

## Article

# Late Cenozoic Evolution and Present Tectonic Setting of the Aegean–Hellenic Arc

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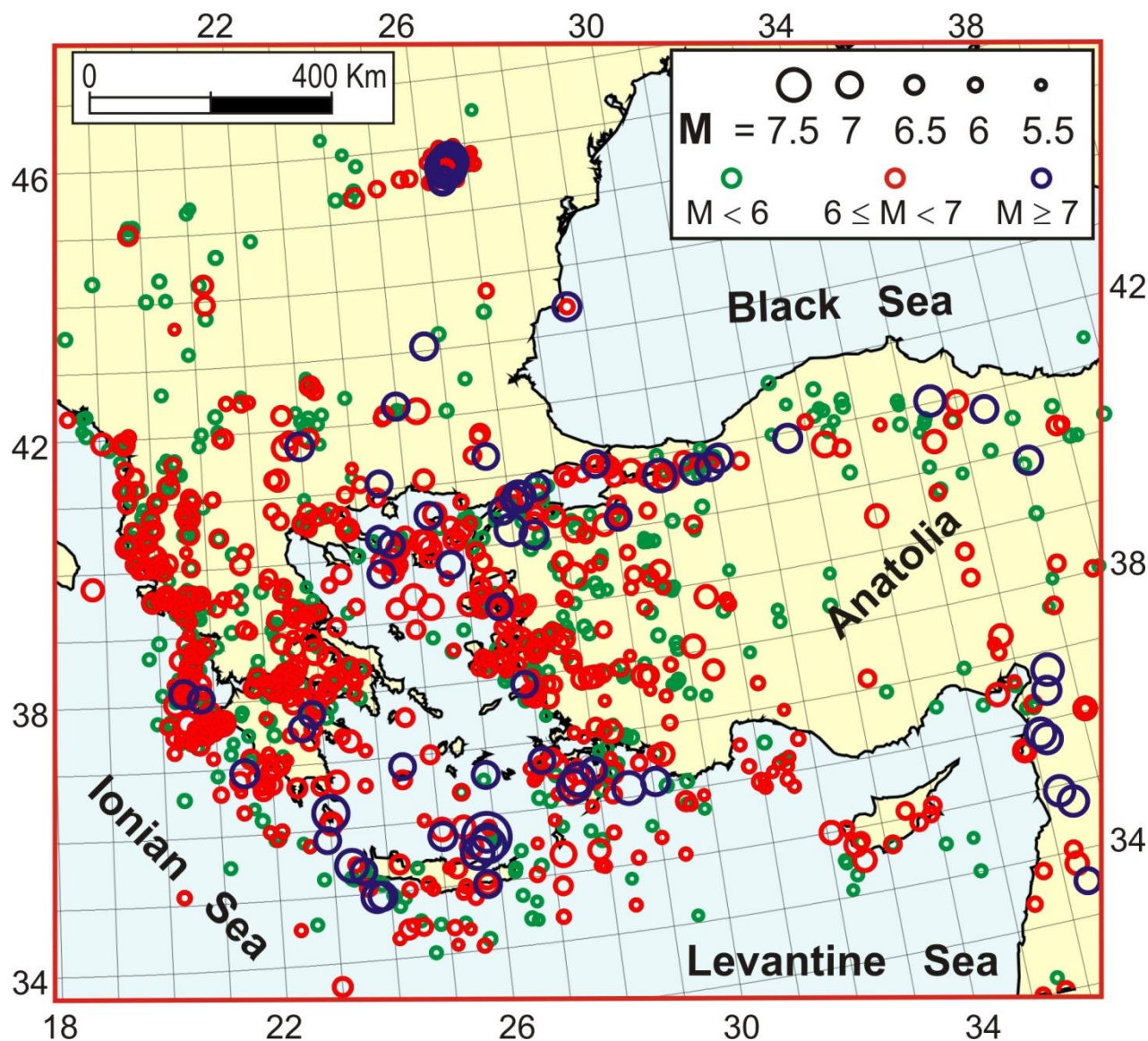
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**Abstract:** The Aegean–Hellenic arc is a deformed sector of a long heterogeneous orogenic system (Tethyan belt), constituted by an inner old metamorphic crystalline core flanked by younger chains of European and African affinity, running from the Anatolian to the Pelagonian zones. Due to the convergence between the Arabian promontory and the Eurasian continental domain, the Anatolian sector of that belt has undergone a westward extrusion, accommodated by oroclinal bending, at the expense of the surrounding low buoyancy domains. Since the late Miocene, when the Aegean Tethyan belt collided with the Adriatic continental promontory, the southward bowing of the Aegean–Hellenic sector accelerated, leading to the consumption of the Levantine and Ionian oceanic domains and to the formation of the Mediterranean Ridge accretionary complex. The peculiar distribution of extensional and compressional deformation in the Aegean zone has mainly been influenced by the different rheological behaviours of the mainly ductile inner core (Cyclades arc) and of the mainly brittle outer belt (Hellenic arc). The bowing of the inner belt developed without involving any major fragmentation, whereas the outer brittle belt underwent a major break in its most curved sector, which led to the separation of the eastern (Crete–Rhodes) and western (Peloponnese) Hellenic sectors. After separation, these structures underwent different shortening patterns, respectively driven by the convergence between southwestern Anatolia and the Libyan continental promontory (Crete–Rhodes) and by the convergence between the Cycladic Arc and the Adriatic continental domain (Peloponnese). A discussion is given about the compatibility of the observed deformation pattern with the main alternative geodynamic interpretations and with the Nubia–Eurasia relative motions so far proposed.

**Keywords:** Aegean arc; Tethyan belt; oroclinal bending; tectonic evolution

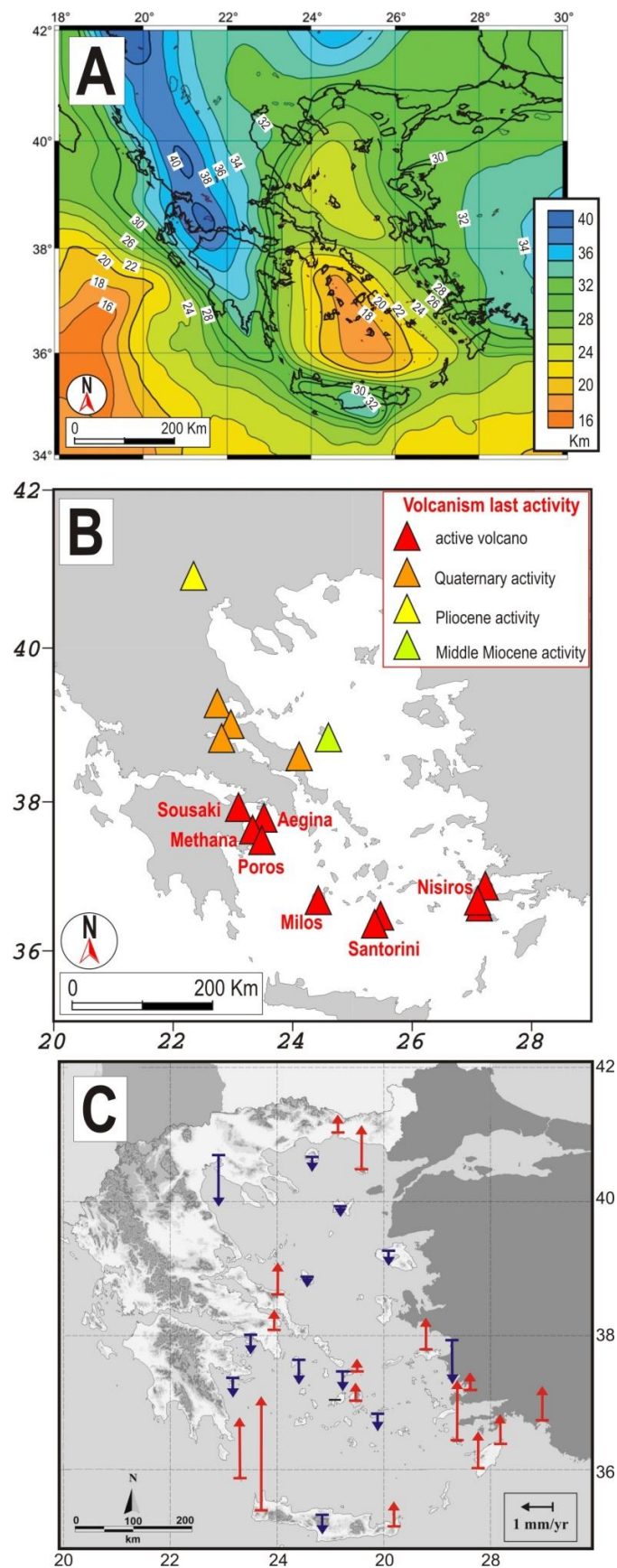
## 1. Introduction

The Aegean–Hellenic system is the most seismic zone of the Mediterranean area (Figure 1). Thus, there is a considerable interest in understanding which driving forces and tectonic processes are responsible for such activity. This basic knowledge could help recognizing the connection between the ongoing deformation pattern and the spatio-temporal distribution of major earthquakes, to be used for tentatively identifying the zones most prone to next major earthquakes (e.g., [1,2]).



