On the geodynamics of the Aegean rift

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Abstract

The Aegean rift is considered to be either a classic backarc basin, or the result of the westward escape of Anatolia, or the effect of a gravitational collapse of an over-thickened lithosphere. Here these models are questioned. We alternatively present a number of geodynamic and magmatic constraints suggesting a simple model for the genesis of the extension as being related to the differential advancement of the upper plate faster or slower than the trench (Doglioni et al., 2009). In the first case a double-verging compressive belt forms, whereas in the second case backarc spreading prevails. More recently, it was inferred that extension in the Aegean area link the extension with the westward extrusion of the Anatolia plate (e.g. McKenzie, 1972; McClusky et al., 2000) or with a post-orogenic collapse combined with slab retreat (e.g. Gautier et al., 1999; Jolivet, 2001). However, all subduction zones have the trench retreating toward the lower plate; they differentiate whether the upper plate is advancing toward the lower plate faster or slower than the trench (Doglioni et al., 2009). In the first case a double-verging compressive belt forms, whereas in the second case backarc spreading prevails. More recently, it was inferred that extension in the Aegean–West Anatolia is associated to a differential advancement of the upper plates (Greek and Anatolian microplates) over the Africa plate (Doglioni et al., 2002). Here we describe this model in more detail and in the light of some new geodynamic and geochemical evidences.

1. Introduction

The formation and evolution of backarc extensional basins constitutes a crucial point in the comprehension of converging plate margins. Recently, it was proposed that the subduction zones are asymmetric as a function of the subduction polarity (e.g., Doglioni et al., 1999, 2007). The polarity is here defined as the main direction of the subduction with respect to the undulated flow of plate motions (Crespi et al., 2007), overall directed toward the “west” (Scoppola et al., 2006). In this model of plate tectonics, the westward drift of the lithosphere facilitates backarc spreading only in the hangingwall of the west-directed subduction zones. However, in some cases, as the Aegean and the Andaman regions, backarc basins are located in the hangingwall of NE-directed subduction zones, questioning this global model. Therefore the origin of the Aegean rift seems at odds with a global polarization of tectonics, and its origin might become relevant for testing the global tectonic pattern. Indeed, in spite of the relatively modest extensional rate affecting the Aegean–Anatolia region and the consequent limited crustal thinning, the Aegean rift has been interpreted as a backarc spreading (e.g., Le Pichon and Angelier, 1979; Horvath and Berckhemer, 1982). Other models of the Aegean area link the extension with the westward extrusion of the Anatolia plate (e.g., McKenzie, 1972; McClusky et al., 2000) or with a post-orogenic collapse combined with slab retreat (e.g. Gautier et al., 1999; Jolivet, 2001). However, all subduction zones have the trench retreating toward the lower plate; they differentiate whether the upper plate is advancing toward the lower plate faster or slower than the trench (Doglioni et al., 2009). In the first case a double-verging compressive belt forms, whereas in the second case backarc spreading prevails. More recently, it was inferred that extension in the Aegean–West Anatolia is associated to a differential advancement of the upper plates (Greek and Anatolian microplates) over the Africa plate (Doglioni et al., 2002). Here we describe this model in more detail and in the light of some new geodynamic and geochemical evidences.

2. Geological outlines on the Mediterranean basin

The Mediterranean basin consists of different domains formed during the Mesozoic to Present interactions between Eurasia and Africa (Fig. 1). Indeed, the Mediterranean lithosphere is made up of: (i) remnants of the Mesozoic Tethys Ocean subducted from the Cretaceous to the Present as a result of the Africa–Eurasia convergence and collision system (central-eastern Mediterranean); or (ii)
Cenozoic lithosphere formed in backarc regions of some of the numerous Eurasia–Africa subduction systems developed in the last 60 Ma (western Mediterranean). Currently, four subduction systems are active in the Mediterranean realm (Carminati and Doglioni, 2004), showing different convergence rates and polarity (Fig. 1A).

