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Tentative list of major deformation events in the Central-Eastern Mediterranean region since the middle Miocene

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Abstract
A list of major constraints to impose on evolutionary reconstructions of the Central-Eastern Mediterranean region for the middle Miocene period to the Present is proposed. Each constraint is constituted by a tentative description of the deformation and related tectonic regime which affected major active zones during each evolutionary phase. This information has been derived from the analysis of the available observations in the various branches of Earth Sciences, trying to select clearly recognized deformations, possibly supported by independent observations. The list here reported has been used to constrain the evolutionary reconstruction proposed by Mantovani *et al.* (1997).

Key words *Mediterranean – major deformations – geodynamic constraints*

1. Introduction

A huge amount of information is now available on the present and past deformations in the Mediterranean area (fig. 1a,b), derived from observations in all branches of the Earth Sciences. However, only limited parts of this information are generally used in geodynamic reconstructions. This is probably due to the fact that experts in a given field tend to focus attention and rely on the kind of information they know best, both concerning the quality of data and the underlying geodynamic implications, and to consider the tectonic contexts they are most familiar with.

This under-exploitation of the available data may imply an insufficient constraining power in the geodynamic modeling, which could explain why rather different hypotheses are all presented as satisfactory solutions (see, *e.g.*, Lavecchia, 1988; Serri *et al.*, 1991; Van Dijk and Okkes, 1991; Mantovani *et al.*, 1992, for extended reviews of the geodynamic models so far proposed).

To mitigate this problem, experts in different fields should join forces to compile a list of major pieces of evidence and of the related geodynamic implications. This, however, would only represent the first step of the work, since a lot would remain to be done to turn the set of evidence into a number of major constraints to impose on the kinematic pattern and the underlying dynamic framework.

In order to stimulate contributions in this direction, this work reports a preliminary attempt to elaborate a list of the major deformations which have characterized the evolution of the Central-Eastern Mediterranean region since the middle Miocene.

Most of the «points» reported in the list concern information on the kind of deformation, the associated stress field and possibly the deformation rate which is supposed to have af-

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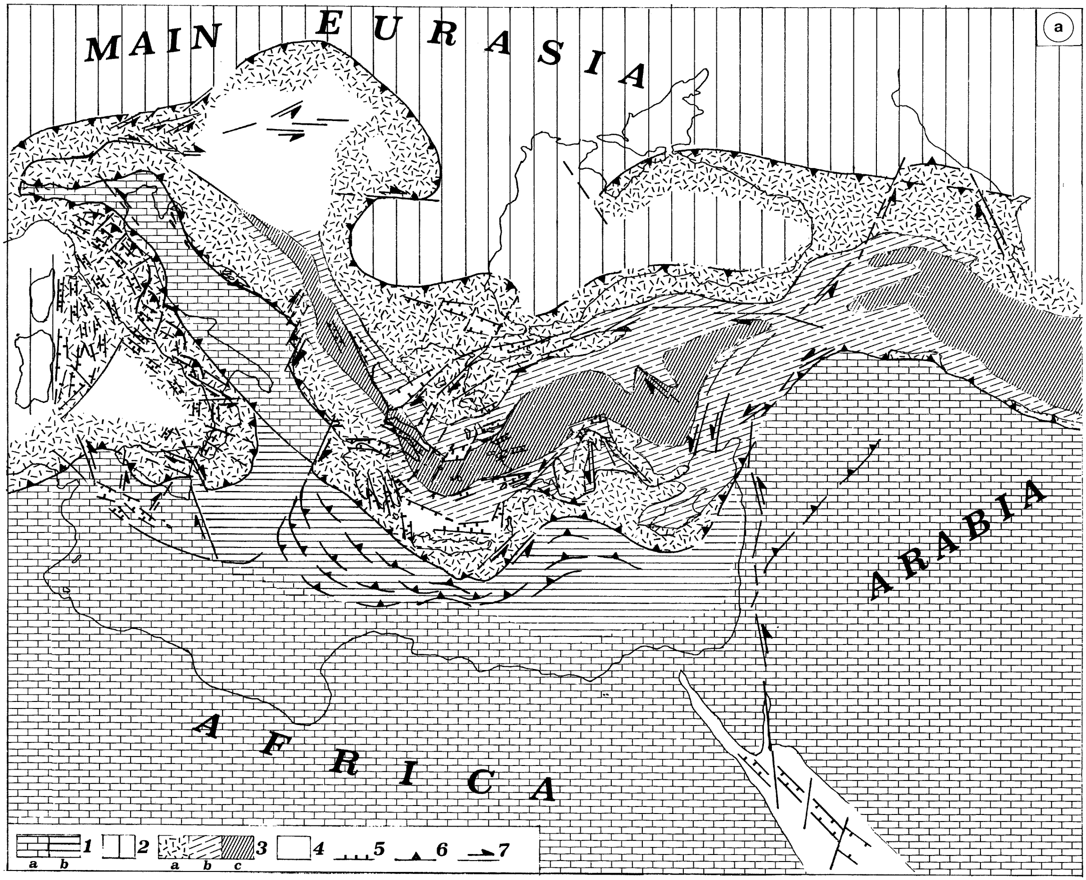
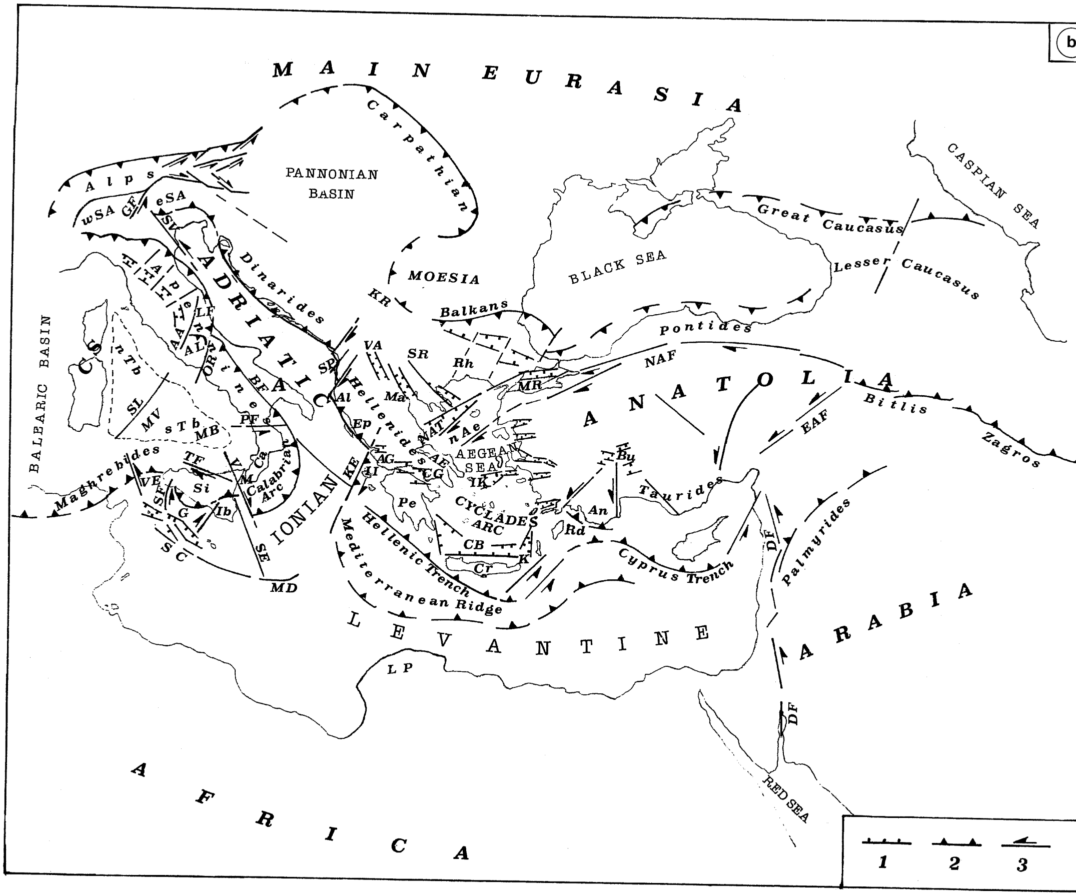


Fig. 1a,b. a) Tectonic sketch of the Central Mediterranean area. 1) African-Adriatic domain with continental (*a*) and thinned crust (*b*). 2) Eurasia domain. 3) Deformation belt: *a*) Eurasian and African deformed margins; *b*) outcropping oceanic remnants (flysch, ophiolites, etc.); *c*) internal massifs (from Boccaletti and Dainelli, 1982). 4) Zones affected by Neogene-Quaternary tensional activity. 5,6,7) Main tensional, compressional and transcurrent features. b) Main tectonic features and geographical references. 1,2,3) Main tensional, compressional and transcurrent features. A = Apulia region; AA = Ancona-Anzio line; AE = Attica-Eubea basin; AG = Ambracique Gulf; AL = Abruzzi-Latium platform; Al = Albanides chain; An = Antalya; BF = Bradanic Foredeep; Bu = Burdur zone; Ca = Calabria; CB = Cretan basin; CG = Corinthian Gulf; Cr = Crete; CS = Corsica-Sardinia massif; DF = Dead Sea transcurrent fault system; EAF = East Anatolian fault system; Ep = Epirus;

fected a given zone during a certain period. Some of the points report interpretations of integrated sets of data which seems to be widely recognized in the relevant literature.

