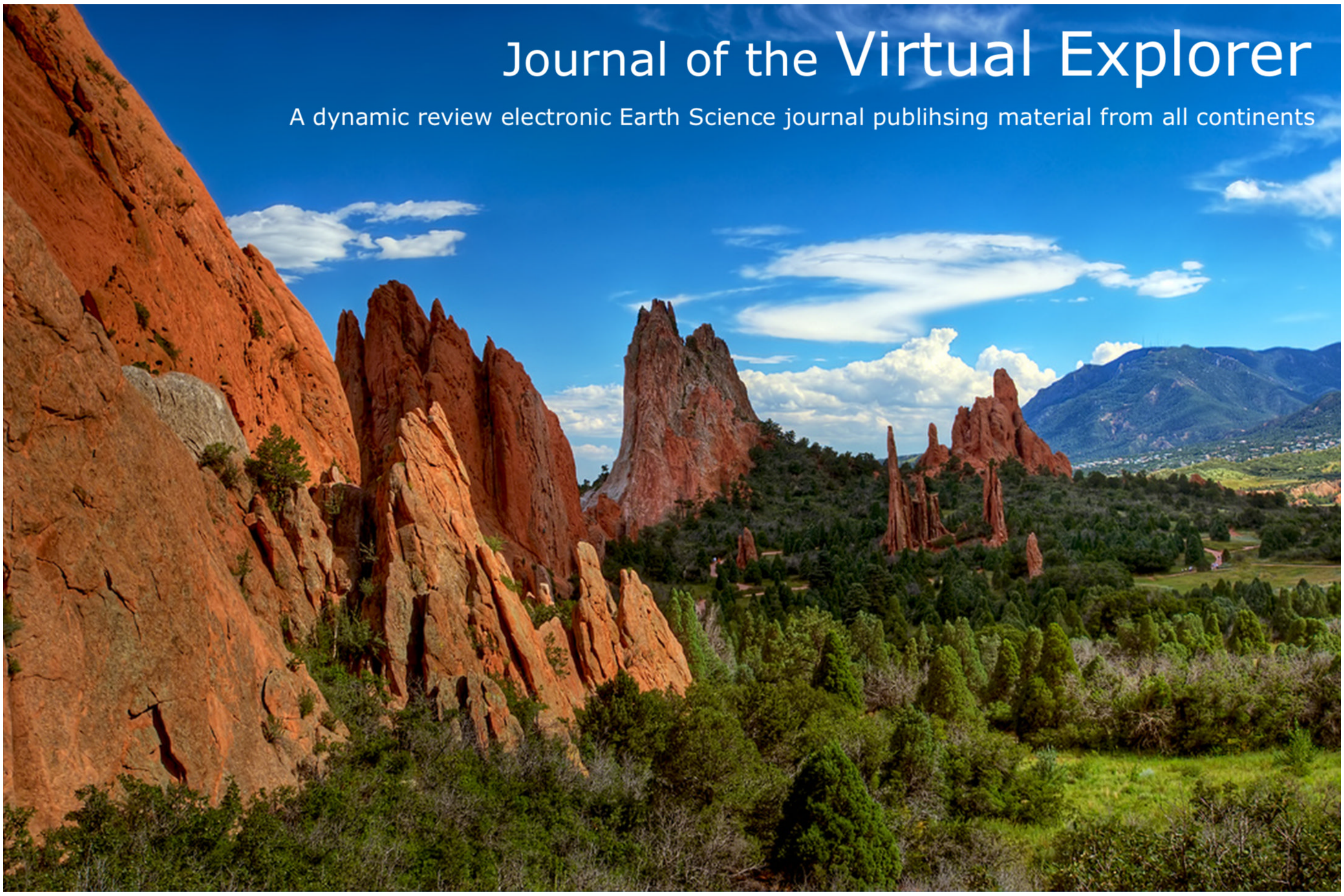

On the extension in western Anatolia and the Aegean sea

C. Doglioni, S. Agostini, M. Crespi, F. Innocenti, P. Manetti, F. Riguzzi, Y. Savasci

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Reconstruction of the evolution of the Alpine-Himalayan orogen, 2002.

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Abstract: Geodynamic reconstructions of the eastern Mediterranean suggesting either westward anatolian extrusion, or gravitational collapse of thickened lithosphere, can be ruled out because plates velocity vectors increase from eastern Anatolia to the Aegean and Greece. This contradicts the basic rule that the velocity field decreases moving away from the source area of the energy, i.e. the supposed squeezing of Anatolia due to the Arabia indenter, or the collapse of the Anatolian orogen. Moreover the topographic gradient between Anatolia and the Ionian deep basin is too small (<1°) for providing sufficient energy relief able to explain present deformation. The simplistic view of the westward Anatolian escape would rather close the Aegean Sea. We interpret the extension in western Turkey, Aegean sea, Greece and Bulgaria as a result of the differential convergence rates between the northeastward directed subduction of Africa relative to the hangingwall disrupted Eurasian lithosphere. Considering fixed Africa, the faster southwestward motion of Greece relative to Cyprus-Anatolia determines the Aegean extension. A new IERS-based solution is presented. The differences in velocity are ascribed to differential decoupling with the asthenosphere. Unlike west-Pacific backarc basins where the asthenosphere replaces a subducted and retreated slab, the study area represents a different type of extension associated to a subduction zone, where the hangingwall plate overrode the slab at different velocities, implying internal deformation. The slab may be folded by the isostatic rebound of the mantle beneath the ‘backarc’ rift, and stretched for the increasing length of the slab between Greece and Anatolia. A sort of window then formed both in the slab and between the two upper plates, allowing uprise of mantle derived Na-rich magmas.

http://virtualexplorer.com.au/article/2002/49/extension-in-western-anatolia-and-the-aegean-sea