The western Mediterranean mainly consists of basins developed in the last 30–40 Ma. The progressive southeast-ward retreat of the Apennines–Maghrebides subduction system led to the development of the Provençal, Valencia, Alboran, Algerian, and Tyrrenian basins (Boccaletti et al., 1974; Carminati and Doglioni, 2004). These basins...
can be regarded as typical backarc basins and are characterized by marked crustal thinning and subsequent formation of new oceanic crust. Instead, the Eastern Mediterranean is basically a relic of the Neotethys (e.g., Garfunkel, 2004) whose original passive margins are still preserved to the east and to the south in the Levantine and Herodotus basins (Fig. 1A). To the north, the Eastern Mediterranean is characterized by the northeast-directed subduction of the African plate under the Crete and Cyprus trenches. It is noteworthy that, whereas thinner oceanic lithosphere is currently subducting under Crete, a thicker continental margin is subducting into the Cyprus trench (e.g., Robertson, 1998). Unlike the Western Mediterranean, the only rift of the Eastern Mediterranean formed in a “backarc” setting is the Aegean Sea. This basin formed by thinning of a variety of tectonic units, which were mainly emplaced during the Upper Cretaceous–Paleocene convergence–collision processes (e.g., Boccaletti et al., 1974; Robertson et al., 1991). These units are continental massifs such as the Serbo-Macedonian-Rhodope, the Pelagonian–Attic-Cycladic, and the external Hellenides that are connected by ophiolitic belts such as the Vardar and Sub-Pelagonian-Pindos (Fig. 1B). It is noteworthy that: 1) the Hellenic subduction system was active since at least the Late Cretaceous and the “backarc” rift developed later; and 2) despite the long-lasting formation of the Aegean basin (~40 Ma), the extension rate is relatively low, so that no oceanic crust was generated. This area is currently undergoing a widespread regional extension, which can be dated back to the Eocene–Early Miocene (e.g., Seyitoğlu and Scott, 1996; Gautier et al., 1999; Jolivet, 2001; Doglioni et al., 2002). Western Anatolia has a structural architecture very similar to the Aegean region. Indeed, the Sakarya massif is considered to be analogous to the Rhodope massif (e.g., Sengör and Yilmaz, 1981); the Izmir–Ankara zone is similar to the Vardar ophiolite belt; the Menderes massif can be correlated to the Pelagonian massif, whereas the Lycian nappes could represent the eastward projection of the Pindos ophiolites (Fig. 1B) (e.g. Robertson et al., 1991; Stampfli, 2000).

3. Geophysical constraints on the slab geometry

The subduction of the African plate under the Crete trench is marked by seismic and volcanic belts, which display a well-defined Benioff plane extending down to about 160–180 km (USGS Earthquake Catalogue, http://neic.usgs.gov/neis/epic/, Konstantinou and Melis, 2008, Fig. 1B). Earthquake focal mechanisms indicate down-dip extension in the slab (Papazachos et al., 2005). The angle of the subduction, calculated according to the earthquake hypocenters distribution is about 16°, and tends to flatten beneath NE–SW cross-sections, that is the direction of the subduction parallel to the relative plate motions) (Papazachos et al., 2005). The slab dip increases (up to 30–40°) moving southward where the subduction is oblique or in lateral ramp. Shear wave splitting analysis of SKS phases reveals the occurrence of an anisotropic fabric of the lithospheric mantle in the Central-Northern Aegean, suggesting a pure shear extensional mechanism involving the whole lithosphere (Kreemer et al., 2004).

The direction of this anisotropy (Hatzfeld et al., 2001) is NE-trending in the Central and Northern Aegean and Western Anatolia. However it is worth noting that, closer to the convergent margin, its direction is rotating, from NW in the Peloponnesus to roughly ENE close to Rhodes (see Figs. 2 and 6 of Kreemer et al., 2004). Furthermore, low S-wave velocities in the mantle beneath North-Central Aegean found by Bourova et al. (2005), suggest the occurrence of a low-velocity channel at relatively shallow depths (around 100 km), compatible with a flat slab beneath the overriding Aegean rift (Fig. 2), matching the slab shape drawn according to hypocenter locations (e.g., Christova and Nikolova, 1993). The tomography of the area shows a steep high velocity body, which is usually referred as the evidence for the Hellenic slab (Fig. 3A). However, the seismicity has a very different trend, pointing for a less inclined, shallower and shorter slab (Fig. 3B). The misalignment between tomography and slab-related seismicity highlights a fundamental problem, i.e., question on the real nature of the high velocity body.

In the entire Aegean–Western Anatolian region, no earthquakes deeper than 180 km occur (http://neic.usgs.gov/neis/epic/), and only shallow to intermediate earthquakes are present in the Central and Northern Aegean. These events have maximum depths of 55–60 km, well inside the lithosphere of the upper plate. The slab depicted by seismicity does not correspond to the slab imaged by tomography. We prefer to rely on the geometry of the slab detected by the seismicity (e.g., Christova and Nikolova, 1993; Papazachos et al., 2005) rather than to consider the tomography (e.g., Piromallo and Morelli, 2003), which is based on a mantle velocity model. The absence of deep seismicity contradicts the presence of a deep slab since the few cm/yr convergence rate and the oceanic nature of the down-going slab should rather favor a deep seismicity.

Most of the authors interpret faster bodies detected by tomography as colder bodies, i.e., subducted slabs (e.g., Wortel and Spakman, 2000; Piromallo and Morelli, 2003). However seismic waves speed variations may depend by some other factors, such as chemical heterogeneities in the mantle, or phase changes (e.g., Trampert et al., 2004). Recently, as an alternative hypothesis, Doglioni et al. (2009) proposed a mechanism able to generate the ghost of a slab along E- or NE-directed subduction zones; deeper mantle portions, more rigid and with faster seismic velocities with respect to the shallower mantle, may be upraised in response to the slab suction. This would generate volumes of faster mantle much wider than the usual slab thickness and it could explain the absence of seismicity below 200 km in the Aegean realm because there the slab would be missing.

According to this view, we interpret the faster body detected by tomography not as a real subducting lithosphere, but as a volume of uplifted mantle sucked up by the slab that is moving southwestward relative to the mantle, a direction opposed to the dip of the subduction.