It is reasonable to expect that this list is influenced by our cultural background and by our point of view on Mediterranean geodynamics. We tried to attenuate the effects of personal tendencies by discussing the possible geodynamic implications of each piece of evidence with recognized experts on the related fields.

In some cases, however, personal choices could not be avoided, as, for example, when we encountered very different interpretations.



eSA = eastern Southern Alps; G = Gela nappes; GF = Giudicarie fault system; Ib = Iblean Mounts; II = Ionian Islands; IK = Ikaria basin; K = Karpathos basin; KE = Kefallinia fault; KR = Kraishtides area; LF = Laga flysch; LP = Libyan Promontory; M = Messina sphenocasm; Ma = Macedonia; MB = Marsili basin; MD = Medina Mounts; MV = Magnaghi and Vavilov basins; MR = Marmara basin; nAe = northern Aegean region; NAF = North Anatolian fault system; NAT = Northern Aegean trough; nTb = northwestern Tyrrhenian basin; OR = Ortona-Roccamonfina line; Pe = Peloponnesus; PF = Palinuro fault; Rd = Rhodes; Rh =Rhodope; SC = Sicily Channel; SF = Sciacca fault; SE = Syracuse Escarpment; Si = Sicily; SL = Selli Line; SP = Scutari-Pec line; SR = Struma fault system; sTb = south Tyrrhenian basin; SV = Schio-Vicenza line; TF = Taormina fault system; VE = Ventura plateau; VA = Vardar fault system; V = Vulcano fault; wSA = western Southern Alps.

Sometimes we chose to adopt the most recent or better documented interpretations, at other times we preferred the interpretations which seem to be more coherent with the surrounding deformation pattern.

The timing of the observed deformations is often the object of debate and thus the uncertainty on this parameter may be greater than that on the nature of the related tectonic event. In this case, we attributed different weights to the above two aspects of the same phenomenon.

In this first attempt we only considered the central and eastern parts of the Mediterranean area, since, after the Oligo-Miocenic opening of the Balearic basin, this zone was the scenario of the most intense tectonic activity. Furthermore, as argued in Mantovani *et al.* (1997), we believe that the tectonic processes in the Central and Eastern Mediterranean sectors were closely connected with each other, whereas the minor post-middle Miocene tectonic activity in the Western Mediterranean did not have a significant influence on the deformation pattern of the remaining Mediterranean area.

The list proposed in this work constitutes the data basis of the geodynamic reconstruction described by Mantovani *et al.* (1997). The events are separated into three evolutionary phases (as proposed by Mantovani *et al.*, 1997) and for the Central and Eastern Mediterranean sectors.

As concerns the presumed structural and tectonic setting prior to the evolutionary phases here considered, we make reference to Mantovani *et al.* (1997).

It is highly probable that important constraints are not included in the list here proposed. This may be due to the fact that they are not available to us or that we arbitrarily underestimated their importance. Let us hope that more and more complete and reliable compilations will follow this first attempt.

Previous attempts to recognize the main recent deformations in more or less limited sectors of the Mediterranean area are given by a number of authors (see, *e.g.*, Channell and Horvath, 1976; Biju-Duval *et al.*, 1977; Dewey and Sengor, 1979; Burchfiel, 1980; Boccaletti and Dainelli, 1982; Dercourt *et al.*, 1986; Dewey *et al.*, 1986; Ricou *et al.*, 1986; Philip, 1987; Lavecchia, 1988; Mercier *et al.*, 1989; Patacca *et al.*, 1990; Sartori, 1990; Rebai *et al.*, 1992; Mantovani *et al.*, 1992, 1996).

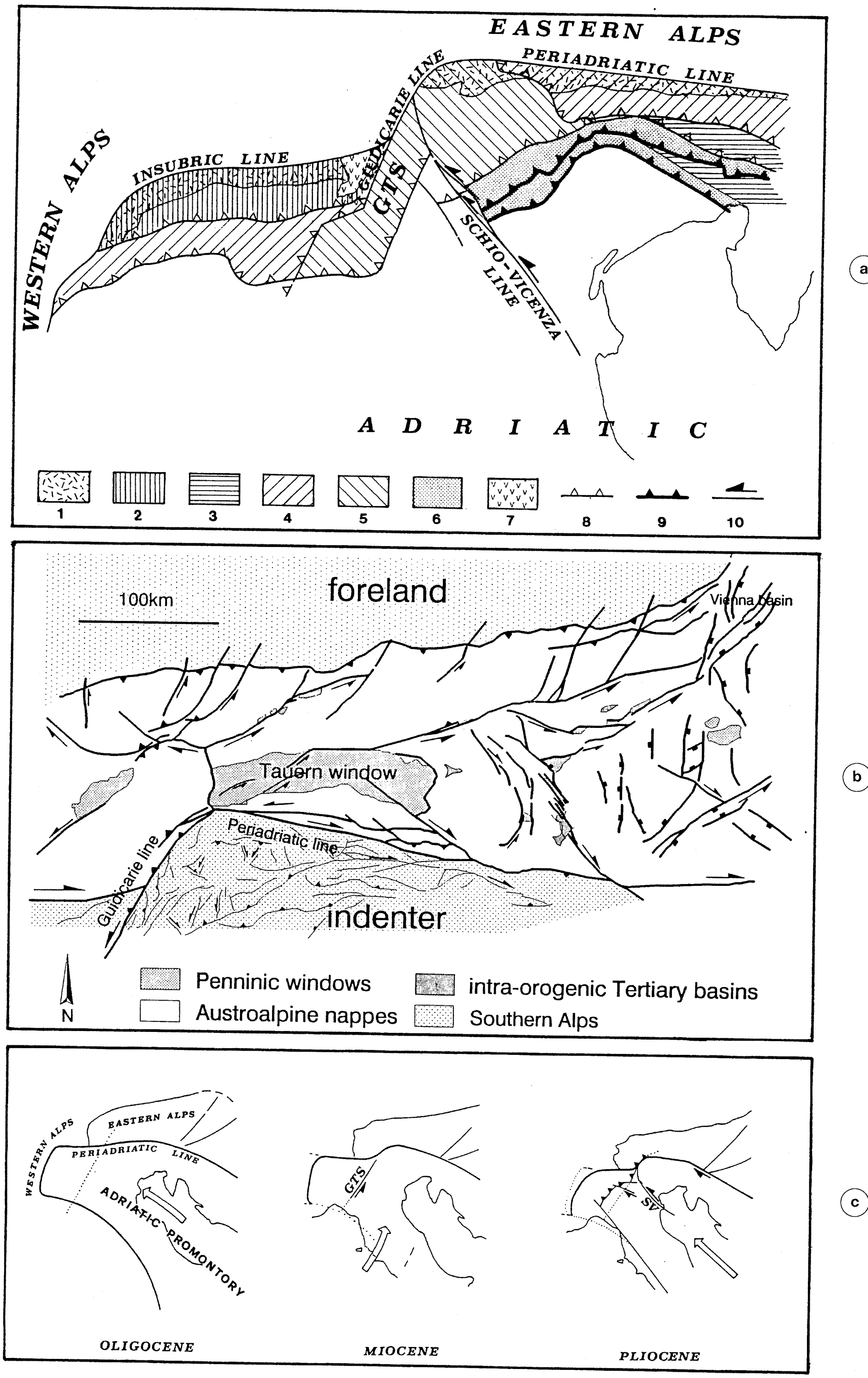
2. Major deformation events

2.1. *First phase: from the middle Miocene to the lower Messinian*

2.1.1. Central Mediterranean

Around the Tortonian, sinistral transpressional activity occurred along the SW-NE trending Giudicarie fault system, see fig. 2a (Semenza, 1974; Castellarin and Vai, 1986; Massari, 1990; Castellarin *et al.*, 1992a), determining a profound change in the deformation pattern of the Southern and Eastern Alps. Since this event, shortening activity mainly affected the Alpine sector lying east of the Giudicarie system, see fig.2a (Castellarin and Vai, 1986; Massari, 1990; Castellarin *et al.*, 1992a) and an acceleration of compressional stresses occurred in the Eastern Alps (Grundman and Morteani, 1985). This last regime was accommodated by crustal thickening, uplift and lateral escape of crustal wedges towards the Pannonian basin (see, *e.g.*, Royden *et al.*, 1983; Laubscher, 1983; Ratschbacher *et al.*, 1991a,b). This lateral flow of masses, driven by sets of conjugate strike-slip faults (fig. 2b), caused tectonic denudation of the zone affected by most intense squeezing and uplift, *i.e.* the present

Fig. 2a-c. Deformation pattern and kinematic interpretation of the Adriatic-Eurasia interaction zone. a) Main structural elements and chronological classification of deformations in the Southern Alps: 1) metamorphic basement rocks; 2) upper Cretaceous deformation; 3) Eocene deformation (Dinaric belt); 4) mainly Chattian-Serravallian deformation; 5) mainly Serravallian-Tortonian deformation, the SW-NE left-lateral transpressional Giudicarie system is identified by GTS; 6) post-Tortonian compressional belt; 7) Paleogene Adamello granitoides; 8) pre-Tortonian thrust fronts; 9) Tortonian to Present thrust fronts; 10) transcurrent faults (after Castellarin and Vai, 1986 and Castellarin *et al.*, 1992a, modified). b) Fault pattern in the Eastern Alps, presumably connected with the lateral escape of crustal wedges (after Ratschbacher *et al.*, 1991a). c) Kinematic pattern of the Adriatic plate proposed by Semenza (1974, modified) to explain the distribution of deformations in the Alps and surrounding regions. GTS = Giudicarie Transpressional System; SV = Schio-Vicenza line. Big arrows indicate the dominant motion of the Adriatic plate with respect to Eurasia.