**Figure 1.** Distribution of major earthquakes  $M \geq 5.5$  in the study area (since 1000 A.D.). Data from [3–16].

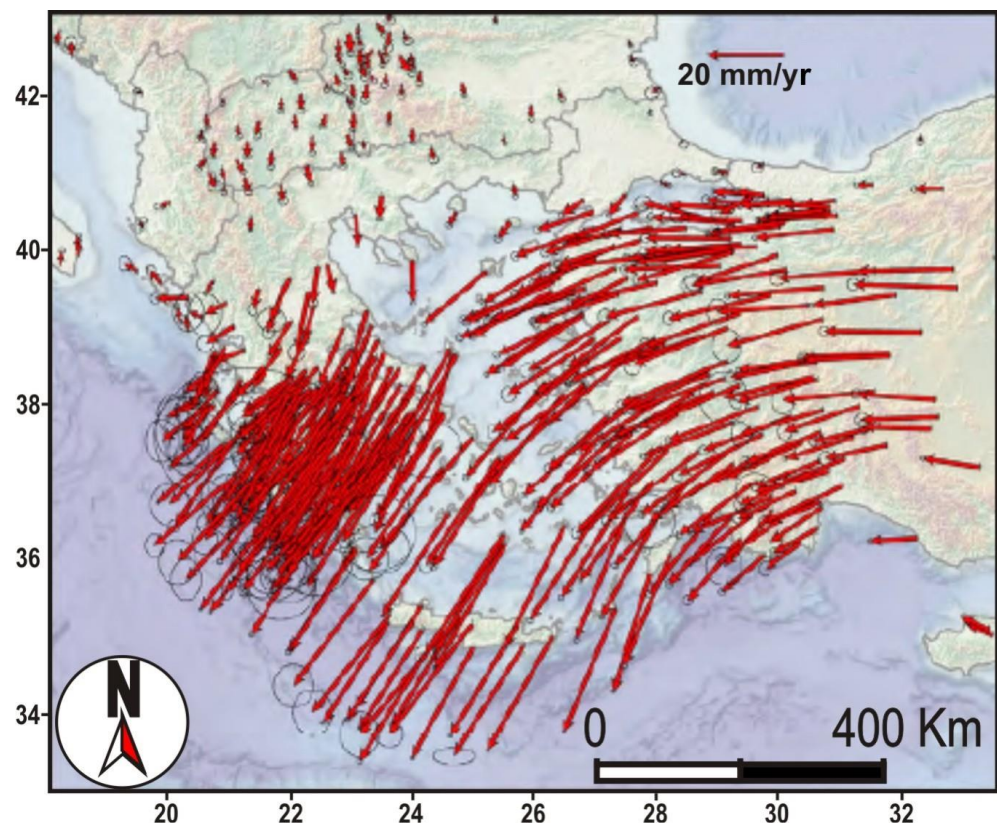
The Aegean region presents a complex distribution of zones characterized by different lithologies, structural features, morphologies, strain regimes and kinematic patterns. We separated the evidence relating to long-term deformation (Figure 2), mainly derived by geological or geophysical investigations, from the short-term deformations (Figure 3), inferred from geodetic and seismicity data. This was because the second kind of data may relate to the transient displacement and strain fields triggered (as postseismic relaxation) by the strong earthquake sequence that has developed along the entire North Anatolian Fault system (NAF) since 1939 [17–19].



**Figure 2.** Main long-term evidence in the Aegean region. (A) Map of the Moho depth modified after [20,21]. (B) Distribution of volcanic activity modified after [22]. (C) Vertical displacement rates in Aegean coastal zones during the Holocene assessed by geoarchaeological data (modified after [23]).

The map of the Moho depth (Figure 2A) indicates that the thinnest crust is observed in the Cretan basin, the Karpathos trough (15–20 Km) and the North Aegean trough (20–25 Km). The Cretan and Mykonos–Ikaria basins are deep troughs, where a relatively thin crustal basement is buried beneath a thick sedimentary cover (e.g., [24] and references therein). The thickest crust is observed in the Hellenides, along the Greek–Albanian zone. Seafloor bathymetry shows that the Cyclades sector is a shallow sea, with many islands where exhumed deep crustal rocks outcrop (e.g., [25] and references therein). Transtensional tectonics has occurred in the northern Aegean zone since the late Miocene–early Pliocene [26–28], while extension and subsidence have developed in the southern Aegean zone since the early Pliocene in the western Cretan basin, and since the Pleistocene in the eastern Cretan basin and Karpathos trough (e.g., [24,29]).

A generalized uplift has affected the Peloponnese and mainland Greece since the early Pliocene (e.g., [30] and references therein; [31]), with Holocene uplift rates running from 0.6 to 1.8 mm/yr (Figure 2C). The Adriatic zone beside the Hellenides was folded and uplifted in the Pliocene. The western part of Cephalonia and the Zakynthos islands experienced an uplift of at least several hundreds of meters in the Pleistocene [32]. Uplift rates of some mm/yr have been observed along the Hellenic Arc [22]. Crete underwent a ~1 km uplift between 5 and 3 Ma [33–35]. The analysis of the Plio-Quaternary deformation in the central Aegean area suggests that the Cyclades massif has been simultaneously affected by E-W compression and S-N extension (e.g., [36,37]).



**Figure 3.** Horizontal velocity field in the Anatolian Aegean area, defined by geodetic data (modified after [38]).

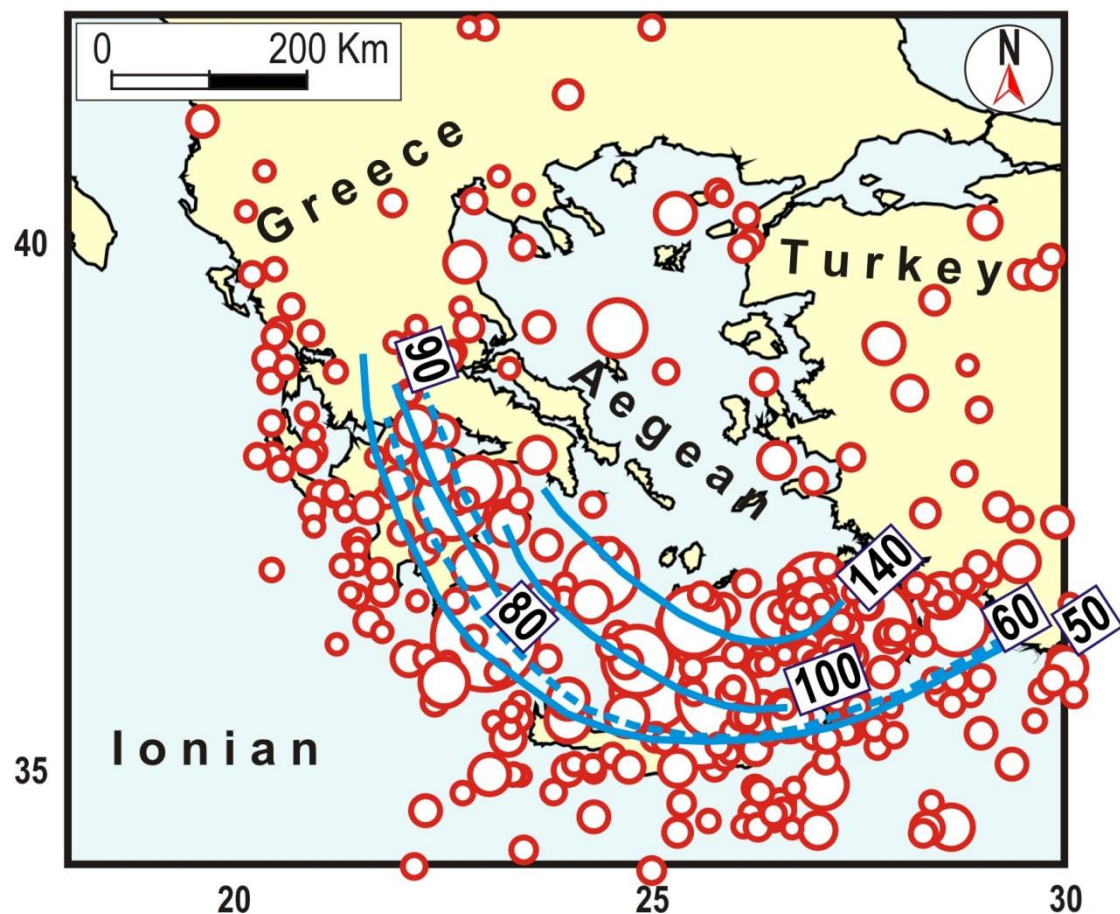
The present kinematic pattern (Figure 3), defined by geodetic observations (GPS, e.g., [38–40]) shows that Anatolia is moving roughly westward at 15 to 25 mm/yr relative to Eurasia, while the southern Aegean (and much of the Peloponnese) moves SW-ward at 30–35 mm/yr. A strong reduction in velocities occurs north of the North Aegean trough. In the southern and western Peloponnese, GPS data suggest 5 mm/yr of E-W extension. In central Greece, the extensional trend turns to S-N (5–20 mm/yr), being most likely

responsible for the generation of E-W troughs, such as the Corinth and Ambracique troughs (e.g., [21,41–43]). In southwestern Anatolia, the velocity field suggests a WNW-ESE extensional rate of 5 mm/yr.