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Introduction

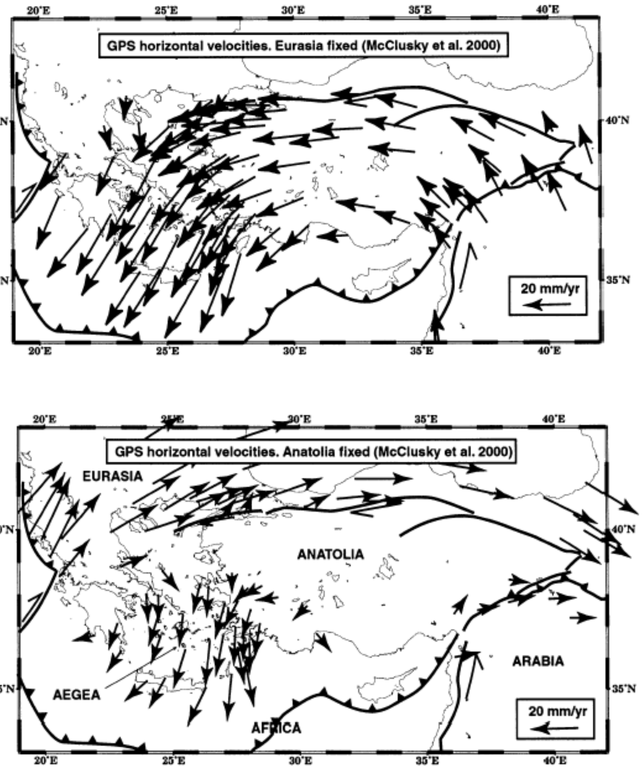
The west-Anatolian-Aegean area underwent a series of multiple continental collisions which started in the Mesozoic (Sengor and Yilmaz, 1981). The collisional zone (e.g., the Vardar-Izmir-Ankara suture zone) migrated southwestward or southward, down to the present position of the Cyprus-Hellenic subduction zone. There are several papers that described the opening of the Aegean Sea and the surrounding extensional areas as a backarc basin (e.g., Horvath and Berckhemer, 1982, Royden, 1993) in the hangingwall of the Hellenic subduction zone (Christova and Nikolova, 1993) superimposing the Mesozoic to present orogens. Since McKenzie (1972), the migration of the Hellenic arc has been related to the northward push of the Arabian plate, and the interpreted consequent westward escape of the Anatolian plate. This idea was and still is so popular that in the literature it became a sort of dogma for the eastern Mediterranean geodynamics (e.g., Yaltõrak et al., 1998; Martinod et al., 2000; Bozkurt, 2001, and references therein). Recently, Seyitoglu and Scott (1996) and Gautier et al. (1999) questioned this point, showing that the Aegean extension pre-dates (Upper Oligocene-Lower Miocene) the Africa-Eurasia and Arabia-Eurasia collisions (Middle-Upper Miocene) and therefore it cannot be a consequence of it. Extension in the “Aegean-Anatolia” realm is probably even older, dating back to the Late Eocene in Eastern Bulgaria as indicated by the early stages of opening of the Burgas Basin (Doglioni et al., 1996). It was then claimed that the gravitational collapse (Seyitoglu and Scott, 1996) allowed by extensional boundary conditions after 30 Ma (slab retreat) could be the primary cause for post-orogenic extension in the Aegean and Western Anatolia (Gautier et al., 1999; Jolivet, 2001). Two main magmatic signatures occur in the study area since Miocene times, i.e., 1) typical subduction-related, southwestward migrating Miocene to Present calk-alcaline volcanism and 2) Upper Miocene-Quaternary alkaline Na-rich magmatism (Benda et al., 1974; Yilmaz et al., 2001).

In this paper we discuss an alternative geodynamic scenario for the eastern Mediterranean. It is based on a multidisciplinary approach of structural field work, new geodetic computations, magmatic constraints and tectonic modeling.

Geodynamic setting

Africa, Greece, Anatolia, Eurasia and Arabia: these are the plates involved in the geodynamic reconstructions of the eastern Mediterranean. Deformation is very active in the all area (Papazachos and Kiratzi, 1996; Bozkurt, 2001). The most prominent geodynamic factor in shaping the eastern Mediterranean is the northeast directed subduction of Africa underneath Greece and the Anatolia plate (Eurasia). Seismic lines across the Cyprus arc (Sage and Letouzey, 1990; Vidal et al., 2000), at the southern margin of the Anatolia plate, show clear active compression and deformation of the seafloor.

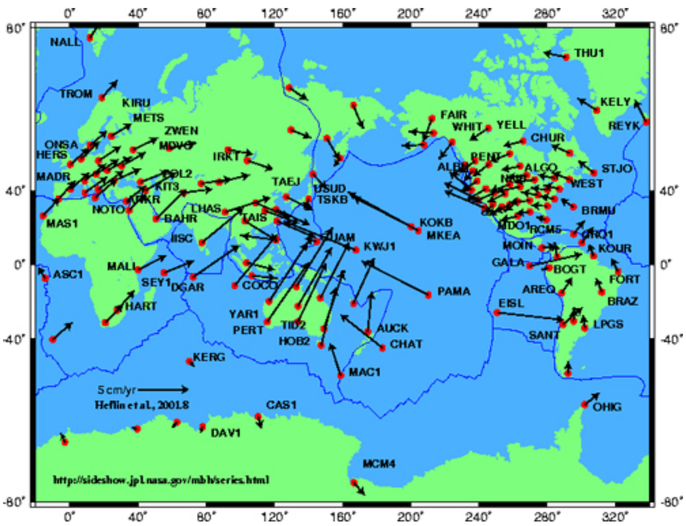
Figure 1. GPS velocity field



GPS velocity field after McClusky et al. (2000) relative to Eurasia (above) and to Anatolia (below). Note the increase in velocity moving westward which is not consistent neither with squeezing of the Anatolia due to the supposed indenter of Arabian, nor to a gravitational collapse. The velocity should rather decrease moving away from the energy source. See discussion in the text.

The Aegean Sea is generally considered as a backarc basin due to the aforementioned subduction (Le Pichon and Angelier, 1979). However, according to Makris (1978) and Makris et al. (2001), the Aegean Sea is characterized by a relatively thick crust (20-25 km), in spite of a long standing subduction, probably active at least since Cretaceous times (Meulenkamp et al., 1988). The subduction zone migrated southwestward, and the associated orogen was later replaced by the extension. In the Aegean Sea, Alpine crustal thickening with HP/LT were followed by non co-axial crustal-scale extension (Rosenbaum et al., 2002). This is coherent with first the emplacement of thrust-sheets of basement slices, later cross-cut by extensional or transtensional faults. Extension and associated magmatism also were and still are south-southwest migrating (e.g., Jackson and McKenzie, 1988), and they developed particularly since the Oligocene, while subduction began much earlier. "Normal" backarc basins (e.g., the Tyrrhenian Sea) associated with west-directed subduction zones open very fast (10-20 Myr) and they are always contemporaneous to the subduction. Moreover they are characterized by oceanization and eastward migration of extension and related magmatism, features directly surrounded by a frontal accretionary wedge. On the contrary, the accretionary wedge of the Hellenic subduction is the southeastern prolongation of the Dinaric thrust belt, where no back-arc rift comparable to the Tyrrhenian Sea occurs.