EASTERN ALPS
PERIADRIATIC LINE
GIUDICARIE LINE
WESTERN ALPS
INSUBRIC LINE
GTS
SCHIO-VICENZA LINE
A D R I A T I C
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foreland
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Giudicarie line
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intra-orogenic Tertiary basins
Southern Alps
b
WESTERN ALPS
EASTERN ALPS
PERIADRIATIC LINE
ADRIATIC PROMONTORY
GTS
SV
OLIGOCENE
MIOCENE
PLIOCENE
c

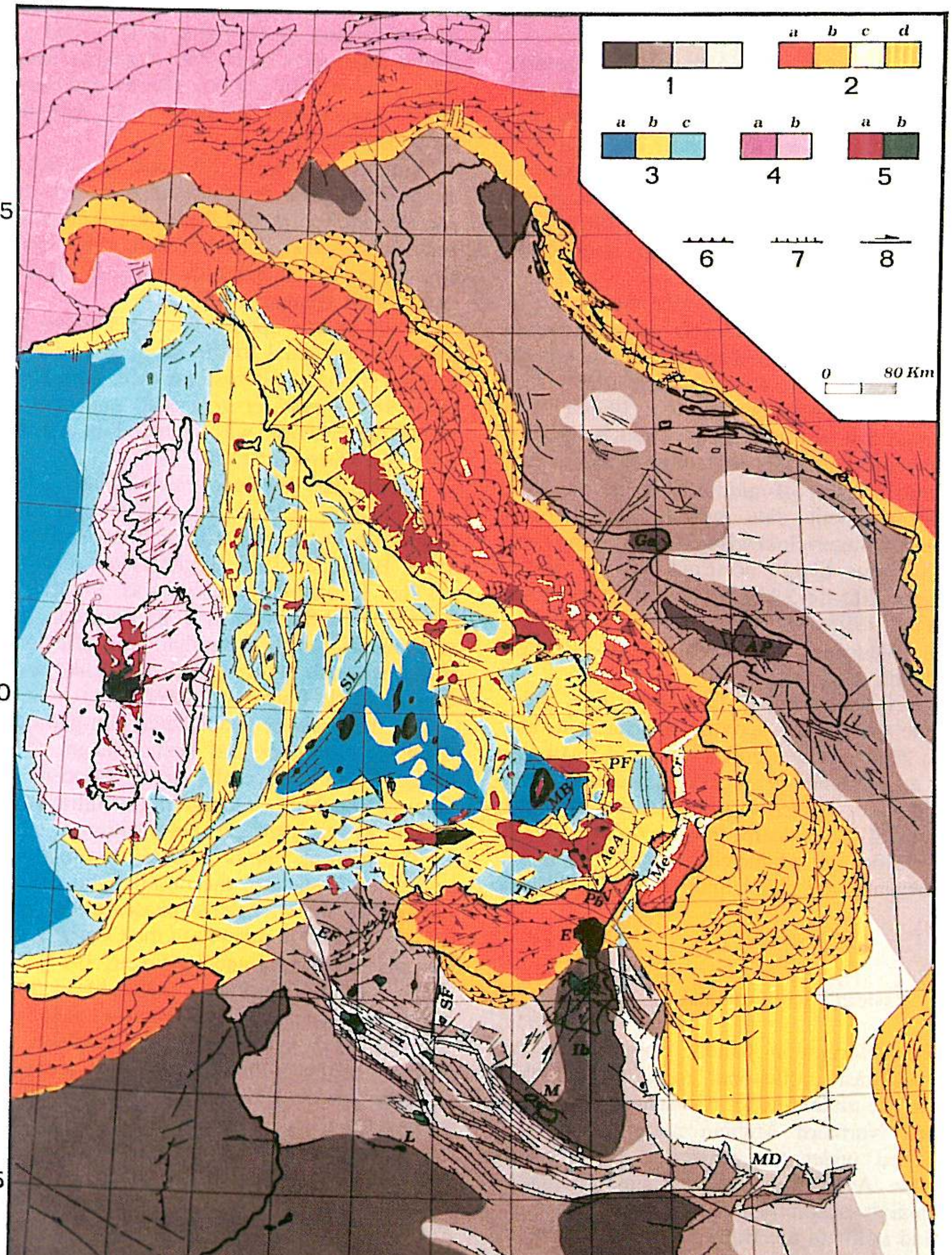
Tauern window. A modeling of this extrusion pattern is given by Ratschbacher *et al.* (1991a,b). The starting of the Giudicarie event was accompanied by strong reactivation of the previous Dinaric structures to the east of the Southern Alps (Castellarin *et al.*, 1992a).

Around the upper Tortonian, intense extensional tectonics, roughly E-W to SW-NE trending, started in a zone previously occupied by both Alpine and Apenninic belts, causing the formation of the Tyrrhenian basin (fig. 3) lying north of the Selli line (Finetti and Del Ben, 1986; Kastens *et al.*, 1988; Sartori, 1990; Mascle and Réhault, 1990). An extensional rate of roughly 2 cm/yr is estimated by Sartori and Capozzi (1997). After this phase, the Selli line separated a deep basin to the NW from a more or less emerged land to the SE (Fabbri and Curzi, 1979; Sartori, 1990). This extension was also accompanied by magmatic activity during the upper Miocene (7.5-6 My, see Juteau *et al.*, 1984; Barbieri *et al.*, 1986; Serri *et al.*, 1991).

Some features generated in the previous evolutionary phases (Oligocene and lower-middle Miocene), such as the Corsica basin and other small troughs along the Eastern Sardinia margin and the exumation of metamorphic rocks in the Apuane Alps and Elba island, in Tuscany, could be interpreted as evidence of an earlier starting of the extension in the Northern Tyrrhenian sector (Zitellini *et al.*, 1986; Carmignani and Kligfield, 1991). However, Sartori and Capozzi (1997) suggested that the same features can be explained as second order effects of a compressional setting: the Corsica basin was passively transported on top of the Alpine-preTortonian Apenninic belt during the rotation of Corsica and of the adjacent chains; the exumation of metamorphic rocks in the Apuane Alps and Elba island may be related to the occurrence of extension on top of an overthickened accretionary wedge (Platt, 1986); the small preTortonian basin in the Eastern Sardinia margin may be related to transpressional-transtensional deformations during the rotation of Sardinia.

A relative tectonic quiescence characterized the Apenninic belt from roughly the Tortonian to the intra-Messinian, as indicated by the continuous subsidence (Sahelian Cycle) in the whole belt and the adjacent foredeep (Parotto and Praturlon, 1975; Di Nocera *et al.*, 1976; Castellarin and Vai, 1986; Vai, 1987; Sgrosso *et al.*, 1988; Borsetti *et al.*, 1990; Sartori, 1990; Boccaletti *et al.*, 1992; Castellarin *et al.*, 1992a). Prior to the Messinian, the Apenninic belt-foredeep system was characterized by a

Fig. 3. Morphological and tectonic map of the Central Mediterranean area. 1) Africa-Apulia foreland: the different tones correspond to bathymetry or altitude, the darkest tone indicates the relative highs; the lightest tone corresponds to the thinned Ionian foreland. 2) Main orogenic belts: orange (*a*) = relative highs; dark yellow (*b*) = external sectors of the belts, mostly buried beneath the periAdriatic foredeeps and the Ionian sea; light yellow (*c*) = main intraorogenic troughs; striped yellow (*d*) = «cobblestone area» (see Rossi and Sartori, 1981). 3) Extensional area: dark blue (*a*) = zones affected by intense crustal stretching, with exposure of oceanic crust; yellow (*b*) = highs mainly constituted by remnants of Alpine and preMessinian Apenninic belts and by fragments of continental margins; light blue (*c*) = zones affected by moderate crustal stretching. 4) Eurasian domain: dark violet (*a*) = Alpine belt; light violet (*b*) = Corsica-Sardinia block. 5) Magmatic bodies related to subduction (*a*) or extensional processes (*b*). 6,7,8) Main compressional, extensional and transcurrent features. AeA = Aeolian Arc; AP = Apulia zone; C = Catanzaro trought; Cr = Crati graben; E = Etna volcano; EF = Egadi fault system; Ga = Gargano high; Ib = Iblean Plateau; L = Lampedusa island; M = Malta island; MB = Marsili basin; MD = Medina seamonts; Me = Mesima graben; Pb = Peloritani block; PF = Palinuro fault; SF = Sciacca fault; SL = Selli Line; TF = Taormina fault system; V = Vulcano fault. (After Funiciello *et al.*, 1976; Boccaletti *et al.*, 1985; Finetti and Del Ben, 1986; Ambrosetti *et al.*, 1987; Bigi *et al.*, 1989; Sartori *et al.*, 1989; Slejko *et al.*, 1989; Turco *et al.*, 1990; Grasso and Ben Avraham, 1992; Cinque *et al.*, 1993; Monaco, 1993; Carmignani *et al.*, 1994; Sacchi *et al.*, 1994).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more or less straight shape roughly trending 320°N (Meloni *et al.*, 1997).