Fig. 2. (A) S–N cross-section of upper mantle structure inferred from S-wave speed perturbations along the Aegean region (redrawn from Bourova et al., 2005; no vertical exaggeration). Blue portion represents zone with positive anomalies, and is interpreted as the subducted slab, whereas red portions represent negative anomalies. (LVZ = Low velocity zone; NAF = North Anatolian Fault). (B) 3D reconstruction of the Hellenic slab based on seismicity (after Christova and Nikolova, 1993). Location of the sections in Fig. 1.

Please cite this article as: Agostini, S., et al., On the geodynamics of the Aegean rift, Tectonophysics (2009), doi:10.1016/j.tecto.2009.07.025
4. Kinematic constraints on plate motions

The active tectonics of the Aegean region has been the subject of many analyses, both concerning its structural setting and its geophysical characteristics (e.g. Jackson, 1994; Gautier et al., 1999). Given the occurrence of normal faults and graben systems throughout the Aegean–Western Anatolian region (e.g. McKenzie, 1972; Yilmaz et al., 2000; Kreemer et al., 2004) and the focal mechanisms of crustal earthquakes, all these papers agree on the fact that the region is currently undergoing extensional tectonics.

A significant improvement in the description of this extension was recently provided by geodetic studies, which allow the reconstruction of detailed plate kinematics based on GPS-data (e.g., Le Pichon et al., 1995; Reilinger et al., 2006). The horizontal velocities measured in the Eastern Mediterranean area evidence the occurrence of a counterclockwise rotation of a broad region relative to Eurasia, from the Arabian plate, to the Turkey, and to the Aegean, at rates in the range of 20–30 mm/yr (Reilinger et al., 2006). In this reference frame, fixed to Eurasia, these authors note that the values in the velocity field increase toward the Crete trench (Fig. 1A) system, stressing the occurrence of extension in the Western Anatolia–Aegean system, with an average NW–SE trend (Fig. 4). However all sites used in this reconstruction pertain to different and disrupted tectonic units, or even independent deforming plates (Arabian, Anatolian, Greek) and do not constrain the rotation of a coherent plate. It is noteworthy that all the plates around the Eastern Mediterranean (i.e. Eurasia, Arabia and Africa) are moving northeast-ward in the ITRF (International Terrestrial Reference Frame), with respect to the hypothetic center of the Earth (e.g., Helffrich et al., 2008). In a hotspot reference frame relative to the mantle, all those plates rather move in the opposite direction (west-ward or southwest-ward) (Gripp and Gordon, 2002).

Given that the most prominent factor in shaping the Eastern Mediterranean area is the subduction of Africa plate underneath the Crete and Cyprus trenches (Fig. 1A), Doglioni et al. (2002) suggested that it could be useful to analyze plate movements keeping the African plate fixed, rather than Eurasia. In this reference frame, the GPS sites located in the Aegean area exhibit faster southwest-ward velocities with respect to the sites in northern Greece, and adjacent areas of the Balkan Peninsula, Western Anatolia, and Cyprus. This means that the region surrounding northern Greece and Western Anatolia are acting as separate microplates and override Africa with different velocities, being the Aegean area the diffuse transfer zone of separation between the two (Fig. 5).

Fig. 3. Tomographic section (A) and seismicity (B) across the Hellenic subduction–Aegean rift (after Piromallo and Morelli, 2003; Papazachos et al., 2000, respectively). White circles in (A) are earthquake hypocenters (M ≥ 5), black box represents the area of section (B). Note the different scales between (A) and (B) (1000 and 200 km depth, respectively). There is a relevant discrepancy between the high velocity body in the tomographic section, usually interpreted as the deep prolongation of the slab, and the seismicity. The earthquakes depict a less inclined and much shorter slab, as also described by shear waves (Fig. 2). Since the accuracy of seismicity is finer than tomography, we infer a possible alternative origin for the high velocity body: the slab is moving SW-ward relative to the mantle and is sucking up the underlying faster mantle. The slab is suffering down-dip extension, compatible with the model.

Fig. 4. GPS-velocity field (Eurasia fixed). Redrawn after Reilinger et al. (2006). Circles represent 1σ velocity uncertainties; circles without arrows represent GPS-stations whose velocity is negligible (lower than error). Note that the pole of rotation of the site arrows might be apparent since the sites are located on different tectonic settings and microplates.