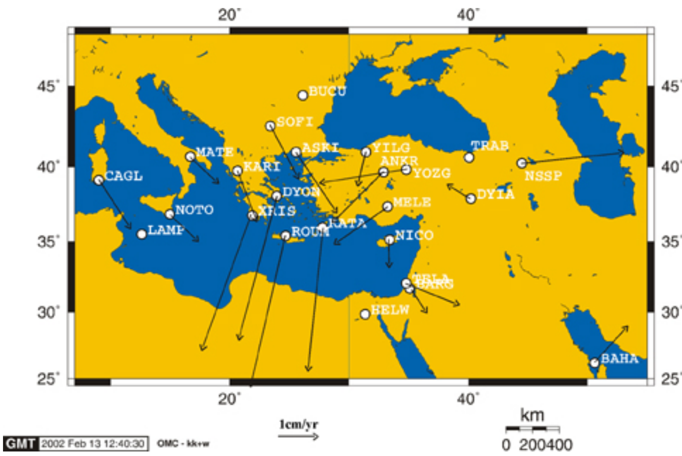
Figure 2. Present day plate motion vectors



Present day plate motion vectors inferred from space geodesy, with reference to an hypothetical fixed Earth center (ITRF, International Terrestrial Reference Frame). After NASA data base, Heflin et al. 2001.8. Note the main north-eastward direction of plates motion in the Mediterranean and in the adjoining plates (e.g., Arabia, India, Europe, Africa).

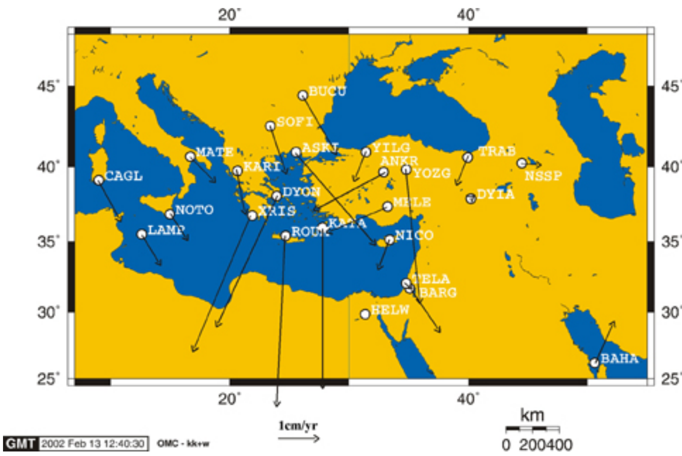
In spite of that, most geodetic analysis in the study area have been computed assuming the motion of the Aegean area and Anatolia relative to fixed Europe or sometimes to Turkey (e.g., Kahle et al., 1998, McClusky et al., 2000, Figure 1), rather than to fixed Africa. This was also related to the paucity of geodetic sites in northern Africa. However, since the main deformation occurs along the subducting plate margin of Africa underneath Eurasia, we tried to analyze the relative motion in this different reference frame.

Figure 3. ITRF97 velocity field



ITRF97 velocity field computed with respect to HELW (Egypt, African plate).

Figure 4. ITRF2000 velocity field



ITRF2000 velocity field computed with respect to HELW (Egypt, African plate).

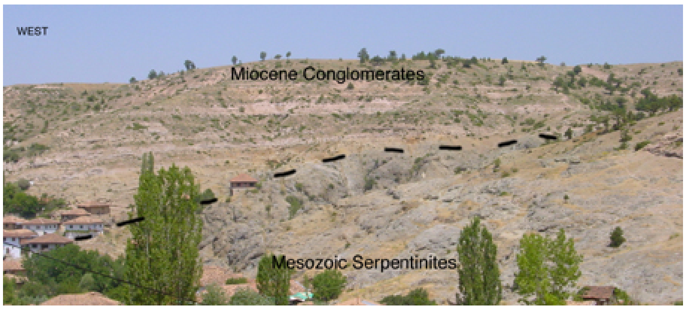
In an “absolute” reference frame, the NASA data base on present global plate motions (Heflin et al., 2001, http://sideshow.jpl.nasa.gov:80/mbh/series.html) referred to the ITRF (International Terrestrial Reference Frame).

It shows how the motions of Africa, Europe, Arabia and Anatolia plates are mainly NE directed (Figure 2). In particular the Arabia plate shows a direction of motion, which strongly deviates with respect to the N-S direction of the paradigm proposed in the dominant literature for the eastern Mediterranean geodynamics. A north-south component in the northwestern margin of the Arabia plate can be expected for left-lateral transpression component at the northwestern part of the Zagros belt. These data of mainly NE-SW direction have been shown to be consistent with directions of plates motion during the Neogene and Quaternary (Doglioni, 1990).

Figure 5. Eocene angular unconformity

Angular unconformity between flat lying Eocene nummulitic shallow water limestone, resting on highly tilted and folded Cretaceous-Early Paleogene Scaglia facies. This documents pre-Eocene active compression in northern Turkey. North to the right. Road between Adapazari and Bileçik, 5 km north of Osmaneli (40° 26' N, 30° 02' E).

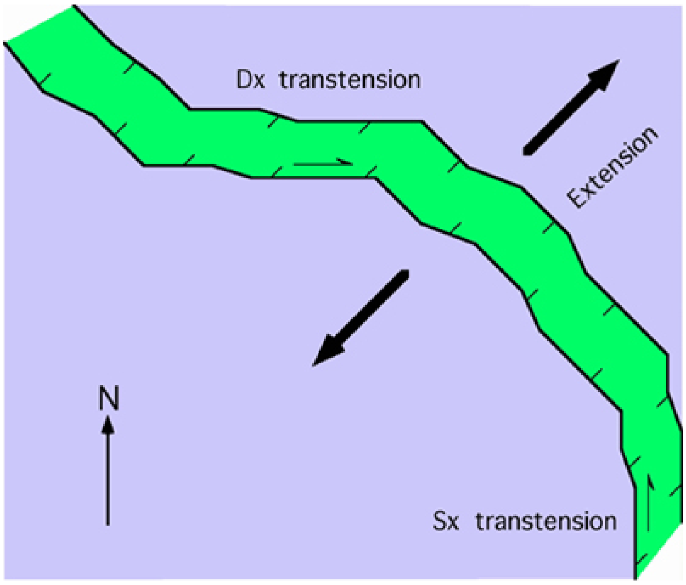
Figure 6. Miocene angular unconformity



Angular unconformity between flat lying Miocene alluvial conglomerates onlapping and pinching out an erosional surface on the ophiolites, indicating pre-Miocene compression-related exhumation of the serpentinites, erosion, and later subsidence associated with extensional tectonics. Outcrop near Çamsu, northeast of Usak, north-western Turkey (38° 52' N, 29° 36' E).