A feature of the Tortonian structural setting which significantly influenced the subsequent evolution of the Central Mediterranean was the presence of a thinned lithosphere in a relatively narrow zone, the Ionian, between the African and Adriatic continental areas (see, *e.g.*, Rossi and Sartori, 1981; Scandone *et al.*, 1981; Dercourt *et al.*, 1986; Malinverno and Ryan, 1986).

The African foreland in the future Sicily channel zone was still closely connected with Africa and was affected by SW-NE compressional stresses (see, *e.g.*, Reuther, 1987; Boccaletti *et al.*, 1987). In particular, Illies (1981) suggested that «an up-arching of roughly 200 m has preceded the physiographic rifting which occurred in the post-middle Miocene times».

Paleomagnetic data indicate that the last phase of counter-clockwise rotation of the Corsica-Sardinia block occurred in the lower Miocene, 20.5-18.5 My ago (Vigliotti and Langenheim, 1995; Meloni *et al.*, 1997). By the middle Miocene, this block did not undergo significant rotations.

The distribution of Oligo-Miocenic calc-alkaline magmatic activity in Sardinia and North Africa, which ended around 13 My ago (see, *e.g.*, Savelli *et al.*, 1979; Bellon, 1981; Beccaluva *et al.*, 1994), suggests that after the opening of the Balearic basin an extended arcuate edifice of subducted lithosphere was lying beneath the Apenninic and Maghrebian belts.

2.1.2. Eastern Mediterranean

The deformation pattern observed at the Arabia-Eurasia collisional zone suggests that around the middle-upper Miocene, most of the thinned Northern Arabian margin had been consumed under the Anatolian belts. Since then the Arabia-Eurasia continental collision was mainly accommodated by crustal thickening and uplift in the Bitlis-Zagros suture zone, that produced widespread deposition of late Miocene-Pliocene continental clastics, foreland type folding and thrusting in previously deformed parts of the Arabian platform, reactivation of folding in the Palmyra zone and a regional regression throughout Eastern Anatolia (Dewey *et al.*, 1973, 1986; Dewey and Sengor, 1979; Sengor and Yilmaz, 1981; Hempton, 1985; Barazangi *et al.*, 1993).

Around the middle Miocene spreading activity in the Red Sea-Gulf of Aden and transcurrent motion at the Dead Sea fault system came to an end (Zak and Freund, 1981; Girdler and Styles, 1982; Schmidt *et al.*, 1983; Coleman, 1985; Hempton, 1987).

In the middle-upper Miocene, Eastern Anatolia was bounded to the north and south by transpressional fault systems (see, *e.g.*, Hempton, 1982; Finetti *et al.*, 1988) and compressional deformation was recorded in the Pontides belt (Dercourt *et al.*, 1986).

During the Miocene calc-alkaline activity took place in the Caucasus region, continuing locally up to the Quaternary (Innocenti *et al.*, 1982; Hempton, 1987; Philip *et al.*, 1989).

The Rhodopian massif was involved in large NNE-SSW sinistral transcurrent faults and the adjacent Balkan chain was affected by compressional deformation (Stanishkova and Slejko, 1991; Dabovsky, 1991).

On the basis of geophysical and geological evidence, Kempler and Ben-Avraham (1987) suggested that since the upper Miocene, the Cyprus trench has undergone a southward migration to reach the present position.

The present Taurus mountains, which were under sea level until the middle Miocene, underwent in the successive phases a strong uplift, up to 1000 m relative to the Central Anatolian plateau (Barka and Reilinger, 1997).

By the late Miocene, a noticeable deformation took place in the Antalya zone (SW Turkey), with extension at the northern side (Burdur region), thrusting activity along the external border with the Levantine basin and transpressional activity and uplift at the western and eastern lateral borders (Eyidogan, 1988; Marcoux *et al.*, 1989; Ben-Avraham and Ginzburg, 1990; Barka, 1992; Taymaz and Price, 1992).

Reconstructions of the upper Miocene shape of the Hellenic arc suggest that it was almost straight, roughly E-W trending (McKenzie, 1972; Le Pichon and Angelier, 1979, 1981; Kissel *et al.*, 1985; Sorel and Mercier, 1988) and that it was affected by a SSW-NNE compression (Dercourt *et al.*, 1986; Ricou *et al.*, 1986).

In the upper Miocene the Northern Aegean zone was affected by a tensional tectonics roughly trending WNW-ESE (Mercier *et al.*, 1989). Contemporaneously, a roughly N-S extension occurred in the Tessaly and Macedonia regions (Caputo and Pavlides, 1993) and the Western Anatolia (Zanchi *et al.*, 1990; Barka, 1992).

Around the early Tortonian the South Aegean landmass broke up and a shallow marine basin began to form north of Crete in the middle Tortonian (Meulenkamp *et al.*, 1977, 1994; Angelier *et al.*, 1982; Lyberis *et al.*, 1982; Mercier *et al.*, 1989).

Some authors suggested that the westward displacement of the Anatolian landmass played an important role in the buckling of the Hellenic arc and in the extensional deformation of the Aegean basin (McKenzie, 1972; Brunn, 1976; Tapponnier, 1977).

In the upper Miocene, shortening processes occurred along the Eastern Adriatic border (see, *e.g.*, Channell and Horvath, 1976; Burchfiel, 1980; Horvath, 1984), with particular regard to the outer Hellenides, where the Epirus zone started overthrusting onto the Apulian platform, continuing until the lower Pliocene (see IGSR-IFP, 1966; Aubouin, 1973; Mercier *et al.*, 1979, 1989). Also in the zone between the Eastern Dinarides and the stable Moesia platform (Kraishtides area), faulting and roughly E-W thrusting activity took place (Stanishkova and Slejko, 1991).

2.2. *Second phase: from the middle Messinian to the upper Pliocene*

2.2.1. Central Mediterranean

By the late Messinian, extensional tectonics in the Northwestern Tyrrhenian came to an end (see, *e.g.*, Finetti and Del Ben, 1986; Kastens *et al.*, 1988; Sartori, 1990).

By the middle-late Messinian, a relatively sharp renewal of orogenic activity, with accretion and outward migration took place in the whole Apenninic belt (Elter *et al.*, 1975; Parotto and Praturlon, 1975; Di Nocera *et al.*, 1976; Ortolani, 1979; Pieri and Groppi, 1981; Ghisetti and Vezzani, 1984; Marabini and Vai, 1985; Castellarin and Vai, 1986; Bigi *et al.*, 1989; Patacca and Scandone, 1989; Sartori, 1989, 1990; Castellarin *et al.*, 1992a; Bernini *et al.*, 1992; Bossio *et al.*, 1993). The most intense activity affected the Southern Apennines, lying south of the Ortona-Roccamonfina line, which also underwent counter-clockwise rotation with respect to the central-northern part of the belt (Di Nocera *et al.*, 1976; Ortolani, 1979; Casnedi *et al.*, 1982; Incoronato and Nardi, 1989; Sartori, 1989; Patacca and Scandone, 1989). This southern part of the chain displays a duplex imbrication style, while the Central-Northern Apennines underwent a more gentle deformation, characterized by a piggy-back imbrication style (Patacca and Scandone, 1989; Bernini *et al.*, 1992; Bossio *et al.*, 1993).

Around the Messinian, the Abruzzi-Latium carbonatic platform started being affected by a roughly NEward compressional regime, which caused the progressive closure of all pre-existing SE-NW troughs and the formation of thrust fronts (Castellarin *et al.*, 1978, 1982; Ghisetti and Vezzani, 1991; Mattei *et al.*, 1991; Acocella *et al.*, 1996). The Laga Flysch underwent strong bending and translation (Casnedi *et al.*, 1982; Mattei, 1987). The zone lying just north of the Ancona-Anzio line was affected by a roughly S-N compressional regime which caused torsion of the previous orogenic fronts. Some authors (Calamita and Deiana, 1988; Lavecchia *et al.*, 1988) suggested that during this phase, the central part of this alignment (Olevano-Antrodoco in recent literature) behaved like a lateral ramp for the further eastward migration of the Northern Apenninic units. In post late Miocene, the Central Apennines underwent differential block rotations: counter-clockwise in the northern sector and clockwise in the southern one (Meloni *et al.*, 1997).