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Moreover, in the hotspot reference frame (Gripp and Gordon, 2002), the Hellenic slab, which is attached to the African plate, is moving westward relative to the mantle (see Figs. 22 and 23 in Doglioni et al., 2007). Therefore it is moving in a direction opposed to the subduction, suggesting that it actually moves out of the mantle. The subduction is active and continues to retreat relative to Africa because the upper plate (Greek lithosphere) is overriding Africa. The subduction is active and continues to retreat relative to Africa because the upper plate (Greek lithosphere) is overriding Africa. Therefore it is moving in a direction opposed to the mantle (see Figs. 22 and 23 in Doglioni et al., 2007). In this interpretation, unlike the west-directed subduction zones that generate a corner flow in the mantle wedge (e.g., Turcotte and Schubert, 1982), the Hellenic type setting in which the slab is supposed to upraise, it should rather generate a suction flow from the underlying mantle (Doglioni et al., 2007). Such an interpretation could explain: i) why the tomographic images show relatively fast velocities along the ideal prolongation of the slab (e.g., a volume of an uplifted underlying mantle sucked up by the slab); ii) the anomalous thickness of the fast velocity body imaged by tomography (several hundred kilometers versus the real 70–100 km thickness of the subducted lithosphere).

5. Why the Aegean cannot be considered a typical backarc

In the Central Mediterranean area, two subduction zones have a still active volcanic arc, the Aeolian and the Aegean, and both are characterized by the formation of a backarc basin, the Tyrrhenian Sea in the hangingwall of a West-directed subduction, and the Aegean Sea in the hangingwall of a NE-directed subduction (Fig. 1A). The southern Tyrrhenian Sea, along with the Provençal and Algerian basins are floored by post 20 Ma oceanic crust, formed in the hangingwall of the eastward retreating Apennines–Maghrebides slab. In contrast, the Aegean basin has some peculiar physiographical and morphological characteristics. The first peculiarity is that the stretching is quite limited when compared to the duration of the subduction (Late Cretaceous, ~60–90 Ma). In fact, the crust is still continental and thicker than 20 km (Makris, 1978; Makris et al., 2001), whereas the backarc basins usually experience fast oceanization contemporaneous with the subduction. For example, the Provençal and Tyrrhenian basins are eastward migrating and rejuvenating backarc basins formed contemporaneously in the hangingwall of the Oligocene to Present Apennines subduction zone (Carminati and Doglioni, 2004). The Aegean rift developed mostly after the emplacement of Hellenic–Taurides nappes (e.g., Jolivet et al., 1994; Jolivet, 2001).

In addition, the Hellenic subduction shows a number of characters that fall into the E– or NE-directed class of subduction zones that include: the shallow dip of the foreland monocline at the base of the accretionary prism in the Mediterranean Ridge (0–2°) (Lenci and Doglioni, 2007); the shallow depth of the seismicity (mostly shallower than 200 km) coupled with the shallow dip of the slab (see above: Christova and Nikolova, 1993); the down-dip extension of the intra-slab seismicity (Papazachos et al., 2000, 2005). Moreover in the Aegean area, there are scattered high-pressure rocks in the thrust sheets of the Hellenides that are disrupted by the normal faults associated with the rift. This type of rocks requires deep seated thrusts cross-cutting the whole crust and possibly the upper mantle, a tectonic setting that typically occurs only along the E– or NE-directed subduction zones (Doglioni et al., 1999). In the opposite W-directed subduction setting, these rocks may occur as scattered relics, but they are inferred to be remnants of pre-existing E– or NE-directed subduction zones (Doglioni et al., 1999). Moreover, extension does not take place only in the backarc area, but is also present in the forearc (e.g., between Crete and the active Aegean arc, Fig. 1B) and in the island of Crete (Papanikolaou and Vassilakis, 2008; van Hinsbergen and Meulenkamp, 2006).

At least three different models have been proposed in the literature to explain the Aegean–Western Anatolia extension:

(i) Some authors consider the Aegean basin to be a typical backarc basin (Fig. 6A) and link extension to slab retreat and steepening (e.g., Berckhemer, 1977; Le Pichon and Angelier, 1979). In this view, the engine driving the extension is both the “margin push force” and the “slab pull force”. This model requires a progressive steepening of the African subducted slab and has recently gained a lot of popularity because of tomographic images showing a steep slab sinking down all through the upper mantle to the mesosphere (e.g., Wortel and Spakman 2000; Piromallo and Morelli, 2003).

(ii) Since the early contribution of McKenzie (1972), another model explains extension in the Aegean–West Anatolia as a result of the Africa–Eurasia collision and the west-ward extrusion of Anatolia driven by Arabia indenter on Eurasia, and the lateral escape bounded to the north by the North Anatolian Fault (NAF, Fig. 6B). In this view, the extrusion spreads out toward the less constrained Ionian margin, and the subduction is seen as a consequence of the Anatolian escape.

Fig. 5. ITRF2000 velocity field computed with respect to HELW (Egypt, African plate), redrawn after Doglioni et al. (2002). Notice the faster southwest-ward motion of the Greek microplate relative to Africa with respect to the motion of the Anatolian microplate. The differential velocity between the two upper plates of the subduction system implies extension in the Aegean realm.
Anatolian escape would rather close the Aegean Sea. Model (iii) geodynamics (Turcotte and Schubert, 1982). Moreover, the westward decrease moving away from the energy source, that is the Arabian collision. Black arrows represent GPS-velocity pattern with respect to a fixed Eurasia – fixed Africa (see above) indicate that the Greek microplate is overriding Africa at the Crete trench towards the SW faster than the Anatolia microplate that overrides Africa at the Cyprus trench (Fig. 7). Thus, extension in the region can simply derive from this velocity pattern, which is responsible for the onset of a diffuse extensional margin between the two plates (Doglioni et al., 2002). Different velocities between the Greek and Anatolian microplates (Fig. 7) could be due to the fact that the Aegean subduction system is coupled to relatively fast southwest-ward migration of the Crete trench (e.g., about 4 cm/yr), whereas the Anatolian subduction is coupled to the slower slab retreat of the Cyprus trench (1 cm/yr: Doglioni et al., 2002). In other words, the hangingwall has a differential velocity, which has to be accommodated by rifting between the Greek and Anatolian microplates (Fig. 8).