In other words, we should rather analyze the eastern Mediterranean plate tectonics in terms of both absolute motions, and relative motions of Greece and Turkey relative to Africa. In fact, assuming Africa as a single plate, since Greece is overriding Africa along the Hellenic trench faster than Turkey along the Cyprus arc, it turns out that there is a positive velocity among Greece and Turkey in the hangingwall of the subduction zone. In fact, Greece is moving SW-ward above Africa faster than Turkey (Figure 3 and Figure 4), which implies extension between Greece and Turkey in the Aegean area and western Anatolia, without any relation in this circuit with the Arabia plate. According to this geodynamic interpretation, during the compressive events associated with northeast directed subduction (Figure 5), basement rocks (both continental and ophiolitic slices) in west Anatolia and the Aegean Sea were uplifted and eroded. Later extension subsided the area, and the basement slices were covered by continental and marine sediments (Figure 6). The extension in western Anatolia and Aegean area was, and still is, dominated by similar absolute framework of plate motion. The previous horizontal NE-SW direction of the sigma 1 in the compressive stage became the trend of the sigma 3 (Figure 7), implying NW-SE trending normal faults or grabens, about E-W-trending right-lateral transtensional faults, and N-S-trending left-lateral transtensional faults. The study area is cross-cut by a net of faults which follow this rule. See a field example of northwest Anatolia with left-lateral transtension in Figure 8.

Figure 7. Miocene-Quaternary faults



Main directions and tectonic meaning of Miocene-Quaternary faults in western Anatolia and Aegean.

Figure 8. Left lateral transtensive fault

Field example, looking northward, of a N-S trending left lateral transtensive fault near Kirkagaç, 20 km NE of Akhisar (39° 06' N, 27° 42' E).

The velocity field and its assessment

On the basis of geodetic observations, in order to infer the velocity field of the Aegean area and Anatolia region and their surroundings, two global solutions were used.

They were computed by the IERS (International Earth Rotation Service) in its institutional activity, based on permanent tracking sites (mainly GPS stations, but also SLR, VLBI, DORIS and recently GLONASS).

In fact, it is well known that reliable geodynamics information can be inferred on a geodetic basis only by permanent geodetic networks (Beutler, 1998). That also because the analysis of coordinate time series of permanent tracking highlights that spurious, long period oscillations, due to mismodelling and unknown site effects, they may be present and corrupt the geodynamical interpretation. Therefore they must be filtered out in advance (Langbein and Johnson, 1997; Mao et al., 1999).

The mentioned IERS global solutions provide the coordinates, velocities and RMS of each tracking site, obtained after the adjustment of a certain number of sub-network solutions analysed by local data centers, under the hypothesis of steady velocity for each site. The number of stations contributing to the global solution is annually updated. It is worthwhile to recall that, in the geodetic terminology, a set of points with their coordinates and velocities realizes a reference frame. The frames produced by IERS are named IERS Terrestrial Reference Frames (ITRF), they are referred to a precise epoch, as the coordinates change continuously in time (Boucher and Altamimi, 2001).

The two solutions (Figure 3 and Figure 4) considered in the present research are the most recent realizations of ITRF, ITRF97 and ITRF2000, both estimated at epoch 1997.0 (1997, Jan 1).

Concerning the Eastern Mediterranean and the Anatolian region, we analysed the ITRF97 and ITRF2000 solutions focusing our attention on a subset of GPS and SLR stations covering the area of our interest. We select 24 tracking stations (Table 1), 13 observed continuously by GPS (12) and SLR (1) and 11 by mobile SLR, sites considered less reliable than the permanent ones (Altamimi, pers. comm.).

We computed the northward and eastward components of velocity Vn, Ve (Table 1). We analyzed the 2D velocities (Figure 3 and Figure 4) with their azimuth for the selected sites with respect to HELW (Helwan, Egypt, African plate), starting from the Vx, Vy and Vz components listed in the ITRF97 and ITRF2000 solutions (http://lareg.ensg.ign.fr/ITRF/).

Figure Table 1. Geodetic stations and velocity components

SITES				ITRF97		ITRF2000	
		Lat.	Lon.	Vn (m/yr)	Ve (m/yr)	Vn (m/yr)	Ve (m/yr)
SOFI	GPS	42.56	23.39	-0.014	0.007	-0.012	0.004
BUCU	GPS	44.46	26.13	-	-	-0.016	0.009
NSSP	GPS	40.22	44.5	0.003	0.027	0	0.005
NOTO	GPS	36.87	14.98	-0.007	0.008	-0.006	0.004
CAGL	GPS	39.13	8.99	-0.013	0.008	-0.011	0.006
MATE	GPS	40.65	16.7	-0.007	0.009	-0.006	0.006
NICO	GPS	35.14	33.4	-0.007	0	-0.007	-0.003
YOZG	SLR	39.8	34.8	-0.004	-0.022	-0.036	0.004
DIYA	SLR	37.92	40.2	0.004	-0.006	-0.013	-0.012
TRAB	GPS	40.99	39.78	-	-	-0.007	-0.003
MELE	SLR	37.38	33.2	-0.01	-0.014	-0.007	-0.015
ANKR	GPS	39.89	32.76	-0.014	-0.014	-0.01	-0.019
BAHR	GPS	26.21	50.61	0.01	0.009	0.011	0.005
YIGI	SLR	40.93	31.43	-0.009	-0.002	-0.007	-0.003
DYON	SLR	38.08	23.93	-0.038	-0.01	-0.034	-0.016
ASKI	SLR	40.93	25.57	-0.015	0.01	-0.024	0.021
ROUM	SLR	35.4	24.69	-0.039	-0.009	-0.045	-0.002
KARI	SLR	39.73	20.66	-0.013	0.005	-0.011	0.002
KATA	SLR	35.96	27.78	-0.038	-0.004	-0.042	0
XRIS	SLR	36.79	21.88	-0.036	-0.013	-0.036	-0.015
BARG	SLR	31.72	35.08	-0.006	0.004	-0.01	0.007
HELW	SLR	29.86	31.34	0	0	0	0
LAMP	GPS	35.5	12.61	-	-	-0.008	0.005
TELA	GPS	32.07	34.8	-0.006	0.014	-0.002	0.003

Geodetic stations and velocity components

Reliability control

Since the global solutions are continuously updated, depending on the new availability of permanent sites, we decided to crosscheck their stability concerning the area under investigation.