Since the Messinian, the evolution of the Central-Northern Apennines has been characterized by the concomitant development of extensional tectonics in the internal zones and compressional deformation along the external fronts (fig. 1a,b). Both types of phenomena presented a progressive migration towards E to NE (see, *e.g.*, Elter *et al.*, 1975; Parotto and Praturlon, 1975; Bartolini *et al.*, 1983; Castellarin and Vai, 1986; Ghisetti and Vezzani, 1991; Mattei *et al.*, 1991; Patacca *et al.*, 1993; Boccaletti *et al.*, 1992; Lavecchia *et al.*, 1994; Bartole, 1995). Paleomagnetic data (Meloni *et al.*, 1997) suggest that the external sector of the Northern Apennines underwent an «oroclinal» bending. Castellarin and Vai (1986) and Castellarin *et al.* (1992b) suggested that the Messinian-Pliocene arc-shaped deformation pattern of the Northern Apennines can be attributed to a compressional regime roughly parallel to the Apenninic trend. This hypothesis is also consistent with the occurrence of transpressional deformations along the major transversal discontinuities, separating the Northern, Central and Southern Apenninic segments *i.e.* the Olevano-Antrodoco (Castellarin *et al.*, 1978; Calamita and Deiana, 1988; Lavecchia *et al.*, 1988) and Ortona-Roccamonfina (see, *e.g.*, Locardi, 1988; Patacca and Scandone, 1989; Cavinato *et al.*, 1994 and references therein).

The outward displacement of the Apenninic units occurred at the expense of the adjacent Adriatic foreland which underwent downward flexure beneath the advancing belt, as suggested by geological analysis and by a high number of exploration wells in the Bradanic foredeep (Casnedi *et al.*, 1982; Mostardini and Merlini, 1986; Patacca and Scandone, 1989; Patacca *et al.*, 1990, 1993).

The most evident eastward allochtonous character of the Southern Apenninic units and the development of the adjacent foredeep seem to be mostly confined to the region lying south of the Ortona-Roccamonfina (OR) line (Casnedi *et al.*, 1982). Given that the outward migration of the belt had a close relationship with the downward flexure of the Adriatic foreland, one can reasonably suppose that the OR line was reflected deep down by a lithospheric tear fault that allowed the different flexural patterns of the parts of the Adriatic slab which were lying south and north of this fault (Royden *et al.*, 1987; Patacca *et al.*, 1990; Mantovani *et al.*, 1992). The importance of the OR line, as a possible border between the oceanic-like and continental-like character of the Adriatic subducting margin is also stressed by the interpretation of magmatic data (Serri *et al.*, 1991; Francalanci and Manetti, 1994).

During this phase, an average migration rate of the compressional front of 5-6 cm/yr in the Southern Apennines has been estimated by Patacca *et al.* (1990). The hightest rate (8 cm/yr) is estimated for the Messinian.

Intense crustal stretching took place in the Central Tyrrhenian (*i.e.*, the Magnaghi and Vavilov basins) from the Messinian to the middle-upper Pliocene. Emplacement of oceanic crust in this zone essentially occurred from 4 to 2.5 My ago, with extensional rates of about 4 cm/yr (Sartori and Capozzi, 1997). The morphology of the troughs and the analysis of drilling data indicate that the dominant trend of extension was roughly W-E (Kastens *et al.*, 1988; Sartori *et al.*, 1989).

A late Messinian-lower Pliocene magmatic episode, with uprising of material classified as «subduction-related», occurred in the Tyrrhenian area (fig. 4) with an arc-shape distribution (see Kastens *et al.*, 1988; Savelli, 1988; Bigi *et al.*, 1989; Francalanci and Manetti, 1994).

By the Messinian, the shortening pattern in the Southern Alps underwent a significant change, with the activation of a SE-NW sinistral fault, the Schio-Vicenza line (fig. 2a), and of thrust fronts oriented in an almost perpendicular direction (Cavallin *et al.*, 1984; Castellarin *et al.*, 1992a; Cantelli and Castellarin, 1994). After this event, only weak deformation affected the sector of the Alps lying west of this lineament (fig. 2a) whereas the Eastern Southern Alps have been affected by significant shortenings up to the Present (Castellarin *et al.*, 1992a).

Paleomagnetic studies on Plio-Quaternary rocks indicate a 19° clockwise rotation of Apulia (Tozzi *et al.*, 1988). Some objections have been, however, raised by Marton (1993) on the quality of these data.

Since the late Messinian-early Pliocene, Sicily and the surrounding zones (fig. 3)

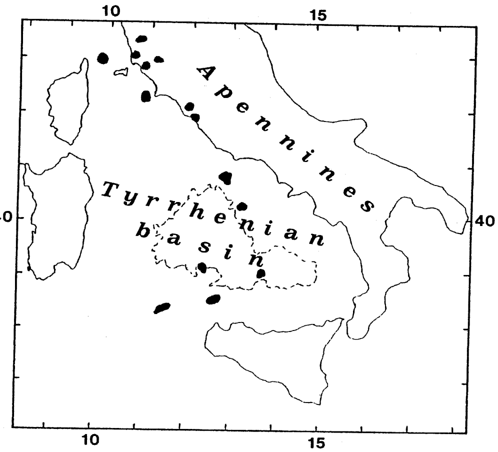

Fig. 4. Subduction-related magmatism (black spots) in the Tyrrhenian area during the Messinian-lower Pliocene (6-4.5 My). The dashed line contours the Tyrrhenian bathial plain (after Sartori, 1986; Bigi *et al.*, 1989).

recorded the start of some significant deformations:

a) The structural border between the Iblean plateau and the Ionian thinned zone, *i.e.* the Syracuse escarpment, was affected by vertical movements of the order of several hundred metres (Carbone *et al.*, 1982; Grasso and Lentini, 1982; Sartori, 1989). The vertical offset created by these movements decreases from north to south (Finetti and Morelli, 1973) and seems to taper out in the area of the Medina Seamounts (Reuther *et al.*, 1993). During this phase, «extensional related» magmatism developed in the Iblean zone (see, *e.g.*, Barberi and Innocenti, 1980).

b) The border between Sicily and Calabria, *i.e.* the Taormina fault system, was characterized by a dextral transpressional regime (see, *e.g.*, Finetti and Del Ben, 1986; Patacca and Scandone, 1989).

c) Compressional deformation trending SW-NE developed in the Ventura plateau and adjacent zones (Bigi *et al.*, 1989; Argnani, 1993). The Gela nappe underwent SWward bending (Ghisetti and Vezzani, 1984; Grasso *et al.*, 1990).

d) During the Pliocene, the Sicily channel was affected by extensional tectonics and strike-slip movements which led to the formation of some troughs (see, *e.g.*, Illies, 1981; Finetti, 1984; Boccaletti *et al.*, 1987; Cello, 1987; Reuther, 1987; Calanchi *et al.*, 1989; Argnani, 1993; Catalano *et al.*, 1994).

e) Paleomagnetic data indicate a clockwise rotation of about 35° of the Caltanissetta basin since the early Pliocene (Scheepers and Langereis, 1993) and a slight (about 10° counterclockwise Plio-Pleistocene rotation of the Iblean foreland (Besse *et al.*, 1984). Meloni *et al.* (1997) suggest that these estimates are affected by significant uncertainties.

2.2.2. Eastern Mediterranean

In the late Miocene-early Pliocene dextral transcurrent activity began along the North Anatolian fault (McKenzie, 1972; Dewey and Sengor, 1979; Barka and Hankock, 1984; Sengor *et al.*, 1985; Dercourt *et al.*, 1986; Hempton, 1987; Barka, 1992).

The timing of activation of the East Anatolian fault is placed by some authors in the late Miocene-early Pliocene (Hempton, 1987; Lyberis *et al.*, 1992 and references therein) and by other authors in the upper Pliocene (see, *e.g.*, Saroglu *et al.*, 1992).

Slip rates along the North and East Anatolian faults estimated by different kinds of evidence (geological, seismological or geodetic data) respectively vary from 5 to 38 mm/yr and from 6 to 31 mm/yr (Barka and Reilinger, 1997).

Since the early Pliocene (4.5 My), a second extensional episode, with severe crustal stretching, affected the Red Sea-Gulf of Aden and transcurrent displacement occurred along the Dead Sea fault system (LaBreque and Zitellini, 1985; Mart and Rabinowitz, 1986; Hempton, 1987).

Only minor deformation affected the Pontides belt after the upper Miocene (Boccaletti and Dainelli, 1982; Dercourt *et al.*, 1986; Finetti *et al.*, 1988).

The orogenic deformation in the Balkans and the transcurrent activity (SW-NE fault system) in the Rhodopian massif slowed down gradually and almost ceased around the end of the Pliocene (Stanishkova and Slejko, 1991).

Western Turkey has undergone a roughly ENE-WSW dextral strike-slip tectonics in the northernmost sector and dominantly N-S extension in the central-southern part, with the formation of E-W trending grabens (Eyidogan, 1988; Yilmaz, 1989; Zanchi *et al.*, 1990).

The Northern Aegean region was affected by dextral transtensional tectonics, with the formation of a system of SW-NE strike-slip faults (Barka and Hancock, 1984; Sengor, 1985; Dewey *et al.*, 1986; Hempton, 1987; Mercier *et al.*, 1989; Taymaz *et al.*, 1991; Barka, 1992).

In the late Messinian-early Pliocene, the Central Aegean metamorphic belt (Cyclades arc) was affected by intense compressional deformations and uplift which determined plurikilometric-sized faults and the first deposition of continental facies, after the Miocene marine sedimentation (Buttner and Kowalczyk, 1978; Durr *et al.*, 1978; Mercier *et al.*, 1987).