In mainland Greece and western Anatolia, the analysis of focal mechanisms indicates an S-N extensional regime, associated with E-W compression (e.g., [38]). Strike-slip faulting is documented in the North Anatolian fault (NAF), while reverse faulting occurs in the Hellenic Trench, the Ionian Sea off western Greece and northwest into Epirus and Albania (e.g., [22]).

The strongest and most frequent seismicity (Figure 1) affects the entire Hellenic arc, with particular regard to the boundary between the Hellenides and the Adriatic plate. Intense activity also affects western Anatolia and the northern Aegean trough. The Cyclades massif in the central Aegean region is affected by minor crustal seismicity and a low strain rate (e.g., [38]).

The occurrence of subcrustal earthquakes under the Aegean zone (down to about 150 km) and the geometry of the Hellenic accretionary chain indicate the presence of subducted lithosphere. The shape of this body has been tentatively reconstructed by many authors [44,45]. A recent slab geometry, suggested by [46], is shown in Figure 4.

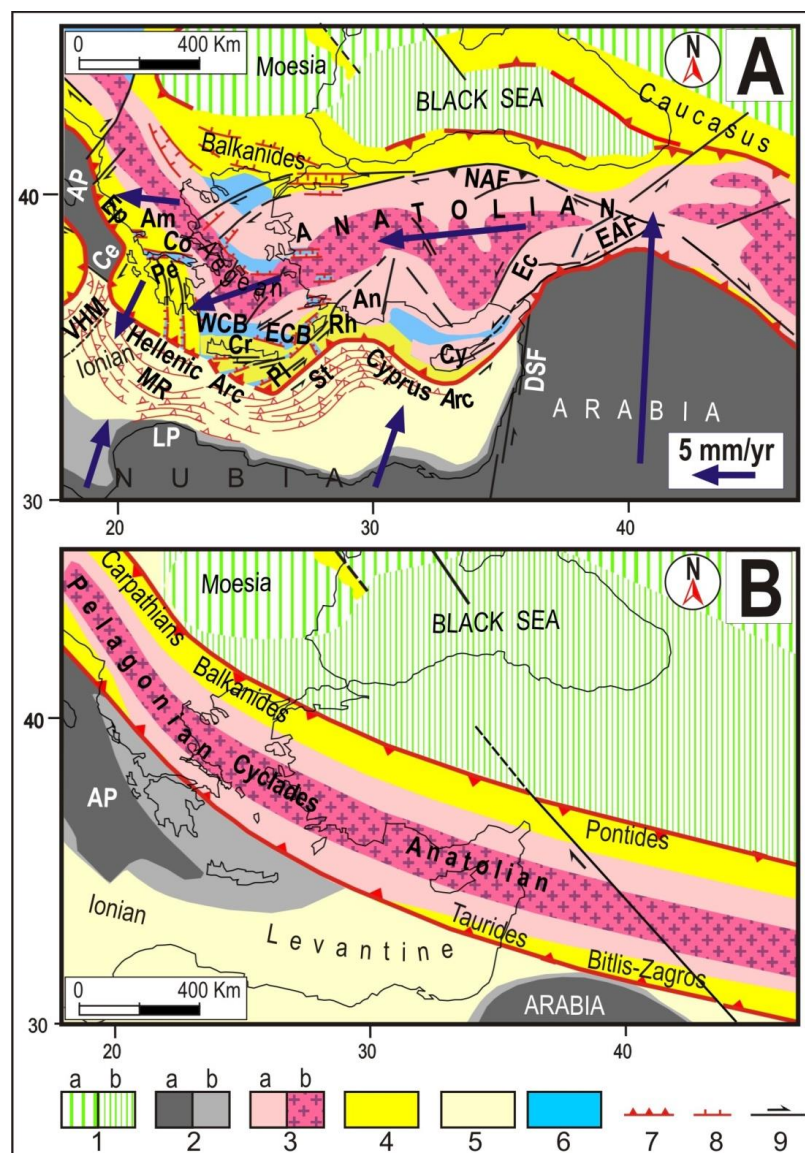


**Figure 4.** Distribution of subcrustal earthquakes ( $h > 30$ ) and tentative reconstruction of the Hellenic slab (isodepth lines from [46] and references therein).

The present configuration of the main lithological domains (Figure 5A) suggests that the Aegean zone is a strongly deformed sector of the long orogenic system that in the Oligocene was running from the Iranian–Anatolian to the Carpathian regions (Figure 5B). This system is constituted by an inner metamorphic crystalline belt, generated by the consumption of the

northern NeoTethys oceanic domain, flanked by two external neogenic accretionary belts of European and African affinity ([47–52] and references therein; [53–55]).

Some authors, on the basis of tomographic data, suggest that the subduction of the African margin (Ionian–Levantine domain) under the Aegean arc started in the Mesozoic (e.g., [56,57]), whereas a much later onset (middle Miocene to early Pliocene) of this consuming process has been proposed by other authors [46,58–62]. Most reconstructions of the Aegean area preceding the subduction suggest an almost rectilinear boundary [31,46,56,63–65], implying a subsequent bowing of the Aegean arc.



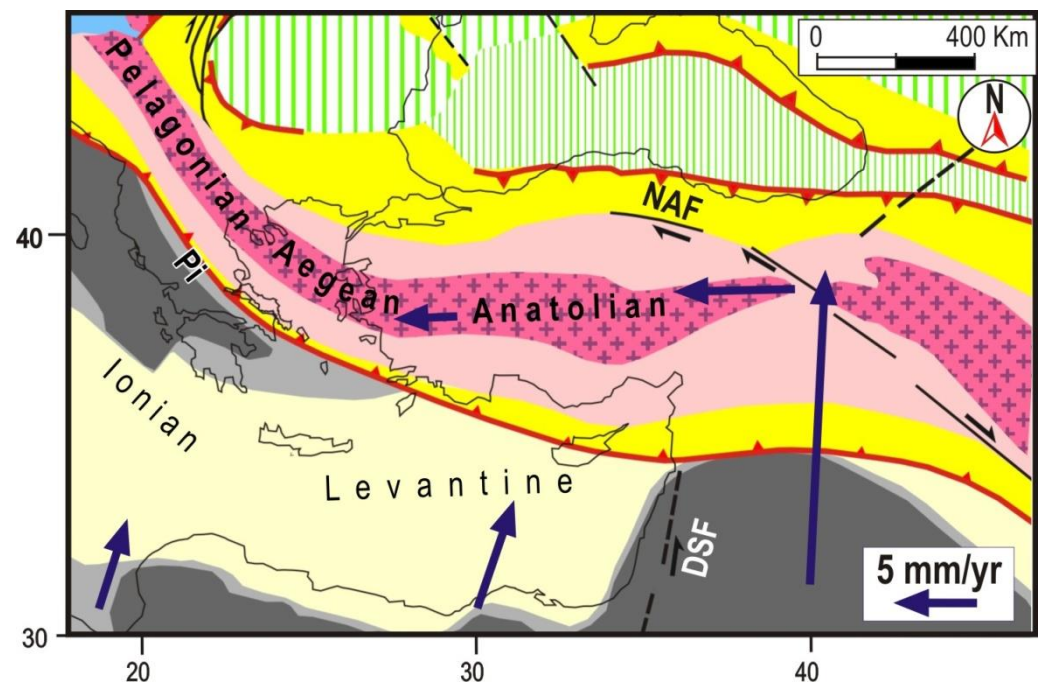
**Figure 5.** (A) Present tectonic setting. (1) Continental (a) and thinned continental (b) Eurasian domains. (2) Continental (a) and thinned continental (b) African, Adriatic and Arabian domains. (3) Inner core of the Tethyan belt, constituted by ophiolitic units (a) and crystalline massifs (b). (4) External accretionary belts with European and African affinity. (5) Ionian and Levantine oceanic domains. (6) Zones affected by crustal thinning. (7–9) Compressional, tensional and strike-slip features. The blue vectors indicate the long-term kinematic pattern [66,67]. The present shape and composition of the Tethyan belt was taken from [68–70]. Am—Ambracique trough; AP—Adriatic plate; An—Antalya; Ce—Cephalonia fault system; Co—Corinth trough; Cr—Crete; Cy—Cyprus; DSF—Dead Sea fault system; EAF—East Anatolian fault; Ec—Eceemis fault; Ep—Epirus; ECB—eastern Cretan basin; LP—Libyan promontory; MR—Mediterranean ridge; NAF—North Anatolian fault; Pe—Peloponnesus; Pl—Pliny fault; Rh—Rhodes; St—Strabo fault; WCB—western Cretan basin VHM—Victor–Hensen Medina fault system. (B) Tentative reconstruction of the Oligocene–early Miocene configuration of the Tethyan belt and surroundings.