This simple intra-upper plate velocity gradient is a different mechanism from the one that can be envisaged in the Tyrrenian Sea, where the backarc area is the loci where the asthenosphere replaces the lithosphere subducting and retreating beneath the Apennines. It is noteworthy that, in a cross-section, the Hellenic system is composed of three plates (Africa, Greek and Anatolian), whereas the Apennine subduction system involves only two plates, Africa (or Adria in the north), in the footwall and Europe in the hangingwall (Fig. 9). These different kinematic constraints and mechanisms are also supported by heat flow data (Hurtig et al., 1991) that are much higher in the Tyrrenian Sea (~200–250 mW) with respect to the Aegean sea (~100–120 mW). When compared to the W-directed Apennines subduction, the higher topography of Greece, in spite of the Aegean rift, better fits with the E–NE-directed subduction zones, as does the gravity signature of the Hellenic subduction system (Harabadia and Doglioni, 1998).

The subduction of the Ionian oceanic (?) lithosphere (de Voogd et al., 1992; Panza et al., 2003, 2007) should determine dehydration of the slab. The fluids are partly metasomatizing the mantle wedge and generating the magmatic arc. However, the large flux of lithosphere into the subduction should generate large amount of fluids, which remain entrapped in the asthenosphere (e.g. Peccerillo et al., 2008). These fluids are able to decrease the viscosity in the low velocity layer (Manea and Gurnis, 2007), hence triggering a faster decoupling between lithosphere and underlying mantle. The western portion of the subducting lithosphere is oceanic (the crust is 11–16 km thick, with a 4–6 km thick sedimentary cover, Makris and Stobbe, 1984; de Voogd et al., 1992), whereas its eastern portion is represented by a continental margin. Therefore the larger amount of fluids released by the Hellenic subduction with respect to the Anatolian subduction could explain the faster southwest-ward motion of the Greek microplate with respect to the Cyprus–Anatolia segment of the Africa plate subduction. This hypothesis is supported by the occurrence of a low-viscosity channel under the Aegean lithosphere, as evidenced by the low values of P-wave velocities (Papazachos et al., 2005) and absolute S-waves velocities (Bourova et al., 2005) (Fig. 2).

6. Aegean–Western Anatolian magmatism

6.1. Age distribution

Since the Paleogene and throughout the Neogene, the Aegean–Anatolian region has been characterized by several episodes of
magnitic activity (Figs. 10 and 11; Table 1). The older products are the Upper Eocene to Upper Oligocene calc-alkaline volcanics of Rhodope, Thrace and North–West Anatolia (e.g., Innocenti et al., 1984; Yanev et al., 1998; Altunkaynak and Genç, 2007), which are sometimes associated with intrusive bodies.

Since the Miocene, orogenic products are found throughout Western Anatolia and the Aegean, with a rough trend of younging southward (Fig. 11; Table 1). This geographic trend indicates that the slab has been retreating consistently (i.e., the subduction hinge was moving towards the African lower plate). The orogenic volcanic belt, which definitely constitutes the most abundant and widespread activity in the region, starts at the base of Early Miocene both in the northern Aegean and northwestern Anatolia (e.g., Aldanmaz et al., 2000; Innocenti et al., 2005 and references therein) and is matched in Western Anatolia by quasi-contemporaneous emplacement of plutonic bodies (e.g., Bingöl et al., 1982; Delaloye and Bingöl, 2000). To the south in the Central Aegean region, analogous widespread high-K calc-alkaline volcanism took place a few million years later, between the Early and Middle Miocene. Significantly younger are less abundant high-K calc-alkaline volcanics found in and south of the Cycladic and Menderes massifs (Figs. 10 and 1A), which are mostly Upper Miocene.

Along with the calc-alkaline volcanism, mainly anatectic S-type granites were emplaced inside the Cycladic and Menderes Massifs during the Middle to Late Miocene (e.g., Hetzel et al., 1995; Pe-Piper and Piper, 2002). Further south occurs the Pliocene–Quaternary South Aegean volcanic arc (Fig. 10), which can be dated back to Early Pliocene and it is still active (e.g., Francalanci et al., 2005 and references therein).