Therefore, we performed 3σ and 2σ tests between ITRF97 and ITRF2000. We particularly computed the residuals between corresponding velocity components, the overall RMS (1s) for Vn, Ve, and finally we tested these residuals at two significance level (3σ and 2σ) in order to identify possible outliers, i.e., inconsistencies between the two global solutions.

Moreover, we made the same tests between each ITRF and the solution provided by McClusky et al. (2000) for the common sites, after having estimated and removed the velocity drift between the different reference frames. It is useful to recall that McClusky et al. (2000) solution is mainly based on non-permanent stations and was adopted as geodetic background for the most recent geodynamic interpretations of the Aegean Sea and Anatolian area.

The relevant results of the statistical solution comparisons are the following:

- 3σ sign. level for all comparisons: all the sites pass the test, except YOZG (ITRF2000)
- 2σ sign. level ITRF97-ITRF2000 comparison: NSSP, YOZG, ASKI and TELA do not pass the test (mean values σVn=0.9 cm, σVe=1.0 cm)
- This means that velocities change more than 2σ and may be interpreted as an improvement in velocity estimate for GPS site (NSSP and TELA, more observations available in 2000) and in low reliability of some SLR mobile sites (YOZG, ASKI)
- 2σ sign. level ITRF97-McClusky comparison: NSSP, YOZG, MELE, YILG, HELW and TELA do not pass the test (mean values σVn=0.3 cm, σVe=0.6 cm)
- 2σ sign. level ITRF2000-McClusky: YOZG, TRAB, MELE, YILG, DYON, ASKI, ROUM and HELW do not pass the test (mean values σVn=0.8 cm, σVe=0.7 cm)

It should be noted that the estimated agreement between ITRF solutions and McClusky et al. (2000) solution is remarkably poorer than the claimed precision of the McClusky et al. (2000) solution velocity, which, on the contrary, seems to be too optimistic in spite of the non-permanent data set on which they are based.

The illustrated comparisons show a good coherence between the ITRF solutions at centimetric level, also at the more severe 2σ sign. level, since only 4 sites exhibit high residuals; on the contrary, the agreement with the McClusky et al. (2000) solution seems to be more problematic.

Therefore, in our opinion, the performed assessment of the available geodetic solutions for the Anatolia region and its surroundings suggests: to base geodynamical investigation on the IERS solutions at present; to account the average precision of these solutions, which is at centimetric level.

Magmatic constraints

The area extending from the circum-Aegean region to Central Anatolia is characterized by the presence of widespread Tertiary and Quaternary volcanism, showing a complex spatial and temporal distribution. Two extensive regions, separated by a narrow junction area, can be distinguished.

The Aegean-West Anatolia Area includes the Rhodope Massif-Thrace, the Central Aegean Sea and western Anatolia up to the South Aegean Active Volcanic Arc. In this area, a calc-alkaline and alkaline vulcano-plutonic associations occur, spanning in age from Upper Eocene to Present.

In the northernmost part of the area (Rhodopian Massif and western Thrace), apart from the Upper Cretaceous calc-alkaline rocks, the volcanism took place from Upper Eocene (~ 37 Ma, Priabonian) to Upper Oligocene (~24

Ma, Yanev, 1998; Innocenti et al., 1984). Erupted products have a petrogenetic affinity, variable from high-K calc-alkaline to shoshonitic with minor ultra-potassic products. Scattered K-basanites and horblend lamprophyres (camptonites) of 28-26 Ma with an OIB geochemical imprint are also found (Marchev et al., 1998). The volcanism shows a southward migration and a decreasing volume with time of the emitted products; a parallel average K2O decrease is also observed. The volcanic activity is associated with a plutonism encompassing a larger interval of time (41-15 Ma, Christofides et al., 1998).

In western Anatolia an Upper Eocene-Oligocene magmatism is also recognized; however, the volume of igneous (intrusive and volcanic) products is significantly smaller than that observed in the Rhodopian-Thracian zone (Yilmaz et al., 2001). Most of the magmatic activity developed during Lower-Middle Miocene (up to about 14 Ma), as well as in several islands of the Central Aegean Sea (Fytikas et al., 1984; Pe-Piper and Piper, 1989). The volcanic products are accompanied by cogenetic plutonic rocks (e.g. Mount Kozak, Altunkaynak and Yilmaz, 1998; Evciler Pluton, Genç, 1998; Izmir, Ayvalik and Bodrum, Savasçin, 1990). The igneous suite is ranging in composition from high-K andesite to rhyolite. Large dacitic-rhyolitic ignimbritic covers of lower Miocene age are widespread in western Anatolia and in eastern-northern Aegean Sea, as in the islands of Lesvos and Limnos. The upper part of this volcanic sequence is characterized by the presence of K-rich basic-intermediate rocks with shoshonitic affinity, sometimes showing ultrapotassic feature. Recently, lamproites as subvolcanic dykes and small lava flows have been found in the central part of western Anatolia (Gediz area), at the end of this orogenic volcanic cycle (14-15 Ma, Savasçin et al., 2000).

After 14-15 Ma the volcanic activity virtually ceased. It resumed in the Late Miocene after about 4 Ma. The erupted products are, in general, poorly evolved, being alkali basalts the most widespread terms; they frequently display a slight potassic affinity (e.g. Eskishehir, Francalanci et al., 2000). These alkali basaltic rocks are found in scattered occurrences in Western Thrace (Yilmaz and Polat, 1998), western Anatolia (Aldanmaz et al., 2000) and central and eastern Aegean Sea (Fytikas et al., 1979; Seyitoglu and Scott, 1992). The subduction-related geochemical signature of these products is generally slight or absent, and several lavas show an OIB character. In Quaternary times, Na-alkali basalts with a clear OIB signature were erupted in western Anatolia (Kula area) and in NW Aegean Sea, where they are located on the prolongation of the North Anatolia fault system (Psathura islet).

The outcropping area of the Eocene-Middle Miocene orogenic rocks is limited southward by the crystalline Actic-Cycladic and Menderes massifs, which are constituted by high P-T metamorphic rocks which reached the eclogite or blueschist-facies peak conditions around 40-50 Ma and successively overprinted by high T/ medium P metamorphism during Late Oligocene- Early Miocene (26-16 Ma, Bröcker et al., 1993). This latter metamorphic phase was accompanied by the intrusion of granites generally of Middle Miocene age (Altherr et al., 1982), locally associated with anatectic Late Miocene-Early Pliocene rhyolites. These volcanics are either exposed (e.g., Antiparos, Innocenti et al., 1982) or occurring as acid clasts in the tectono-sedimentary units overlain the blueschist unit (Sanchez-Gomez et al., 2002). Overall, geochronological data suggest that the Cycladic area has been invaded in the Early-Middle Miocene by large crustal granitic plutons of S- and I-type, that extended their presence also in the Menderes Massif (Delaloye and Bingöl, 2000).