## 2. Late Cenozoic Evolution

The shortening between the Arabian promontory and Eurasia started more than 30 Ma, accommodated by the closure of an ancient ocean, the Neotethys [71,72]. During this phase, a major NW-SE dextral transpressional fault decoupled the Anatolian sector of the Tethyan belt from the Iranian sector (Figure 5B, [48,51,73–75]. Some authors [76,77] suggest that the indentation of Arabia occurred much later (late Miocene–early Pliocene), supposing that such an event was responsible for the thrusting and uplift in the Bitlis suture zone. However, such a hypothesis cannot explain why, since the early Miocene shortening occurred north of the Bitlis zone, in the Pontides, Caucasus, Carpathians and Magura zones (e.g., [48,51,52,67,73,78–81].

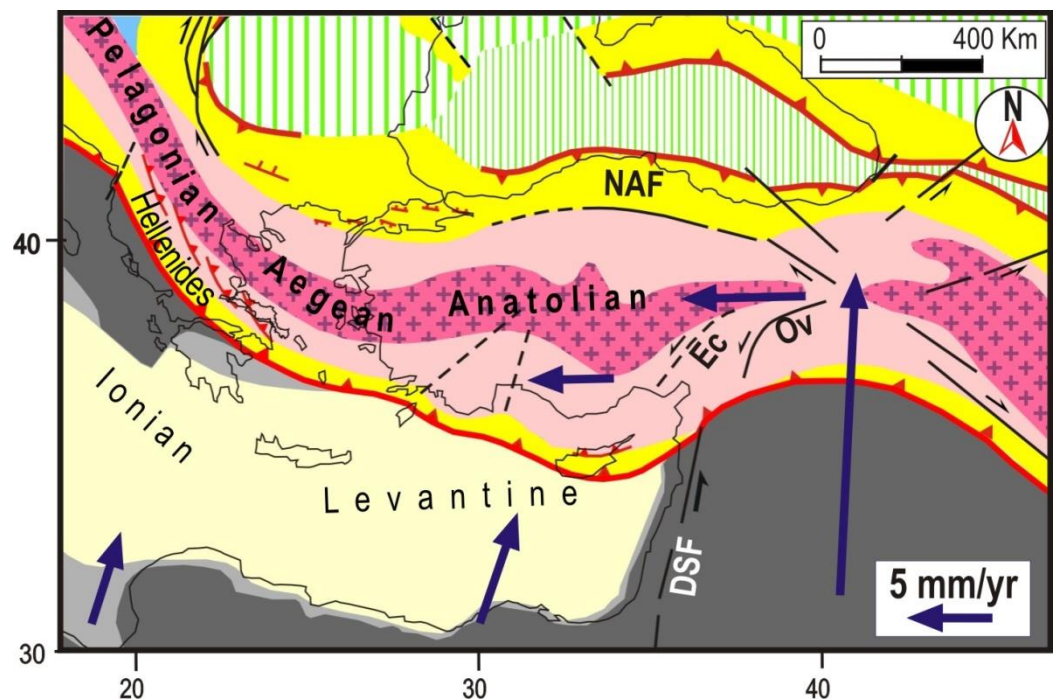
Once such thinned domains were mostly consumed, about 12–15 Ma, the Arabia–Eurasia convergence caused a considerable uplift and compressional deformations in the Tethyan belt, forming the Anatolian plateau (e.g., [76,82–84]. This shortening produced major palaeogeographic, sedimentological and tectonic changes [85,86], accompanied by volcanic activity [76,87], in eastern Turkey (Figure 6). Then, the gradual increase in resistance against any further shortening led to the activation of lateral decoupling faults, which allowed the eastern Anatolian wedge to extrude westward [48,51,73,74,81]. The onset of sliding along the North Anatolian fault (NAF) is mostly placed in the late Miocene, while this onset for the East Anatolian fault (EAF) is generally placed much later (late Pliocene, e.g., [88–90] and references therein). To tentatively explain this considerable mismatch between such onset times, [91] suggests that the activation of the EAF was preceded by the activation of other transpressional faults parallel to the EAF, but located more to the west (in particular, the Ovacik and Ecemis faults, Figure 7).

Geological observations suggest that in central Anatolia, the NAF activated around 8.5 Ma [89,90] and around 5 My in its western part, around the Sea of Marmara [27,92]. During this phase, the westward extrusion of Anatolia was accommodated by the roughly SW-ward bending of the western Anatolian–Aegean–Pelagonian Tethyan belt, at the expense of the Levantine, Ionian and Pindos thinned domains.



**Figure 6.** Middle Miocene paleogeographic setting. Present geographical contours (thin black lines) are reported for reference. The kinematics with respect to a Eurasian reference frame [66,67] are tentatively indicated by the blue arrows (scale in the inset). Pi—Pindos zone. Symbols, colors and other abbreviations as in Figure 5.

This context went on until the late Miocene–early Pliocene (Figure 7), when the western Aegean Tethyan belt collided with the continental part of the Adriatic domain, after the complete consumption of the thinned Pindos zone. Initially, this collision caused a shortening and uplift in both the colliding structures, evidenced by the inversion of previous normal faults in the Ionian domain during the Tortonian–late Messinian interval (e.g., [93]) and by a thrusting and uplift in the western Aegean zone (Central Greece, [94,95]). Once having consumed the thinned domains, the strong increase in E–W compression accelerated the southward oroclinal bending of the Aegean Tethyan belt, at the expense of the Levantine and Ionian oceanic domains (e.g., [96–98]). This consumption process led to the formation of the Mediterranean ridge accretionary complex along the outer boundary of the entire Hellenic arc (Figure 8).

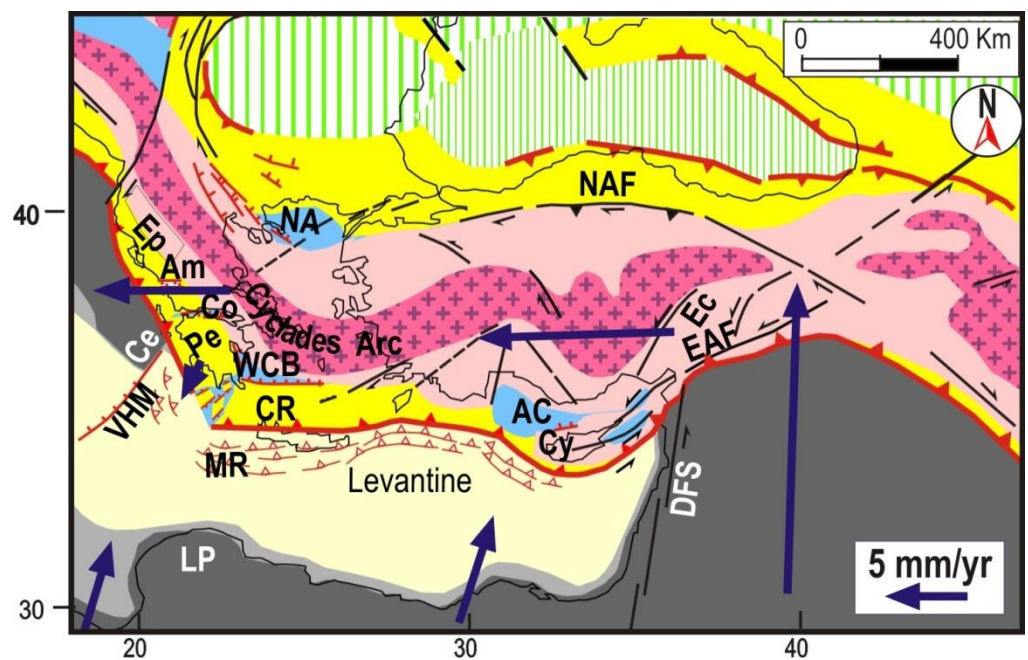


**Figure 7.** Late Miocene–early Pliocene paleogeographic setting, preceding the collision between the Aegean Tethyan belt and Adria. Ec—Eceemis fault; Ov—Ovacik fault. Symbols, colors and other abbreviations as in Figure 5.