It is noteworthy that from the northern to the southern part of this region, activity starts at different times but is characterized by magmas sharing the same geochemical characteristics. They are all high-K calc-alkaline and range from basaltic andesite to rhyolite in composition, with basalts being virtually absent, and andesitic lavas and rhyolitic tuffs being the most abundant products. In addition, volcanic activity follows a similar geochemical evolution all over the region (Fig. 11; Table 1). The older high-K calc-alkaline activity is strictly followed by the emission of shoshonitic products, with a general K2O-increasing trend mirrored by abundance decrease of erupted products. In some places, this trend evolves to sporadic emission of younger ultra-K products that sometimes have lamproitic affinity (Innocenti et al., 2005).

In addition, at the beginning of Late Miocene the Central part of the Aegean–Anatolian region is the locus of a narrow belt of high-Mg andesitic products (Pe-Piper and Piper, 2002; Agostini et al., 2005). From Late Miocene onwards, alkali basaltic lavas were emplaced throughout the Aegean, Thrace and Western Anatolia. These basaltic may display either a K- or Na-alkaline affinity and are characterized by a roughly southward younging trend. They are commonly scattered, low-volume occurrences, except for the Kula volcanic field that covers an area of about 350–400 km² (Tokçäer et al., 2005).

6.2. Geochemical and isotope characters of the volcanic products

A number of recent studies (e.g., Francalanci et al., 1990; Aldanmaz et al., 2000; Alci et al., 2002; Innocenti et al., 2005; Agostini et al., 2007) were aimed at identifying the sources involved in the genesis of the volcanism in the region, as well as their evolution and the interactions between the different sources. The rocks belonging to the
calc-alkaline, shoshonitic and ultra-K associations have the typical geochemical and isotope features of supra-subduction rocks, since they are enriched in Fluid Mobile Elements (FME) with respect to High Field Strength Elements (HFSE), as shown by their high Ba/Nb and Rb/Zr ratios. They are also characterized by high $\text{^{87}Sr/^{86}Sr} \approx 0.707\text{–}0.710$, and low $\text{^{143}Nd/^{144}Nd} \approx 0.5122\text{–}0.5125$ ratios.

The Late Miocene to Pleistocene alkali basalts can be subdivided into two main groups because some of them are mildly SiO$_2$-undersaturated and have potassic affinity whilst some others are strongly undersaturated and sodic alkaline. Interestingly, the K-basalts have limited variability in major elements but have very high variability in trace element abundances, FME/HFSE ratios and Sr and Nd isotope ratios ($\text{^{87}Sr/^{86}Sr} \approx 0.704\text{–}0.708；\text{^{143}Nd/^{144}Nd} \approx 0.5124\text{–}0.5129$). In contrast, the sodic alkaline rocks show greater major element variation (especially those from Kula, Fig. 10B), ranging from basanites to phono-tephrites but have quite constant FME/HFSE, Sr and Nd isotope ratios ($\text{^{87}Sr/^{86}Sr} \approx 0.7032；\text{^{143}Nd/^{144}Nd} \approx 0.5129$). The geochemical and Sr–Nd isotope features of calc-alkaline, shoshonitic and ultra-K magmas in Western Anatolia point out that these magmas were generated from the same mantle domain, which has been interpreted to be a mantle wedge strongly modified by a subduction component (Innocenti et al., 2005). Differing enrichments of subduction-related components, with a minor role played by the assimilation of crustal materials, may explain the geochemical variability observed in these rocks.

The Na-alkaline basalts show the typical geochemical characteristics of intraplate (OIB-type) magmas and were sourced in the sub-slab asthenosphere. By contrast, the K-alkaline basalts derive from some kind of interaction between sub-slab melts and the mantle wedge. From Pliocene to present, arc volcanism related to Aegean subduction is taking place in the South Aegean (e.g. Keller, 1982; Pe-Piper and Piper, 2002). The mantle source of this calc-alkaline volcanism is considered to be the asthenospheric mantle wedge, as the case for the older Central Aegean–Western Anatolian activity. With respect to the Miocene calc-alkaline belt, the Pliocene–Holocene South Aegean arc exhibits a greater variation in Sr, Nd and Pb isotopes, as well as FME/HFSE ratios, in the less evolved rocks, which has been related to different amount of a metasomatizing component added to the mantle source (Francalanci et al., 2005).

In Fig. 12, Sr isotope ratios of Aegean–Anatolian volcanics are plotted against their age. Here we subdivided the region into three zones, from south to north. Using the $\text{^{87}Sr/^{86}Sr}$ ratio as an indicator of subduction component, we found that in all of the three zones there is a change from strongly subduction modified ($\text{^{87}Sr/^{86}Sr} \approx 0.708$) to unmodified mantle sources. It is noteworthy that, through time, rocks sharing the same mantle source (i.e., calc-alkaline, shoshonitic and ultra-K) mark a progressive lowering of the subduction signal. This is particularly evident for the rocks of Central Western Anatolia, where a more coherent dataset is present (Fig. 12B). In addition to the mantle-sourced magmas, anatectic granites and rhyolites characterized by
high \( ^{87}\text{Sr}/^{86}\text{Sr} \) ratios (0.709 to 0.715) were emplaced, especially in the areas experiencing a major amount of crustal thinning or extension (Arcian-Cycladic and Menderes massifs) (e.g. Altherr et al., 1982; Innocenti et al., 1982; Altherr et al., 1988; Hetzel et al., 1995).