The calc-alkaline volcanic activity started again in Early Pliocene south of the Attic-Cycladic Massif. In fact, a narrow active volcanic arc formed as expression of the active northeastward-directed subduction. The erupted products are typically calc-alkaline and constitute an association ranging in composition from orogenic basalts to rhyolites (Mitropoulos et al., 1987).

In Central Anatolia area, the volcanic activity developed essentially on the Central Anatolia Cristalline Massifs (Kirsehir-Akdag-Nigde, Whitney and Dilek, 1988) during the Middle Miocene-Quaternary, in two main areas located south-west of Konia and in Nevsehir-Kayseri area. The oldest products are represented by calc-alkaline Serravalian-Tortonian andesites (Besang et al., 1977); the climax of the volcanic activity took place in the Late Miocene-Pliocene when a huge ignimbritic sequence emplaced and formed the Ürgüp-Nevsehir plateau extending on about 12000 km2 (Aydar et al., 1995). The ignimbrite plateau includes lacustrine and fluvial deposits; it represents the basement of large composite volcanoes as Karacadag and Melendiz Dag (Mio-Pliocene) and Hasan Dag and Erciyes Dag (Quaternary). Many monogenetic Quaternary centers, either of basaltic or rhyolitic compositions, grew forming cinder and spatter cones and maars structures often with intra-crateric domes.

The erupted products constitute a high-K calc-alkaline association ranging from few basaltic andesites, to dominant andesites and dacites. The ignimbrite units are made up mainly by rhyolites (Temel et al., 1998) The lavas of the main composite Quaternary volcanoes (Hasan Dag and Ercyes Dag) are characterized by relatively lower K2O contents forming a typical calc-alkaline suite (Aydar and Gourgaud, 1998), whereas the basaltic rocks erupted by the monogenetic centers display a geochemical Na-alkaline OIB signature, as inferred by the composition of clinopyroxene (Aydar et al., 1995) and by the trace elements distribution (Kürkçüoglu et al., 1998).

The two areas are connected by a relatively narrow region, north-south-trending, from Afyon to Isparta angle (Kirka-Afyon-Isparta junction-area), in which an alkaline association developed from the Upper Miocene to the Pliocene. The volcanic rocks form a narrow belt, about 200 km long and 50 km wide, and they were erupted mainly along the Antalya fault zone (Yagmurlu et al, 1997). The oldest products (~21-15 Ma, Besang et al., 1977) are represented by the ignimbrite sheets of rhyolitic composition cropping out in the Kirka area and forming the base of the volcanic sequence in the northern part of the Afyon Massif. The igninbritic sequence is capped by K-alkaline lavas, breccias and pyroclastic flows with a Middle-Late Miocene age (from 14.8 Ma to 8.6 Ma, Besang et al., 1977; Savasçin et al., 1995). Two main potassic associations are documented in the Afyon area (Francalanci et al., 2000). The first one is almost-saturated potassic suite of composition ranging from trachybasalts to trachytes, through shoshonites and latites; these rocks display a typical orogenic geochemical signature with high LILE/HFSE ratios, and Nb, Ta and Ti negative anomalies in the primordial mantle normalized spidergrams. The second association is constituted by strongly alkaline, silica undersaturated to saturated rocks yet keeping a well-defined subduction-related geochemical imprint. The dominant products are phonolitic leucitites.

In Isparta region, two associations have been identified. The predominant rocks are K-rich undersaturated to saturated alkaline volcanics, varying from tephri-phonolites to latites and trachytes with a well-defined subduction-related trace elements signature. In this area a group of volcanic and subvolcanic ultrapotassic silica-undersaturated rocks is also present, significantly enriched in MgO (MgO>10%) and compatible elements, which display OIB- or intraplate-type geochemical imprinting with low LILE/HFSE and REE/HFSE and quite low Sr isotope ratios (~0.7038, Francalanci et al., 1991).

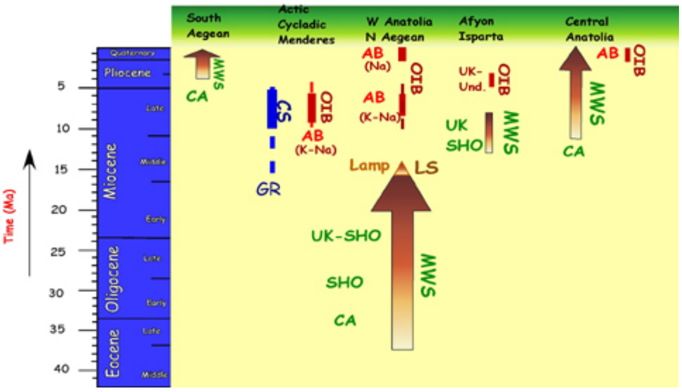
As a whole, in the Anatolia-Aegean area the space-time evolution of the volcanism indicate that the erupted products were fed by different mantle sources, whose activation was strongly related to the geodynamic evolution of the region. On the base of isotopic, geochemical and petrological characteristics of igneous rocks, three main mantle sources have been identified.

The dominant calc-alkaline to shoshonitic products show geochemical features indicative of an origin by a mantle source deeply modified by a subduction-related component. We consider that this source was located in the mantle wedge between the African subducting slab and the overriding Anatolia-Aegean plate. The geochemical variations observed in these products reflect not only the heterogeneity of the source, but also process of interaction of the calc-alkaline magmas with upper crust material, documented in several places in western Anatolia (e.g., Aldanmaz et al., 2000).

The Tertiary orogenic cycle locally ends with lamproitic rocks, interpreted as derived by a lithospheric refractory mantle, affected by a metasomatic event, which enriched the source in LILE and enhanced and lowered the 87Sr/86Sr and 143Nd/144Nd ratios, respectively. The geochemical imprinting of these rocks suggests that the metasomatizing component was yet linked with the subduction process, as stressed by the visible Nb, Ta and Ti negative anomalies in multi-element primordial-mantle normalized diagrams (Savasçin et al., 2000).

From Upper Miocene up to Quaternary, the scattered alkali basalts occurring in all the different sectors of the Aegean-Anatolia realm, exhibit geochemical signature typical of OIB- type source (e.g., Aldanmaz et al., 2000; Yilmaz et al., 2001; Wilson et al., 1997; Aydar et al., 1995). They are interpreted as an evidence of the activation of a relatively deep sub-slab mantle source, which was made possible by the absence of the subducting slab or by its inactive presence.