Due to their different rheological behaviour, the belts constituting the Aegean arc underwent different deformation patterns in response to the E–W compression. The inner, mainly ductile, core (Cycladic arc) underwent a southward bowing without major fragmentations, whereas the outer brittle belt (Hellenic arc), being affected by intense belt-parallel tensional stresses in its most curved sector, broke in two branches: the Peloponnese and Crete–Rhodes (Figure 8). After separation, these belt sectors progressively released their previous horizontal flexure, by clockwise (Peloponnese) and counter-clockwise (Crete–Rhodes) rotations, respectively. The divergence between these migrating sectors and the inner Cyclades massif induced extension in the interposed zone. This tectonic mechanism may explain why crustal thinning, with a dominant S–N extensional trend, occurred in the western Cretan basin (WCB) from the late Miocene to the late Pliocene [26,99–102], and why such a regime only affected a limited, almost triangular zone (Figure 8).

The hypothesis about the mainly ductile behaviour of the inner Tethyan belt is primarily suggested by the present shape of this structure, characterized by the continuity and major bendings. Since that belt generated along the boundary of an oceanic domain, it seems unlikely that its present strongly deformed shape was the original one.





**Figure 8.** Middle Pliocene tectonic setting. AC—Antalia–Cilicia basin; Am—Ambracique trough; Ce—Cephalonia fault; Co—Corinth trough; CR—Crete–Rhodes (eastern Hellenic arc); Cy—Cyprus; EAF—East Anatolian fault system; Ep—Epirus; LP—Libyan promontory; MR—Mediterranean ridge; NA—North Aegean trough; Pe—Peloponnesus; WCB—western Cretan basin; VHM—Victor Hensen–Medina fault. Symbols colors and other abbreviations as in Figure 5 and Figure 7.

The hypothesis that the Tethyan belt transmitted the westward push of the Anatolian wedge to the Adriatic–Hellenides structures is suggested by the shortening that these zones underwent since the middle–late Miocene, involving a thrusting and uplift (e.g., [94,103–105]) and a clockwise rotation of about  $40^\circ$  of northern Hellenides between 15–13 and 8 Ma [57,106]. The analysis by [57] excludes the possibility that significant paleomagnetic rotations occurred before 15 Ma. In the Pliocene, the convergence between the Cycladic arc and Adria caused the southward escape of the Peloponnesus wedge and the formation of E–W troughs, such as the Corinth and Ambracique ones (Figure 5, [41,52,107,108]). The above compressional context is still active, as suggested by the occurrence of strong and frequent seismic activity, characterized by compressional and transpressional mechanisms ([67] and references therein).

The divergence between the southward bowing Aegean arc and Rodope can explain why in the North Aegean and northwestern Anatolian zones transtensional tectonics accelerated in the Pliocene, with the formation of dextral fault systems, purely transcurrent to the east and prevalingly extensional to the west (e.g., [26,29,109–112,113] and references therein).

The slip rate along the NAF fault system increased from about 3 mm/yr in the late Miocene to about 20 mm/yr in the last 2.5–3 Ma [91]. This acceleration could be due to the activation of the eastern Anatolian fault system (e.g., [88,89,91]), which allowed the complete decoupling of the Anatolian wedge from Arabia and/or to the fact that the progressive southward bending of the Aegean arc reduced the resistance against the westward displacement of Anatolia, favouring the activation of decoupling fault systems in eastern Anatolia.

During the Pliocene, the Cyclades massif was affected by an intense compressional deformation and uplift, which produced the first deposition of continental facies, after the Miocene marine sedimentation [26,95,114]. Large-scale overturned and upright folds with axes parallel to the extensional trend (roughly S–N) have been recognized in the Cyclades islands (e.g., [36,115]). The fact that in this zone several Ma of crustal stretching failed to produce any net crustal thinning was interpreted by [36] as an effect of contemporaneous

thickening caused by E-W shortening, through the exhumation of lower crustal material [116]. The simplest explanation of the above features in the Cyclades zone involves the E-W compression induced by the westward escape of Anatolia, as suggested by other authors (e.g., [59,117,118]).

After their separation, the eastern and western Hellenic arcs underwent different deformation patterns (Figure 9A,B). In Greece, the E-W compression between the Cycladic arc and the continental Adriatic domain caused a marked uplift and roughly southward escape of narrow crustal wedges (slices) in the Peloponnesus, at the expense of the Ionian and Levantine oceanic domains. The S-N extension induced by the divergence between mainland Greece and the escaping Peloponnesus wedges generated the Corinth and Ambracique troughs (e.g., [108,119]). Since the compressional context was exerted by two different converging structures, the Cycladic arc on one side and the shorter Adria southern boundary (Cephalonia fault) on the other side, the E-W compression was accompanied by a clockwise torsion. This mechanism may explain the angular divergence and dextral shear that formed some longitudinal troughs in the southern Peloponnesus (e.g., [119–121]). The dextral Cephalonia transpressional fault marks the overthrusting of the Peloponnesus wedge on the southernmost Adriatic continental domain (e.g., [26,122,123]).

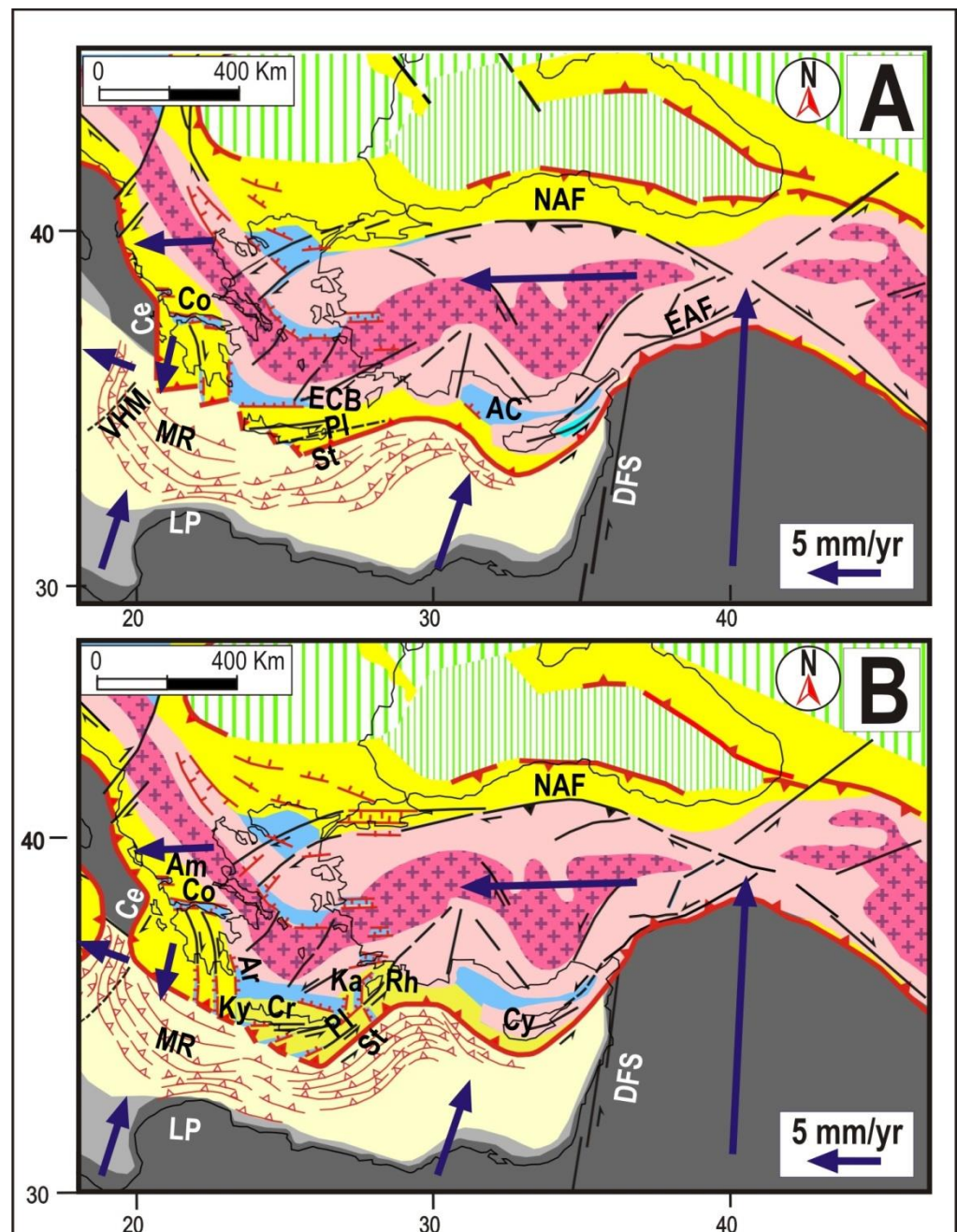
The bending of the Aegean arc was not the only effect of the enhanced E-W compression that followed the late Miocene–early Pliocene collision of the Aegean Tethyan belt with the continental Adriatic domain. This acceleration also produced further southward bending of the eastern Anatolian Tethyan belt (Figure 8; Figure 9). The timing of this deformation is suggested by the fact that the late Miocene compressional and extensional deformation patterns, respectively, occurred along the outer and inner sides of the Cyprus arc (e.g., [124–126]). This inference is based on the hypothesis that the detachment of the Cyprus wedge from Anatolia, with the formation of the Antalya and Cilicia basins, was determined by the belt-parallel extensional stress that occurred in the outermost sector of the Tethyan belt in response to bending (Figure 8). Belt-parallel stress in the outer belt also caused the thinning and subsidence of the narrow land strip connecting the Cyprus arc to easternmost Anatolia.