Around the late Miocene-early Pliocene crustal stretching occurred in the Central-Western Cretan basin (fig. 1a,b) along with a fast uplift of Crete (Buttner and Kowalczyk, 1978; Meulenkamp *et al.*, 1994).

Since the late Miocene a broad accretionary zone developed along the external border of the Hellenic arc, the so-called «Mediterranean ridge» (Finetti, 1976; Underhill, 1989).

In the late Miocene-early Pliocene, the outer border of the Eastern Hellenic arc (Crete-Rhodes), changed from a convergent to a sinistral transpressional boundary (Jacobshagen *et al.*, 1978; Dewey and Sengor, 1979).

Calc-alkaline volcanic activity took place in the Southern Aegean basin (Fytikas *et al.*, 1984; Papazachos and Panagiotopoulos, 1993), most probably connected with the subduction of the Ionian-Levantine foreland beneath the Hellenic arc (Le Pichon and Angelier, 1979).

Around the late Miocene, the roughly E-W shortening process in the Northern Hellenides (Epirus) underwent a noticeable slowdown-cessation. Since then a relative tectonic quiescence affected this zone until the middle-upper Pleistocene (see, *e.g.*, Mercier *et al.*, 1979, 1989; Auroux *et al.*, 1984).

Paleomagnetic observations (Kissel *et al.*, 1985; Speranza *et al.*, 1995) suggest that since the late Miocene-early Pliocene the Hellenides underwent a clockwise rotation of roughly 25° whereas no rotation is recorded in the Dinarides. The decoupling between these differentiated kinematic behaviours is supposed to be due to the Scutari-Pec discontinuity (see, *e.g.*, Kissel *et al.*, 1985, 1995). This hypothesis is supported by geological and geophysical evidence (Anderson and Jackson, 1987b; Čermák, 1993) and by an abrupt change in the deflection of the Adriatic lithosphere beneath the Eastern Adriatic chains (Moretti and Royden, 1988).

During the late Miocene-early Pliocene strong uplift affected the Albanides (Sulstarova *et al.*, 1980).

Dextral transpressional activity developed along the southern border of the Adriatic platform (Kefallinia fault) since the Pliocene (see, *e.g.*, Sorel, 1989; Mercier *et al.*, 1989), reactivating old normal faults (Sorel *et al.*, 1976). This discontinuity (fig. 1a,b) continues offshore SWward for some 200 km in the Ionian area (Finetti and Del Ben, 1986; Van Dijk and Okkes, 1991; Reuther *et al.*, 1993).

Since the lower Pliocene a reactivation and initiation of reverse faults occurred in western mainland Greece indicating a continent-continent collision (Underhill, 1989).

By the lower Pliocene, the Peloponnesus underwent uplift for several kilometres, spliced up in a number of blocks separated by transcurrent faults (Stiros, 1988) and rotated clockwise for about 25° (Kissel and Lay, 1988). In the southern sector of this block, NNW-SSE extensional tectonics led to the formation of the present gulfs (Kelletat *et al.*, 1978).

By the early Pliocene, extensional tectonics, roughly oriented S-N, led to the formation of troughs, such as the E-W Corinthian Gulf and the Attica-Euboea NW-SE basins (Buttner and Kowalczyk, 1978; Berckhemer and Kowalczyk, 1978; Stiros, 1988).

From the lower Pliocene to the lower Pleistocene, the internal Aegean domain underwent a NNE-SSW extension (Mercier *et al.*, 1987, 1989).

In the Serbo-Macedonian and Vardar zones, a NE-SW trending extensional regime began in the late Miocene, with the activation of the Struma and Vardar tensional features. Plio-Quaternary volcanism with a continental rift character developed along these graben structures (Pavlides and Kilias, 1987; Milev and Vrablyanski, 1988; Dinter and Royden, 1993).

2.3. *Third phase: from the late Pliocene-early Pleistocene to the Present*

2.3.1. Central Mediterranean

Around the late Pliocene-early Pleistocene, the outward migration of the Southern Apennines and the downward flexure of the adjacent Adriatic foreland underwent slowdown/cessation (see, *e.g.*, Ortolani, 1979; Casnedi *et al.*, 1982; Ghisetti and Vezzani, 1982; Patacca and Scandone, 1989; Patacca *et al.*, 1990, 1993; Cinque *et al.*, 1993).

By the late Pliocene, crustal stretching almost ceased in the Magnaghi-Vavilov basin, while a new extensional phase, with a NW-SE trend and emplacement of oceanic basalts began in the Southernmost Tyrrhenian, causing the formation of the narrow Marsili basin, see figs. 1a,b and 3 (Finetti and Del Ben, 1986; Kastens *et al.*, 1988; Sartori, 1989).

Since the late Pliocene-early Pleistocene, some evidence of a compressional regime more or less parallel to the main Apenninic trend have been recorded in the Northern and Southern Apennines (Boccaletti *et al.*, 1992; Sacchi *et al.*, 1994). This regime was accompanied by fast uplift (see, *e.g.*, Boccaletti *et al.*, 1992; Ciaranfi *et al.*, 1983; Sacchi *et al.*, 1994) and by the activation of out of sequence compressional fronts (Castellarin and Vai, 1986; Boccaletti *et al.*, 1992; Cinque *et al.*, 1993).

Since the late Pliocene-early Pleistocene, bowing in the Northern and Central Apennines arcs underwent a significant acceleration, as testified by paleomagnetic observations (Meloni *et al.*, 1997) and by the increase in deformation rates both along the external fronts (see, *e.g.*, Vai, 1987; Sartori, 1989; Boccaletti *et al.*, 1992) and the internal troughs (see, *e.g.*, Bartole, 1995).

Geological analyses on a NS left lateral transcurrent fault system crossing the Olevano-Antrodoco line and the focal mechanisms available in the same area indicate a stress field characterized by a NW-SE compression and a NE-SW extension. This regime has been active since the middle Pleistocene (Cello *et al.*, 1995).

In the Southernmost Apennines, Patacca and Scandone (1989), Turco *et al.* (1990), Monaco (1993) and Cinque *et al.* (1993) suggested the presence of a left-lateral fault system since the late Pliocene-early Pleistocene (figs. 5 and 6a-d). Since the middle Pleistocene, the activity of the NW-SE Cilento-Pollino sinistral strike-slip zone has slowed down considerably (Cinque *et al.*, 1993; Moretti *et al.*, 1994).

Since the early Pleistocene, Calabria and the Southernmost Apennines respectively experienced clockwise (15°) and a counter-clockwise (23°) rotations (Sagnotti, 1992; Scheepers *et al.*, 1994). Paleomagnetic evidence might indicate that Calabria has not rotated as a unique block (Meloni *et al.*, 1997). A fast uplift, with maximum rates of 1.5 mm/yr, occurred in these regions (Ghisetti and Vezzani, 1982; Ciaranfi *et al.*, 1983). This effect was coeval with the occurrence, in the central sector of the Southern Apennines, of normal faults and intramontane basins characterized by an arcuate shape roughly parallel to the external thrust fronts (Cinque *et al.*, 1993). Uplift also affected the Bradanic foredeep and the adjacent Apulian region, after a Pliocene-early Pleistocene subsidence phase (Doglioni *et al.*, 1994).

Since the late Pliocene-early Pleistocene, the Calabria-Peloritani block (figs. 1a,b and 3) was affected by a considerable acceleration of tectonic activity, as indicated by several pieces of evidence:

a) Around 0.9-0.7 My extensional activity in the Calabrian longitudinal troughs (Mesima and Crati grabens in fig. 3) changed from roughly 0.1 mm/yr to 1 mm/yr (Westaway, 1993).

b) Intense compressional deformations affected the external Calabrian arc (see, *e.g.*, Rossi and Sartori, 1981; Barone *et al.*, 1982;

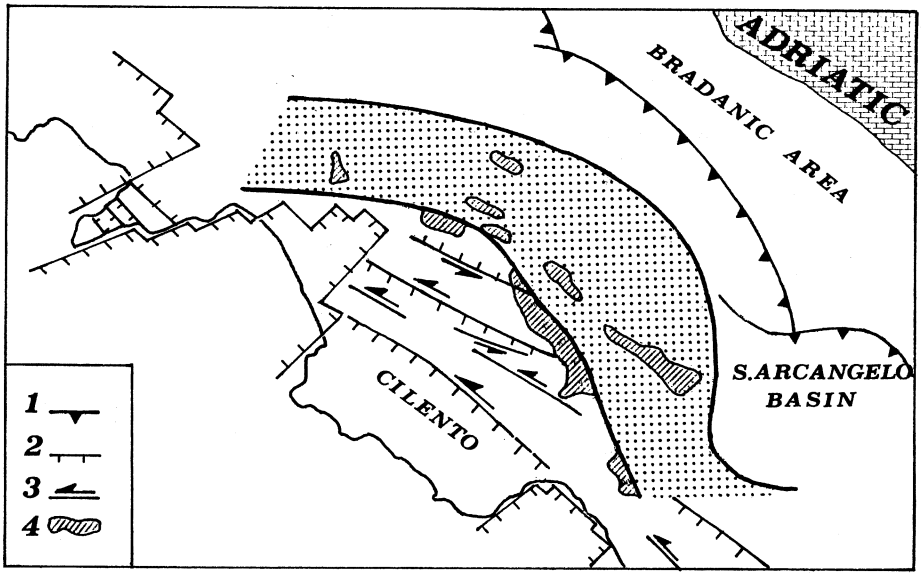

Fig. 5. Example of arc shaped structures in the Southern Apennines. The dotted area represents the Miocenic belt in the Campania-Lucania sector transported and deformed in Pliocene and Pleistocene times and strongly uplifted in early Pleistocene. Transtensional tectonics is evidenced in the inner side of the arc. 1) External front; 2) normal faults; 3) transcurrent faults; 4) main intramontane basins (after Cinque *et al.*, 1993, modified).