Figure 9. Time distribution



Time distribution of volcanism in Aegean–Awest Anatolia area with the indication of the main rock associations and the involved magma-sources. Mantle wedge source (MWS): CA, calc-alkaline and high-K calc-alkaline; SHO, shoshonites; UK-SHO, ultrapotassic shoshonites; Lithospheric source (LS): Lamp, Lamproites. OIB-type mantle source (OIB): AB, alkali basalts (sodic and potassic); UK-Und., undersaturated ultrapotassic phonolitic tephrites. Crustal source (CS): Granitoids and rhyolites; (GR) with dominant crustal signature.

Beside these mantle sources, in the Attic-Cycladic-Menderes Massifs several granites and rhyolites showing a dominant crustal signature emplaced during Middle-Late Miocene, in connection with a regional-scale extensional phase forming the Aegean Basin, (e.g., Altherr et al., 1982; Innocenti et al, 1982). The time-space evolution of the igneous recognized associations and the main involved magma-sources are schematically outlined in Figure 9. The southwestward migration of the subduction hinge was accompanied in the hangingwall by the southwestward migration of calc-alkaline, shoshonitic and lamproitic magmas. These volcanic products emplaced in an already extending area that eventually evolved in a more extreme setting of mantle uprise, source for the OIB basalts.

On the extension in western Anatolia and the Aegean sea.

Westward Anatolian escape? No thanks

In most of the published models of the geodynamics of the eastern Mediterranean, the westward Anatolia escape is interpreted on the motion of Anatolia and Aegean area relative to Europe. Geodetic data show that the velocity increases from east to west and southwest. This contradicts basic physics, because the velocity should rather decrease moving from the energetic source (the assumed N-S Arabia indenter), and it contrasts the fact that in geodynamics inertial forces are negligible (Turcotte and Schubert, 1982). In other words, the velocity cannot increase with respect to the initial boundary forces. Therefore, in case of an indenter, plate velocity should decrease moving from east to west, which is not what we observe.

How the Aegean Sea could open if the Anatolia plate moved and is moving westward? This should have closed the basin rather opened it. In reality, Anatolia is moving westward relative to Europe (Figure 1), but not relative to Greece, which is moving southwestward even faster than Turkey (Figure 1). Anatolia is not moving westward neither with respect to Africa (Figure 4). A supposed pole of rotation of the Anatolia plate in the southeastern Levantine Sea or Egypt is figured out in a reference frame fixed to Europe (Figure 1). However those vectors do not represent the motion of Anatolia alone, but also of Arabia and Greece. Moreover vectors depict a misleading arc since the geodetic sites are not located on undeformed plate interiors but in areas undergoing regional oblique deformation, where the local stress field is deviated with respect to the main direction of plate motion in a the absolute reference frame. Examples are the left-lateral transpression at the northwestern margin of the Arabia plate generating a NW-SE oriented compression in spite of a NE directed motion of Arabia; or the right lateral transtension in the southern Aegean Sea producing a N-S extension or the left-lateral transpression in the south (e.g., the Strabo trench), generating N-S compression, contemporaneous to the southwestward motion of Greece.

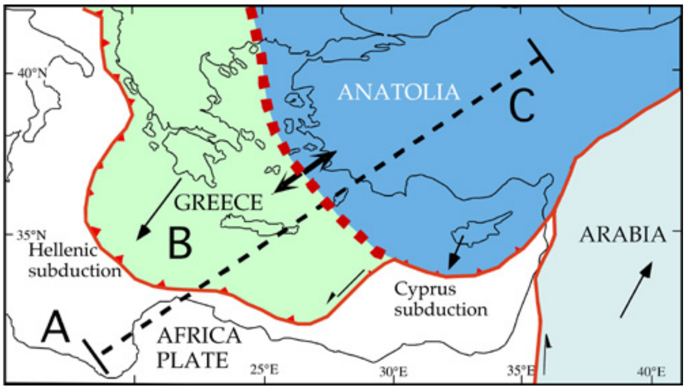
The other quite popular model for the opening of the Aegean Sea and migration of the arc is the gravitational collapse of the thickened lithosphere generated by previous subduction stages. However, also this model presents some odds: 1) The system is still active, with compression in the southern arc and extension in the north Aegean where topography is even below sea-floor. 2) From the deep Ionian (3000 m below sea-level) to the Anatolia Plateau (about 1000 m above sea level) there are about 1000 km with a topography variation of 4 km, which indicates a very gentle slope (less than 1°), too low for generating active gravitational sliding of a brittle crust. Much larger topographic gradients with similar ‘unconstrained’ margins did not generate any lithospheric-scale gravitational collapse: see the arc of the Western Alps and the oceanic Provençal basin to the south.

Analogue modeling (Martinod et al., 2000) tried to mimic the tectonic evolution of the eastern Mediterranean in terms of lateral Anatolian escape and gravitational collapse. These simulations are very helpful, but in their model it is not considered that also Africa is subducting underneath the Hellenic and Cyprus arcs, and not only the Arabia plate underneath Eurasia. Moreover, apart scaling problems, neither the increase in velocity field towards the southwest nor the actual topographic gradient are not discussed. A gravitational collapse model that explains extension in the Aegean area is hard to reconcile when the topography and geodetic data relative to Africa are analyzed and combined.

Geodynamic model

The convergence between Africa and Greece can be hypothesized back to the Eocene at a conservative mean speed of 3 cm/yr. With this velocity, a 1200 km long slab should have formed. The present African slab depicted by seismicity has a low dip (15-20°) and is shallow (no deeper than 200-250 km), becoming more horizontal northeastward, and assuming a spoon shape (Papazachos and Comninakis, 1977; Christova and Nikolova, 1993). Most of the seismicity in the Aegean Sea is rather superficial and the high geothermic gradient is probably responsible for the seismic disappearance of the slab underneath the basin. The opening of the Aegean-western Anatolian rift can kinematically be interpreted as due to the faster southwestward advancement of Greece over Africa, with respect to Cyprus-Anatolia over Africa (Figure 10). The rift generated a mantle uplift to compensate the lithospheric thinning. Therefore the underlying slab itself should have been involved and folded by the mantle uprise beneath the rift (Figure 11). The stretching between Greece and Anatolia, and the differential velocity of convergence with the underlying slabs should have generated a sort of "horizontal windows" both in the hangingwall and in the footwall of the subduction, allowing melting of mantle, and generating the OIB magmatism after regular subduction/collision evolution.