In the central Mediterranean region, the late Miocene collision of the Tethyan belt with the continental Adriatic domain caused a major reorganization of the tectonic context, as extensively described by [127,128].

In the late Pliocene, the deformation pattern of the Hellenic arc underwent a major change, due to the collision of Crete with the Libyan promontory, as suggested by geological and morphological evidence [129–132]. After this contact, the Crete–Rhodes sector, stressed by the convergence between southwestern Anatolia and Nubia, underwent SE-ward bending at the expense of the Levantine domain (Figure 9B). This mechanism was probably responsible as well for the land interruption between Crete and Rhodes [41,102,133]. The extension that developed in the wake of the SE-ward bending Crete–Rhodes sector formed the eastern Cretan basin and the Karpathos trough (Figure 9B). Since the resistance of the Libyan promontory only acted on the southern part of the Cretan wedge, the SW-NE compression was associated with a sinistral shear, which caused the slicing of that wedge, involving a faster westward motion of the northernmost slice (Crete island), with respect to the southern slices (Figure 9B). The sinistral relative motion between those slices was allowed by the Pliny and Strabo sinistral strike-slip major faults.

During this phase, the Crete slice (pushed roughly westward) collided with the Kythira slice of Peloponnesus, causing its clockwise rotation and the consequent formation of the Argolides trough (Figure 9). This context may explain the intense uplift recognized in the colliding structures (Crete and Kythira, Figure 2C) and may have favoured the clockwise rotation of southern Peloponnesus as well. The authors of [133] suggest that the onset of this rotation shows a remarkable coincidence with the change in stress regime between 0.8 and 0.3 Ma that affected the Aegean region and with the coeval uplift of the entire Aegean outer-arc (Figure 2C).

The focal mechanisms of subcrustal earthquakes [44,134–137] indicate a complex slab deformation regime dominated by SW-ward extensional and E-W compressional forces, as confirmed by the main features of the strong 2006 earthquake in the Kythira strait. The analysis of the microseismicity related to subcrustal earthquakes along the Hellenic arc (e.g., [134–136]) suggests that the downgoing slab is considerably contorted. Such an interpretation agrees with the results of [138], who observed an undulating topography of the African plate Moho beneath western Crete after an analysis of the migrated receiver functions.



**Figure 9.** (A) Late Pliocene paleogeographic configuration. AC—Antalya–Cilicia basin; Ce—Cephalonia fault; Co—Corinth trough; DFS—Dead Sea fault system; ECB—eastern Cretan basin; EAF—East Anatolian fault system; LP—Libyan promontory; MR—Mediterranean ridge; NAF—North Anatolian fault system; Pl and St—Pliny and Strabo faults; VHM—Victor Hensen-Medina fault system. (B) Late Pleistocene paleogeographic setting. Am—Ambracique trough; Ar—Argolides trough; Cr—Crete island; Ka—Karpathos trough; Ky—Kythira slice; Rh—Rhodes. Colors, other symbols and abbreviations as in Figure 5; Figure 8.

### 3. Alternative Geodynamic Interpretations

Since a reliable recognition of tectonic settings in the study area may be very useful for various purposes, also involving social objectives as the estimate of seismic hazard, we think that any presentation of a new geodynamic hypothesis should be integrated by a description of the possible weaknesses of the main previous interpretations. This may help the identification of a widely accepted opinion about the tectonic setting responsible for seismic activity, which could be used for a tentative recognition of the time–space distribution of major earthquakes [1,2].

#### 3.1. Slab-Pull Model

This hypothesis suggests that the southward migration of the Aegean–Hellenic arc was driven by the retreat of the Ionian–Levantine slab (e.g., [56,63,139,140]). However, the implications of this hypothesis cannot easily be reconciled with some major features of the observed deformation pattern, as discussed in the following:

In the middle–late Miocene, the present slab, identified by the distribution of sub-crustal earthquakes, was not sufficiently developed for generating a sufficient slab-pull force. Thus, the slab-pull model relies on the hypothesis that, in the Miocene, there was a previous well-developed subducted lithosphere. The presence of such a slab is mainly supported by tomographic investigations (e.g., [56,63,141–143]). However, one should explain why that presumed lithospheric body is not affected by seismic activity. Whatever hypothesis is advanced to justify the lack of earthquakes deeper than 160 km, it remains to understand why under the southern Tyrrhenian, where there is a slab generated by the subduction of the same oceanic domain, strong earthquakes occur down to depths greater than 450 km. Insights into the possible uncertainties of tomographic data may also be achieved by the comparison of the deep structural settings derived from different kinds of investigation in the Northern Tyrrhenian–Apennine zone ([144–149]). To understand the real evolution of the subduction process under the Aegean zone, it may be useful taking into account some major aspects of the previous (Eocene–Oligocene) Mediterranean evolution, as discussed by [127,128], in particular the fact that during the long phase of collision between the Africa/Adriatic promontory and Eurasia, the Tethys oceanic domain (located between continental domains) did not undergo any subduction. This kind of behaviour, recognized in other parts of the world (e.g., [150–152]), means that the consumption of oceanic lithosphere may not simply occur as an effect of plate convergence. This behaviour is a consequence of the fact that, in geological time intervals, a mature oceanic lithosphere is characterized by a horizontal compressional strength larger than one of a continental domain, as indicated by long-term rheological profiles (e.g., [153]). However, to explain why in the Neogene the subduction of oceanic domains (Alpine and Ionian Tethys) took place in various Mediterranean zones, it is necessary to recognize what were the conditions that allowed such a process to occur. Since all-consuming processes in that period occurred below extruding orogenic wedges (e.g., [127,128]), one can suppose that the starting of subductions was favoured by the load of orogenic extruded material. This overload may have perturbed the previous equilibrium in the collision zone, triggering the subduction of the denser oceanic lithosphere [154,155]. Numerical experiments (e.g., [156]) and laboratory modelling (e.g., [154,155,157–160]) have shown that in collisional contexts, the lateral escape of buoyant orogenic wedges is the most convenient shortening process and that crustal extension may develop in the wake of a migrating arc.

The above considerations suggest that the consumption of the Ionian and Levantine oceanic domains in front of the Aegean arc started in the early–middle Miocene, when the Tethyan belt (stressed by the Arabian indenter) began its SW-ward bending. A significant acceleration of this process occurred around the late Miocene–early Pliocene when the Aegean sector of that belt collided with the Adria continental domain. This interpretation is compatible with the onset time of the Aegean subduction suggested by other authors

(e.g., [58–62,112]) on the basis of geological and geophysical observations and of the Plio-Quaternary age of the volcanic arc (Figure 2C).

In our view, the hypothesis that the Aegean subduction started much earlier [141–143] should be accompanied by the identification of the extrusion process that may have triggered such a consumption. In this regard, it is worth noting that despite the strong compression which stressed the Levantine basin for tens of My, that domain did not undergo any subduction under the Nubian continental domain.

