miocene and the late Pliocene-early Pleistocene. All these tectonic changes occurred just after the cessation of important consuming processes. Thus, we believe that the subsequent activations of new shortening processes and the changes of microplate kinematics were determined by the redistribution of stresses which followed the sutures of consuming boundaries.

It is finally argued that some major features of the deformation pattern in the study area can hardly be reconciled with the main implications of a slab-pull mechanism. This might indicate that the gravitational sinking of subducted lithosphere does not necessarily produce trench retreat and the associated surface deformation. This possibility could be connected with the weakening that the slab is expected to undergo in the zone of maximum bending, especially in passive consuming boundaries. Because of this weakening, the slab-pull force could be accommodated by the detachment of the subducted body rather than by the roll-back of the upper lithosphere.

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