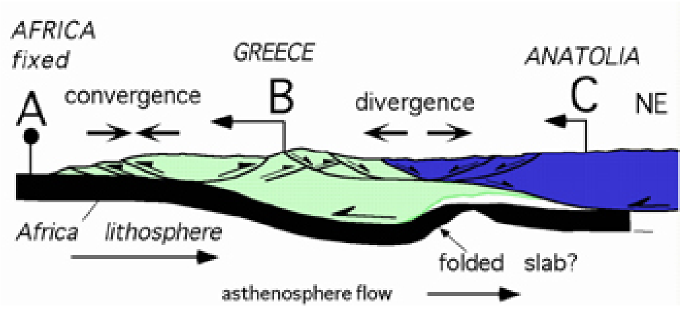
Figure 10. Reconstruction



Considering fixed Africa (A), Greece (B) is overriding Africa faster than Cyprus and Anatolia (C). This implies extension between Greece and Anatolia. Differences in thickness and composition of the subducting Africa lithosphere may determine the faster subduction of the Ionian oceanic lithosphere with respect to the Levantine Sea. The margin between Greece and Turkey is a diffused margin in the entire broad area of extension, and the straight line of the figure is positioned only for distinguishing the two plates. Larger rollback of the Hellenic-Greek slab should determine stretching in the downgoing Africa lithosphere with respect to the Cyprus-Anatolia slab.

No relative motion occurs between the central and eastern Mediterranean, since both sides belong to the same African plate. However, the plate is subducting both below Greece and Cyprus, but at different velocities, or in another reference frame, the hangingwall plate is overriding at different velocities. The Cyprus subduction is in fact slower as indicated by the minor shortening of the Quaternary sediments, by the lower seismicity with respect to the Hellenic subduction, and by geodetic data (Figure 4). In fact thinned continental lithosphere (Makris and Stobbe, 1984) occurs in the footwall of the arc of the eastern Mediterranean, whereas oceanic lithosphere characterizes the Ionian Sea to the west (de Voogd et al., 1992) beneath the Hellenic arc. If no relative motion between central and eastern Mediterranean occurs, the different convergent rates at the two subduction zones has to be related to differential velocities between the hangingwall plates, enabling a faster motion of Greece southwestward over the Ionian relative to Turkey and Cyprus over the eastern Mediterranean, responsible for the extension in between. In other words Turkey is relatively moving apart from Greece toward the northeast in the absolute reference frame, and not converging. This extension may or may not be coeval with compression elsewhere, and the related normal faults and shear zones should flat out in the decollement planes at base of the lithosphere. Similar “backarc” extension could be classified the Andaman Sea rift, in the hangingwall of the western Indonesia subduction zone, in the hinge where the arc advances southwestward faster than Asia over the Indian plate.

Figure 11. Cross-section



Cartoon showing that Greece lithosphere (B) is overriding Africa fixed (A) faster than Anatolia (C), generating extension between B and C. See a reference trace of the section in the previous figure. The extension in the ‘backarc’ is due to differential velocity of the hangingwall lithosphere. The African slab should be folded by the isostatic uplift of the mantle in the rift. This should results in a sort of window in the hangingwall lithosphere that is splitting apart into two independent plates, i.e., Greece and Anatolia, and it would ipothetically be coupled with a sort of horizontal window in the underlying stretched slab, allowing melting and uprise of OIB basalts. These kinematics should be envisaged in a 3D view, with the previous figure, where the faster rollback of the Hellenic trench with respect to Cyprus-Anatolia is determining a stretching of the slab also in a map view. The subduction zone migrated southwestward, and it was replaced by the extension. This is coherent with first the emplacement of the calc-alkaline rocks, and later the OIB basalts in the western Anatolia and the central-northern Aegean Sea.

Since the Aegean backarc basin developed in the hangingwall of a northeast directed subduction, it has been used as a case where the theory that backarc basins form only in the hangingwall of west-directed subduction zones (Doglioni et al., 1999) fails. It has rather been considered as an example of slab retreat faster than the convergence rate, determined by the slab-pull (Royden, 1993). However, from the aforementioned issues, it can be argued that the Aegean rift is not a classic backarc basin. In fact, one of the most debated type of rift are the so-called "backarc basins" related to E-NE-NNE-dipping subductions (e.g. the Aegean Sea or the Indonesia backarc). In this paper they are considered to have a different origin from backarc basins due to west-directed subduction zones because of the following points: a) They have lower rates of opening and subsidence with respect to backarcs related to west-directed subduction zones; b) They are often characterized by thick continental crust in spite of longstanding subduction, while in the opposite subduction, backarcs more frequently experienced fast generation of new oceanic crust; c) They may be inactive during contemporaneous subduction, while this does not occur for west-directed subduction zones; d) They have opposite polarity of opening toward WSW or SSW; e) They often are inverted in compressional regime.

Unlike the Apennines or other accretionary prisms related to west-directed subduction zones, the Hellenic arc and the related Mediterranean Ridge (the accretionary prism, e.g., Camerlenghi et al., 1995) have a low dip foreland monocline (almost flat, Clément et al., 2000), deep metamorphic rocks involved in the orogen, and higher structural and a morphologic elevation. Moreover, the slab has regularly low dip and it is shallow.

In the west-Pacific backarc basins (e.g., Honza, 1995), or in the Apennines, Carpathians, Barbados, Sandwich and Banda arcs subduction zones (Doglioni et al., 1999), the asthenosphere replaced the retreated subvertical slab (type 1 of extension in convergent settings, Doglioni, 1995). Due to the low dip of the Hellenic slab, in the Aegean Sea, the hangingwall has not enough space for a thick asthenospheric wedge like in W-Pacific or Apennines subduction zones. The hangingwall and footwall lithospheres are almost stacked one on top of the other, with thin (if any) sandwiched asthenosphere in between. Therefore the Aegean Sea represents a different type of extension associated to a subduction zone, where the hangingwall plate overrode the slab at different velocities, implying internal deformation (type 6 of extension), as shown in Figure 10 and Figure 11.

Considering the "east-northeastward" mantle flow indicated by the hot spot reference frame (Ricard et al., 1991), the origin of backarc basins due to E-NE-directed subduction should not be due to lithospheric disappearance as found in west-directed subduction zones. It could rather be related to differential drag of the lithosphere in the hangingwall of the subduction (Figure 11), due to different viscosity contrasts generated by lateral heterogeneities that control the amount of decoupling at the interface between lithosphere and asthenosphere. The system is composed by three plates (A-B-C, Figure 11), in contrast with back-arc basins due to W-directed subductions where the backarc may develop with two converging plates or even within one single plate.

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