1. Another major problem of the slab-pull model is the fact that, since the late Miocene, the zone running from central Greece to Albania has been affected by strong E-W compressional deformation, such as crustal thickening and a strong uplift (e.g., [94,103–105]), which is considered to be responsible for the southward escape of the Peloponnesus wedge and for the formation of E-W troughs, such as the Corinth and Ambracique ones (Figure 5, [41,52,107,108]). This compressional stress regime is still going on, as testified by the strong and frequent seismic activity characterized by compressional and transpressional mechanisms ([67] and references therein). The above evidence can hardly be reconciled with an SW-ward pull, whereas it can simply be interpreted as a consequence of the E-W convergence between the Cycladic arc and the Adriatic continental domain.
2. The present shape of the inner metamorphic core of the Tethyan belt does not show any significant interruption or thinning in the Aegean and Anatolian sectors (Figure 5A). This structural continuity would imply that the retreat of the Hellenic slab did not only pull the Aegean sector of the belt, it would have also pulled the entire Anatolian body, a result that cannot easily be imputed to the retreat of a relatively narrow slab.
3. The marked oroclinal bendings of the Aegean arc (Figure 5A) can hardly be explained as an effect of a pull. This deformation would have required a much faster trench retreat in the central part of the arc with respect to the lateral sectors, implying a very peculiar shape of the sinking slab. Conversely, the bowing can easily be interpreted as an effect of the E-W compression induced by the convergence between eastern Anatolia and the Adriatic continental domain.
4. The numerical modelling of slab-pull processes [161–163] suggests that when a realistic parametrization of the overriding plate is adopted, the slab-pull force cannot cause the break of that plate and, thus, the development of the back arc basin.
5. The complex space–time distribution of Plio-Quaternary strain regimes in the Aegean zone is rather different from the one that usually develops in a back arc basin ([24,25,67] and references therein). The northern Aegean zone has been affected by a sinistral transtensional regime. The Central Aegean (Cyclades arc) has simultaneously undergone E-W compression and S-N extension. The southern Aegean was affected by SW-NE extension in the Pliocene (western Cretan basin) and by S-N extension in the Pleistocene (eastern Cretan basin). It seems rather difficult to explain such a complex deformation pattern with the implications of a simple slab-pull driving mechanism. In particular, a southward pull cannot explain the strong E-W compression that was clearly recorded in the Cyclades massifs (e.g., [36]). The above-mentioned authors took into account the possibility that such a deformation is related to the westward displacement of Anatolia, but they concluded that this hypothesis could not explain the fact that E-W shortening in the Cyclades started developing in the early Miocene when the NAF was not yet active. However, this last consideration does not consider that, in the early Miocene, the Tethyan belt was already undergoing belt-parallel compression, driven by the Arabian indenter ([52] and references therein). This regime accelerated in the early Miocene, after the complete consumption of the Magura oceanic zone in front of the migrating Carpathian arc, which considerably increased the resistance against any further NW-ward displacement of the Tethyan belt. This may explain why the outward bending of the Tethyan belt, at the expense of the Levantine–Ionian domain, began at that time, causing a coeval NE-

SW extension and belt-parallel compression in the Tethyan belt, which is considered the genetic mechanism of the crustal deformation and petrological evolution of the Cyclades massifs (e.g., [36]).

The reason as to why, despite the above difficulties, the slab-pull hypothesis has not yet been abandoned may be due to the fact that the present kinematic field inferred by geodetic data (Figure 3) is mostly taken as a representative of the long-term behaviour of the Anatolian–Aegean zone. This speculative assumption led some authors to invoke the action of an additional driving force able to justify the faster migration of the Aegean arc with respect to Anatolia. However, this view does not take into account the possibility that the ongoing kinematic field is a transient pattern, related to the post seismic relaxation that was triggered by the activation of the entire NAF fault since 1939.

The hypothesis that the present geodetic velocity field is rather different from the long-term one is suggested by some major features of the Aegean system. The concave shape of the Hellenic slab (Figure 4) and the fact that the width of the Mediterranean ridge accretionary complex is almost uniform around the entire Hellenic trench from the Cephalonia fault to Rhodes (Figure 1) testifies that the subduction of the Ionian–Levantine domain occurred under the entire migrating arc, involving subduction trends varying from NE to NW-ward, with more or less comparable rates. Thus, it seems very difficult to understand why at present the supposed slab rollback may induce a dominant SW-ward motion of the Aegean zone (Figure 3).

### 3.2. Gravitational Spreading

This hypothesis suggests that the formation of the Aegean arc was driven by gravitational spreading, due to the strong difference of crustal thickness between the Aegean domain and the Ionian Levantine oceanic zone (e.g., [109,116,164]). The implications of this driving force, quantified by laboratory experiments (e.g., [116,165]), cannot easily be reconciled with some major features of the observed deformation pattern:

1. To explain the bowing of the Hellenic arc (Figure 1), one should assume that the spreading rate in the central part of the Arc was higher than in the lateral sectors, in such a regular distribution to produce an Arc shape.
2. The effects of E-W compression, as an uplift, crustal thickening and very strong seismic activity, recognized in the Hellenides belt (e.g., [94,104,166]) are not compatible with the presumed spreading of Aegean masses towards the Ionian–Levantine domains.
3. The E-W shortening recognized in the Cycladic arc [36] cannot easily be explained as an effect of the gravitational spreading of the Aegean masses towards the Ionian–Levantine domains.
4. The crust of the Nubian plate is thicker than the one of the Levantine and Ionian domains, so one should explain why gravitational spreading did not occur at that boundary zone.

## 4. Nubia–Eurasia Relative Motion

The convergence between the two main confining plates, Nubia and Eurasia, constitutes one of the two main driving forces of the Mediterranean evolution. Thus, a reliable recognition of this boundary condition may crucially be important for understanding the geodynamic context that determined the observed deformation pattern. Global kinematic models (GKM, e.g., [167–169]) suggest a SE-NW Nubia–Eurasia convergence trend. However, the implications of this plate’s kinematics cannot easily be reconciled with major features of the Mediterranean deformation pattern, as argued in previous works [66,67,170,171]. Here, we report some considerations about the main difficulties of the above hypothesis and the evidence that supports the SSW-NNE Nubia–Eurasia convergence trend adopted in this work.

1. The southern part of the Adriatic plate moves roughly NE-ward with respect to Eurasia, as suggested by many authors on the basis of geological, geophysical and geodetic information [67,172–177]. Since this motion trend is almost perpendicular to the NW-ward Nubia–Eurasia convergence trend provided by GKM, one could expect to observe a clear Nubia–Adria decoupling zone characterized by significant seismotectonic activity. In particular, the divergence between the Adria plate and the Ionian domain (belonging to the Nubian plate) should produce clear extensional deformations in a zone more or less corresponding to the Apulian escarpment. However, no significant active deformation has been recognized in that structure and its surroundings (e.g., [178]). Numerous attempts at identifying other possible decoupling zones have been undertaken, but the variety of solutions so far proposed, located all over the Adria plate, testifies the scarce significance of the available tectonic and seismic evidence in support of a decoupling fault (e.g., [170,179] and references therein). Even though one could identify the invoked decoupling, it would remain the problem of identifying the driving force responsible for the independent motion of the Adriatic plate. The Nubia–Eurasia convergence trend adopted in this work (SSW–NNE) strongly mitigated the problem mentioned above.
2. The morphology of the outer Hellenic arc is characterized by a SE–NW trench and SW–NE sinistral strike-slip fault systems (Pliny and Strabo, Figure 5). These features, with particular regard to the deformation pattern of the Crete–Rhodes structure discussed earlier (Figure 9), can hardly be reconciled with the presumed NW motion of Nubia. This problem could only be mitigated by assuming that the SW-ward motion of the Aegean arc was much faster than the Nubia–Eurasia convergence. However, the long-term kinematics of this zone derived by geological data and by the present shape of the subducted lithosphere do not support such a possibility. Similar considerations can be determined for the Cyprus–Nubia boundary, constituted by a SE–NW trench and an NE–SW strike-slip fault system (Figure 5A).
3. Other major tectonic features in the central and western Mediterranean area can hardly be reconciled with a NW motion of Nubia, as discussed by [66,67,170,171]. In particular, the SW-ward extrusion of the Calabrian wedge at the expense of the Ionian domain, clearly indicated by the strong deformation, uplift and fracturing of that structure, can hardly be reconciled with a NW ward motion of Nubia (e.g., [127,128]).

Stimulated by the difficulties mentioned above, we thought it opportune to verify the reliability of the conditions assumed by GKM for the determination of the Nubia–Eurasia relative motion. The results of this analysis suggest that the plate mosaic used by GKM (Nubia and Eurasia) is too simple, since the distribution of seismicity and other major evidence indicates that a four-plate model, also involving the Iberia and Morocco microplates, is more realistic. The authors of [66] showed that adopting such a mosaic Nubia–Eurasia motion trend here proposed (SSW–NNE) is compatible, within errors, with all the Atlantic constraints considered by GKM, with the constraints on the relative motions of all plates involved, and can plausibly account for the main Mediterranean geological constraints on the recent/present Nubia–Eurasia motion trend.