Boron and lithium isotope studies contributed to clarify the role of the subduction-derived fluids in modifying the mantle sources involved in the magma genesis and their temporal evolution (T0narini et al., 2005; Agostini et al., 2008). A progressive lowering of \( \delta^{11}\text{B} \), as well as \( \delta^{7}\text{Li} \), throughout the time is observed and is matched by a progressive decrease of B/Nb ratio (Fig. 13). This fact points out that the shift from C-A to U-K rocks was due to the addition of a decreasing amount of metasomatizing fluids in the mantle wedge, from a progressively dehydrating slab. In particular, the ultra-K rocks were sourced in a mantle wedge left as residuum due to the previous extractions of calc-alkaline magmas, and slightly metasomatized by the last fluids coming from an almost dehydrated slab. The high \( \delta^{11}\text{B} \) variability of K-alkaline basalts, along with their limited variation in major elements and Li isotopes, and wider variations of FME/HFSE ratios, Sr–Nd isotopes, point out that these basalts did not result from mixing between intraplate-type and orogenic melts, rather they were formed by interactions between sub-slab asthenospheric magmas and residual slab fluids (T0narini et al., 2005; Agostini et al., 2008).

### 6.3. Magmatic constraints

Some fundamental constraints on the geodynamic evolution of the region can be gained by the time distribution, the geochemical and petrological features of the magmatism. These constraints can be schematically listed as follows:

(i) The northern, central and southern parts of the Aegean–western Anatolian region are characterized by similar evolution of magmatism, with a time shift from north to south (Fig. 11), implying that the same geodynamic setting progrades from north to south.

(ii) Especially in the central Aegean and western Anatolia, we observe an Early–Middle Miocene evolution from calc-alkaline to ultra-K magmas. The geochemical and isotope features of these rocks (see above) implies that they were generated after partial melting endured in the same mantle domain, suggesting the occurrence of a non-convective mantle wedge.

(iii) The progressive decline of slab released fluids coupled to an extreme B–Li negative signature point out that the slab was dehydrated up to almost being dewatered completely. The occurrence of such a process again may be indicative of a stagnant slab (i.e. very slowly sinking, dehydrated as a consequence of progressive thermal perturbation).

(iv) Li isotopes are easily re-equilibrated to common mantle values in the mantle wedge, so that most subduction-related magmas worldwide show no Li isotope variability. The persistence of high Li isotope differences, as well as the B–Li isotope correlation, in Western Anatolia orogenic rocks are thus indicative of a limited interaction between slab fluids and overlaying mantle, that is a very tiny, low-volume, mantle wedge (Agostini et al., 2008).

(v) The youngest Na-alkaline magmas were sourced in the sub-slab asthenosphere and suffered no interaction with slab fluids or any slab component. The genesis of such magmas is usually linked to asthenosphere partial melting after extensional dynamics and mantle upwelling. In addition, to allow such magmas to reach the surface without any contamination by slab material, the occurrence of a slab window is invoked.

(vi) The Late Miocene K-alkaline basalts resulted from interaction between sub-slab magmas (i.e. the intraplate-type magmas represented by the later Na-basalts) with residual slab fluids. This points out that, during the earlier stages of slab window opening, limited interactions between intraplate magmas and slab fluids occurred.

### 7. Summary and conclusions

The Western Anatolia–Aegean region since the Late Oligocene–Early Miocene was, and still is, affected by extensional tectonics taking place in a backarc position. The Aegean area is one of the few examples of a backarc basin in the hangingwall of a NE-directed subduction zone. This would contradict the model of a global polarization of tectonics, which claims that backarc basins form only in the hangingwall of W-directed slabs (e.g., Doglioni et al., 1999).
However, the Aegean rift has a number of different characteristics with respect to typical backarc. Indeed, it is diachronous with respect to the onset of the subduction that dates back at least to the Late Cretaceous. The Aegean rift is characterized by a low degree of stretching, so that no oceanic crust was generated through 40 Ma of rifting. Other structural features of the Aegean region that contrast with W-directed subduction systems and backarc basins are: i) the shallow dip of the foreland monocline at the base of the accretionary prism; ii) the shallow depth of the seismicity with an absence of earthquakes deeper than 180 km; and iii) a low subduction angle (≈16°) in the first 200 km of the subducted lithosphere which tends to flat northward beneath the rift (Fig. 2).