Finetti and Del Ben, 1986; Bigi *et al.*, 1989; Sartori, 1989).

c) An important sphenocasm, the Catanzaro trough (fig. 3), developed between the southern and northern parts of Calabria (see, *e.g.*, Del Ben, 1993 and references therein).

d) The Peloritani block (fig. 3) detached from Southern Calabria, by the Messina sphenocasm (see, *e.g.*, Selli *et al.*, 1978; Ghisetti, 1979).

Since the Pleistocene, deformation rates in the northern part of Calabria slowed down considerably. This phenomenon is interpreted as an effect of the collision of this block with the Adriatic continental margin (Barone *et al.*, 1982; Del Ben, 1993). Since this event, most intense tectonic activity affected the southern part of this block, as clearly testified by neotectonic and seismicity data (see, *e.g.*, Barbano *et al.*, 1978; Barone *et al.*, 1982; Finetti and Del Ben, 1986; Van Dijk and Okkes, 1991).

Around the late Pliocene-early Pleistocene, the Taormina fault (fig. 3) became a predominantly compressional border and transpressional activity began at the S-N to SSE-NNW Vulcano and Sciacca faults (see, *e.g.*, Finetti and Del Ben, 1986; Reuther, 1987). Tectonic activity in the Syracuse escarpment and in the Sicily channel slowed down considerably (Finetti, 1984; Reuther, 1987; Cello, 1987; Calanchi *et al.*, 1989). Catalano *et al.* (1994) suggest that since the late Pliocene-early Pleistocene folding tectonics and uplift affected the Malta-Lampedusa region.

A new magmatic episode classified as «subduction related» has taken place since the early Pleistocene in the Central and Southern Apennines, Roman and Neapolitan magmatic provinces, and in the Southernmost Tyrrhenian area, Aeolian Arc (Beccaluva *et al.*, 1982, 1989; Bigi *et al.*, 1989; Serri, 1990; Francalanci and Manetti, 1994).

Petrological studies of Plio-Quaternary volcanic rocks in the Southern Tyrrhenian-Apennines system suggest an oceanic nature of the lithosphere subducted south of the Ortona-Roccamonfina line (see, *e.g.*, Serri *et al.*, 1991; Francalanci and Manetti, 1994). Conversely, the coeval magmatism lying north of this line indicates a major involvement of the continental lithosphere (Serri *et al.*, 1991).

In Eastern Sicily, just in front of the Iblean plateau an extensional-related magmatic episode, with the formation of the Etna volcano (fig. 3) has developed since 0.5 My (Barberi *et al.*, 1974; Lo Giudice *et al.*, 1982; Romano, 1983; Bigi *et al.*, 1989).

In the late Pliocene, the direction of tectonic transport in the Sicilian chain changed from

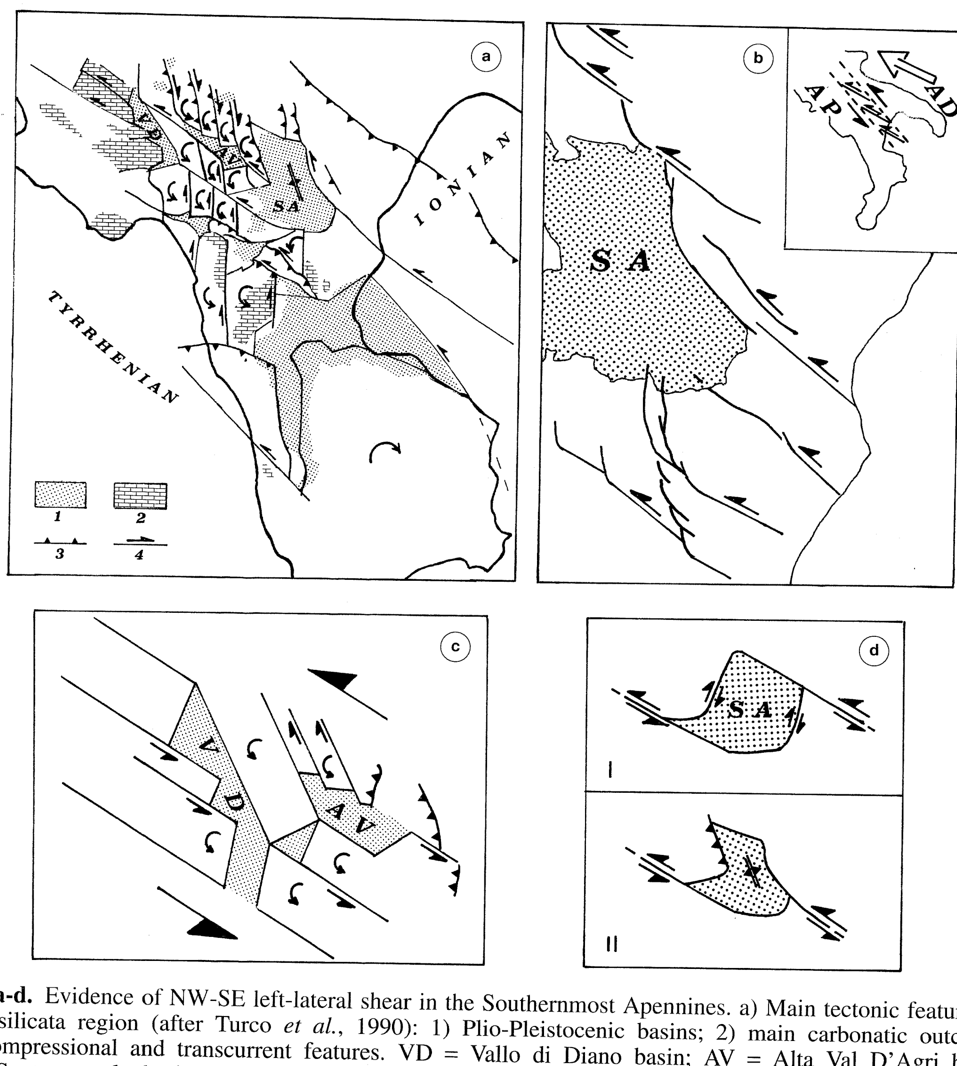

Fig. 6a-d. Evidence of NW-SE left-lateral shear in the Southernmost Apennines. a) Main tectonic features in the Basilicata region (after Turco *et al.*, 1990): 1) Plio-Pleistocenic basins; 2) main carbonatic outcrops; 3,4) compressional and transcurrent features. VD = Vallo di Diano basin; AV = Alta Val D'Agri basin; SA = Santarcangelo basin. b) Main transcurrent faults in the Basilicata zone and interpretative kinematic scheme, after Monaco (1993). AD = Adriatic block; AP = Apenninic domain; SA = Santarcangelo basin. c) Model of the formation of the Vallo di Diano (VD) and Alta Val D'Agri (AV) basins proposed by Turco *et al.* (1990). d) Evolutionary model for the Santarcangelo basin (SA) proposed by Turco *et al.* (1990).

about E-SEward to southward (Oldow *et al.*, 1990). This tectonic evidence is coherent with paleomagnetic observations (Meloni *et al.*, 1997).

Paleomagnetic observations in the sediments overlying the Apulia region (fig. 1a,b) suggest that the Southern Adriatic has not undergone any significant rotation since the Pleistocene (Scheepers, 1992).

2.3.2. Eastern Mediterranean

Since the late Pliocene, the deformation pattern of the zone between the Southern Peloponnesus and Crete (fig. 7a,b) has been dominated by approximately E-W extension with a dextral shear, which produced an important feature, the Kithira trough (Lyberis *et al.*, 1982; Lyon-Caen *et al.*, 1988; Armijo *et al.*, 1992).

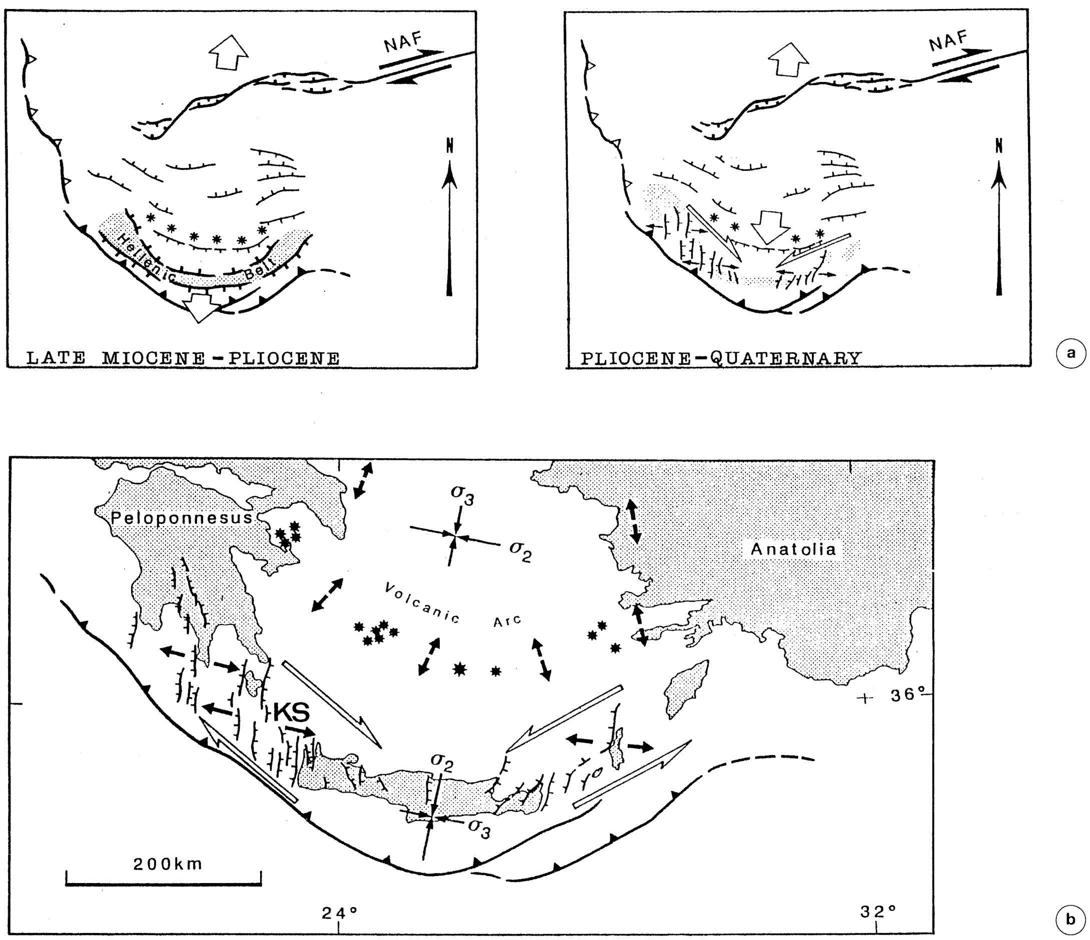

Fig. 7a,b. Miocene-Quaternary evolution (a) and present deformation pattern (b) in the Aegean region. In (a) large open arrows indicate the bulk Aegean extension. In (b) small solid arrows indicate the directions of extension and half arrows represent horizontal shear. KS = Kithira Strait. Other symbols as in fig. 1a,b (after Armijo *et al.*, 1992, modified). It can be noted that the average direction of active extension (σ_3) in the Central-Northern Aegean zone is perpendicular to that in the Crete region.

The so-called «Mediterranean ridge» has been affected by a strong N-S shortening in its central sector just south of Crete. This evidence has been interpreted as an effect of an incipient continental collision between Crete and the Libyan promontory (Ryan *et al.*, 1982; Lyon-Caen *et al.*, 1988; Armijo *et al.*, 1992; Chaumillon and Mascle, 1995).

During the late Pliocene, extensional tectonics began in the eastern sector of the Cretan basin, and the land connection between Rhodes and Crete was cut off (Buttner and Kowalczyk, 1978; Armijo *et al.*, 1992). In the Karpathos basin, the analysis of microseismicity shows reverse faulting with a NE-SW shortening direction (Hatzfeld *et al.*, 1993).

In the lower Pleistocene, a renewal of compressional deformation and uplift took place in the Western Aegean domain, and in particular in the Northern Hellenides, the Albanides, the Ionian islands and the Peloponnesus (Sulstarova *et al.*, 1980; Kelletat *et al.*, 1978; Buttner and Kowalczyk, 1978; Mercier *et al.*, 1989).

Since the early-middle Pleistocene, the deformation pattern of the internal Aegean zone is compatible with an E-W shortening and S-N extension (Mercier *et al.*, 1987, 1989; Lykousis *et al.*, 1995).

In the Quaternary, a roughly N-S extension has occurred in the Corinthian and Ambracique gulfs (Tagari *et al.*, 1993; Hatzfeld *et al.*, 1993).

During this phase, the central-western part of the Cretan basin has experienced very low deformation (see, *e.g.*, Angelier *et al.*, 1982) and is currently affected by a minor seismic activity (see, *e.g.*, Jackson, 1993).

3. Present tectonic setting

Reviews on the large scale recent-present strain field, deduced from the analysis of seismicity taking into account the available stress-*in-situ* measurements and microtectonics, are reported by Philip (1987) and Rebai *et al.* (1992). A tentative reconstruction of rheological profiles in a number of sites throughout the study area is given by Viti *et al.* (1997).

Direct information on present tectonics may be also derived from long range satellite geodesy (VLBI, SLR, GPS), (Zarraoa *et al.*, 1994; Smith *et al.*, 1994; Noomen *et al.*, 1996). However, due to the relatively small time interval covered by observations and the distribution and low density of station points, definitive conclusions on present kinematics cannot be drawn. A discussion on the uncertainties which may affect this type of information is reported by Mantovani *et al.* (1995).

3.1. *Central Mediterranean*

Seismic strain analyses (Anderson and Jackson, 1987a; Jackson and McKenzie, 1988; Albarello *et al.*, 1993; Pondrelli *et al.*, 1995; Mantovani *et al.*, 1996) indicate that a compressional regime currently affects the eastern boundary of the Adriatic microplate from the Eastern Southern Alps to the Hellenic arc, with a horizontal maximum compression which rotates from N-S in the Eastern Southern Alps, to NE-SW along the Dinarides and Hellenides, to E-W in the Epirus.

The most intense seismic activity in the Apennines is determined by tensional seismogenic structures with Apenninic trends (Anderson and Jackson 1987a; Westaway *et al.*, 1989; Viti *et al.*, 1992; Pondrelli *et al.*, 1995). However, if the analysis is extended to smaller events the deformation pattern, at least for the Northern Apennines, seems to be more complex with tensional mechanisms in the inner Apenninic belt and compressional mechanisms along the external fronts (*e.g.*, Eva and Pastore, 1993). This pattern is more in line with neotectonic data (see, *e.g.*, Bartolini *et al.*, 1983; Boccaletti *et al.*, 1985; Bousquet and Philip, 1986; Castellarin and Vai, 1986; Bartole, 1995).

The analysis of fault plane solutions (Viti *et al.*, 1992; Caccamo *et al.*, 1996; Albarello *et al.*, 1997) suggests that the Calabrian arc is affected by a roughly N-S compressional regime in line with geodetic data (Zarraoa *et al.*, 1994; Lanotte *et al.*, 1995) and geological analyses (Philip, 1987; Lo Giudice and Rasà, 1986; Van Dijk and Okkes, 1991). This regime is accompanied by E-W tensional activity along the seismogenic structures responsible for the most intense earthquakes in the region.

Information on crustal structure features along several profiles in the Central Mediterranean region are given by Scarascia *et al.* (1994).

Investigations on intermediate-deep seismicity and tomographic analyses, which point out the presence of cold lithospheric bodies subducted beneath the Calabrian arc and Northern Apennines are reported by Anderson and Jackson (1987a), Spakman (1990), Giardini and Velonà (1991), Amato *et al.* (1993), Cimini and Amato (1993), Frepoli *et al.* (1995).

3.2. *Eastern Mediterranean*

Seismic deformation in the North Aegean is compatible with E-W shortening and S-N extension (see, *e.g.*, Taymaz *et al.*, 1991). This pattern is quite similar to seismic strain deformations observed along the western segment of the North Anatolian fault system (Jackson and McKenzie, 1988; Pondrelli *et al.*, 1995) and suggests a general sinistral transcurrent regime from North-Western Turkey to the Northern Aegean trough.

Both geodetic (Billiris *et al.*, 1991) and seismological analyses (Papazachos and Kiratzi, 1996) suggest that N-S extension in the Corinthian and Ambracique gulfs is currently active and characterized by a rate of roughly 1 cm/yr. This result is in line with geological data concerning the Quaternary deformation pattern (Mercier *et al.*, 1987, 1989). This pattern, which also affects the Peloponnesus and surrounding zones, has been interpreted as an effect of a WNW-ESE compressional regime (Stiros, 1988) compatible with the seismic regime observed in the Northern Aegean. The Kefallinia fault is affected by strong earthquakes, associated with dextral transpressional mechanisms (Scordilis *et al.*, 1985). Seismic deformation analysis indicates that widespread N-S tensional deformations affect mainland Greece, Rhodope and Western Turkey (Taymaz *et al.*, 1991; Papazachos and Kiratzi, 1996). In line with geological data concerning Quaternary deformations (Angelier *et al.*, 1982), the central-western part of the Cretan basin is only affected by minor seismicity. Estimates of seismic strain rates along the Hellenic trench (Papazachos and Kiratzi, 1996) indicate a present-day NE-SW compression. Along the easternmost segment of the trench, this regime is accompanied by SE-NW extension. The presence of intermediate seismicity and the results of tomographic analyses disclose the presence of cold lithospheric bodies subducted beneath the Hellenic trench (see, *e.g.*, Spakman, 1990).

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