The Aegean rift is here interpreted as an atypical backarc basin, whose opening is not related to progressive slab rollback, steepening, and asthenospheric replacement as it can be inferred for example for the Tyrrhenian or the Western Pacific basins. Geodetic data, if

### Table 1

Geochronology and Sr isotopes data of volcanics from Aegean region.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Affinity&lt;sup&gt;+&lt;/sup&gt;</th>
<th>87Sr/86Sr Ref.</th>
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<td>29.5-27.5</td>
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<td>22, 23</td>
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<td>18</td>
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<tr>
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<td>Central Aegean–Central West Anatolia</td>
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<td>3, 19, 30</td>
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</table>

South Aegean–South West Anatolia

| S Aegean                 | Samos                   | 10.2-8.7 | HK Calc-alk           | 24             |
| S Aegean                 | Samos                   | 8.5-8.1  | Sho                   | 26, 27         |
| Aydın                    | Söke                    | 7.0      | HK Calc-alk           | 10, 16         |
| Bodrum                   | Bodrum                  | 10.6-10.5| Sho                   | 15, 27         |
| Bodrum                   | Bodrum                  | 8.1-7.9  | K Bas                | 26, 27         |
| S Aegean                 | Patmos                  | 7.2-6.2  | Sho                   | 13, 31         |
| S Aegean                 | Patmos                  | 4.4      | K Bas                | 31             |

References
computed relative to fixed Africa, indicate that the Greek microplate is overriding Africa southwest-ward faster than the Anatolia microplate (Figs. 7 and 8). This simple intra-upper plate velocity gradient is able to explain extension in the region and the onset of a diffuse extensional setting between the Greek and Anatolian microplates. The differential velocity of the Greek microplate with respect to Anatolia is accommodated by the faster southwest-ward migration of the subduction hinge along the Crete trench with respect to the Cyprus trench. The differential velocity could be explained by the larger dehydration of the Ionian oceanic slab, subducting all along the Hellenic trench, with respect to the eastern Levantine continental segment of the slab, subducting for example along the Cyprus trench. The west-ward Anatolia escape is an unsatisfactory physical interpretation because the velocity increases from east to west, indicating that the source of the energy cannot be the Arabia indenter alone. Moreover, the Hellenic slab (African plate) is moving west-ward relative to the mantle (i.e., in the hotspot reference frame) that is rather moving “out” of the mantle in a direction opposed to that of the subduction. Thus, the subduction is active because the upper plate moves southwest-ward faster than the lower Africa plate, pushing down the slab, and forcing the subduction hinge to migrate in the same direction. The intra-slab down-dip extension is compatible with a subduction zone attached to a surface plate (i.e. the African plate) that is moving in the direction opposed to the subduction (Dolgoni et al., 2007, 2009). This mechanism is feasible in the frame of the “westward” drift of the lithosphere relative to the mantle. The thickness of the continental upper plate (80–100 km), the shallow dip of the slab, and the source of the magmatism from a depth of about 100 km constrain the mantle wedge to be only continental lithospheric mantle or have only a thin underlying asthenospheric layer. However, the differential SW–SSW retreat of the subduction hinge (faster along the Greek segment with respect to the Anatolian one) should produce tear zones and windows in the slab. Such slab ruptures could allow sub-slab asthenospheric upwelling.

Our analysis of the complex magmatic activity in the region provided fundamental constraints to develop and test this model: (i) the orogenic activity exhibits a geochemical evolution from calc-alkaline, to shoshonitic, to ultra-K, pointing out the occurrence of progressive dehydration of the slab (extremely low B isotope values) and a very tiny mantle wedge (variable and low values of Li isotopes) (Figs. 13 and 14A); (ii) the later occurrence of alkali basalts sourced in the sub-slab testify for the occurrence of upwelling and melting of mantle asthenosphere. Moreover, the time shifting from K-alkali basalts, with residual slab imprinting, to OIB-type Na-basalts points out for a progressive influx of sub-slab mantle whose rising is considered due to the stretching of the subducted slab, with the formation of ruptures, or vertical slab windows, allowing such magmas to reach the surface (Fig. 14B).

The Hellenides and Taurides (Fig. 1) have been shown as polygenic orogens where a few microplates gradually docked, sandwiching the intervening oceanic regions (e.g., Pindos, Sub-Pelagonian, Vardar, Izmir–Ankara zone), supporting a very heterogeneous paleogeography and several lithospheric anisotropies as visible in Fig. 1B. For example, the lower K content in the Pliocene–Holocene South Aegean active volcanic arc could be related to a change in composition of the subducting slab such as from a stretched passive continental margin during the Miocene, giving rise to high-K calc-alkaline to shoshonitic magmas, to an oceanic slab during the Plio-Pleistocene (low- to medium-K calc-alkaline magmas). The differential slab retreat was
enhanced by the lower plate lithospheric variability, and kinematically required slab ruptures, which can explain the occurrence of alkaline Na-rich final magmatic products.

Acknowledgements

Randy G. Keller and an anonymous referee are acknowledged for their constructive and helpful comments. We thank M. Fytikas (Aristotle University of Thessaloniki, Greece), N. Koliós (IGME, Athens, Greece), and M.Y. Savaşçın (Dokuz Eylül University, Izmir, Turkey) for stimulating discussions during fieldwork. We are also indebted with G. Panza (University of Trieste, Italy) for stirring discussions on Eastern Mediterranean geophysical context.

F. Innocenti devoted his last days for revising the earlier version of this paper. In his enthusiastic conception of science this paper would not be the last one. For the co-authors, it has been a privilege to work with him for several decades until his very last days.

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