A significant part of the dataset supporting the SE–NW Nubia–Eurasia relative motion trend derives from geodetic observations (e.g., [180–182]). However, in some cases, the use composed of this kind of information may be questioned. For instance, the data located in the Hyblean zone and North Africa (Tunisia, Algeria) are mostly attributed to the Nubian plate, whereas the tectonic evolution of the central and western Mediterranean regions suggests that such zones are not moving in close connection with Nubia [127,128]. As discussed earlier, the geodetic data in the Aegean region were taken as a representative of the long-term kinematics of that zone, without considering the perturbation of the velocity field triggered by the strong seismic sequence along the NAF since 1939. It is worth noting that some authors [169] point out a significant discrepancy between the Nubia–

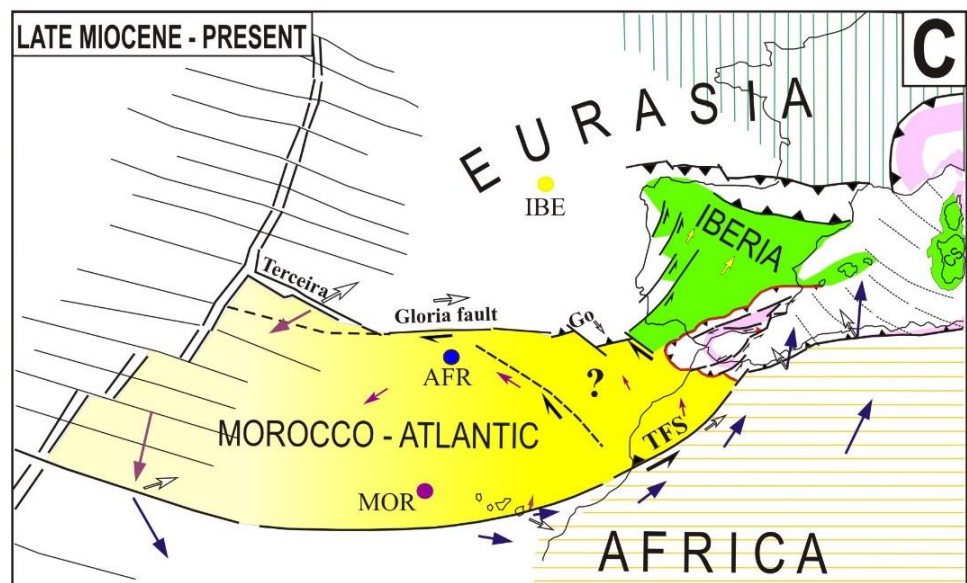
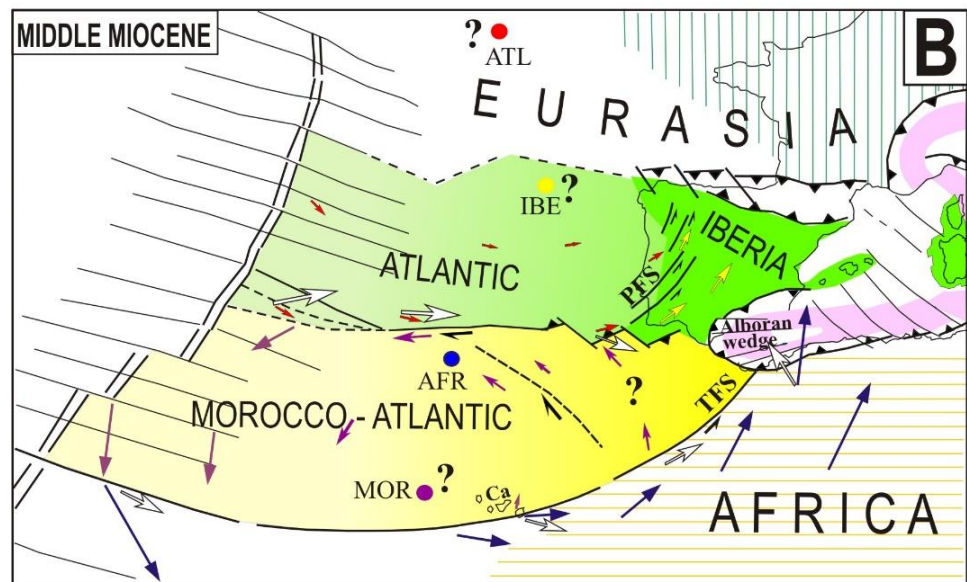
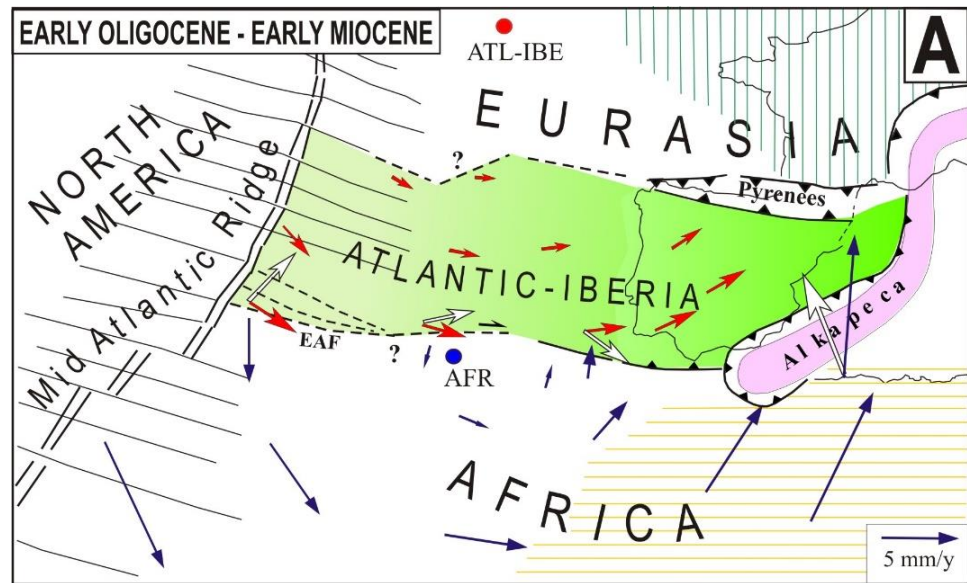
Eurasia Euleran poles inferred from geodetic data and those derived from Quaternary geological evidence.

Insights into the recent/present Nubia–Eurasia convergence trend may also derive from the Atlantic kinematic indicators related to the last tens of MY. The interpretation of these data, carried out by combining the Nubia–North America and Eurasia–North America plate motions [169], suggests that, before 25 My, the Nubia–Eurasia convergence was oriented roughly NNE-ward and that between 25 and 13 My, the convergence slowed down to about 50% and underwent an anticlockwise rotation, passing from NNE-ward to about NW-ward. However, such a considerable change (about 70°) of the Nubia–Eurasia motion trend cannot easily be reconciled with major tectonic features in the Mediterranean area [66,67,170,171]. Furthermore, the supposed change would have required a drastic reorganization of tectonic activity in all active and passive Nubian and Eurasian boundaries.

We rather think that the reconstruction of [169] must be revised, since it does not take into account a major tectonic event (Figure 10) that may have changed the implications of the North Atlantic kinematic indicators related to post 13 My, as schematically reconstructed in Figure 10 and explained in the following. Prior to 25 My, when the African plate moved roughly NNE-ward [169], the convergence with the Iberian domain was accommodated by the consumption of the thinned continental African and Iberian margins, by the deformation of the Alpine belt (Alkapeca in Figure 10A) and by the westward escape of the Alboran wedge. The progressive increase in resistance at this collisional boundary, as the thinned domains were reducing, caused the shortening and slowdown of the Nubia sector mostly involved in the collision. This context reached a critical point around 13 My, when the activation of the trans-Moroccan fault system (e.g., [183] and references therein) allowed the decoupling of the Morocco and adjacent Atlantic domain from the African plate (Figure 10B). After this event, the Morocco–Atlantic block underwent a significant slowdown (being no longer connected with Nubia), whereas the adjacent Nubian domain (no longer restrained by the Moroccan sector) recovered its previous motion rate. Around the middle Miocene, the same collisional context led the Iberian block to decouple from the Atlantic domain. This decoupling was allowed by a system of NE-SW to NNE-SSW sinistral transpressional faults in the northern Portugal zone (PFS in Figure 10B), probably reactivating Paleozoic and Mesozoic weakness zones (e.g., [184–187]). After decoupling, the Iberian block started moving more northward than before (Figure 10C), as indicated by the renewal of sinistral transpressional tectonics recognized in the Pyrenean belt and surrounding zones (e.g., [188]).

The above considerations imply that the Atlantic kinematic indicators located west of the Morocco–Atlantic and Atlantic–Iberia microplates and related to times younger than 13 My are not related to the Nubia–North America and Eurasia–North America relative motions, respectively. If such data were used anyway to derive the Nubia–Eurasia convergence, as conducted by [169], the result would be an apparent anticlockwise rotation of Nubia with respect to Eurasia. If, instead, these data were more realistically used to constrain the relative motions of the four-plate mosaic [66], the result would be compatible within errors with the SSW–NNE Nubia–Eurasia motion trend here proposed.

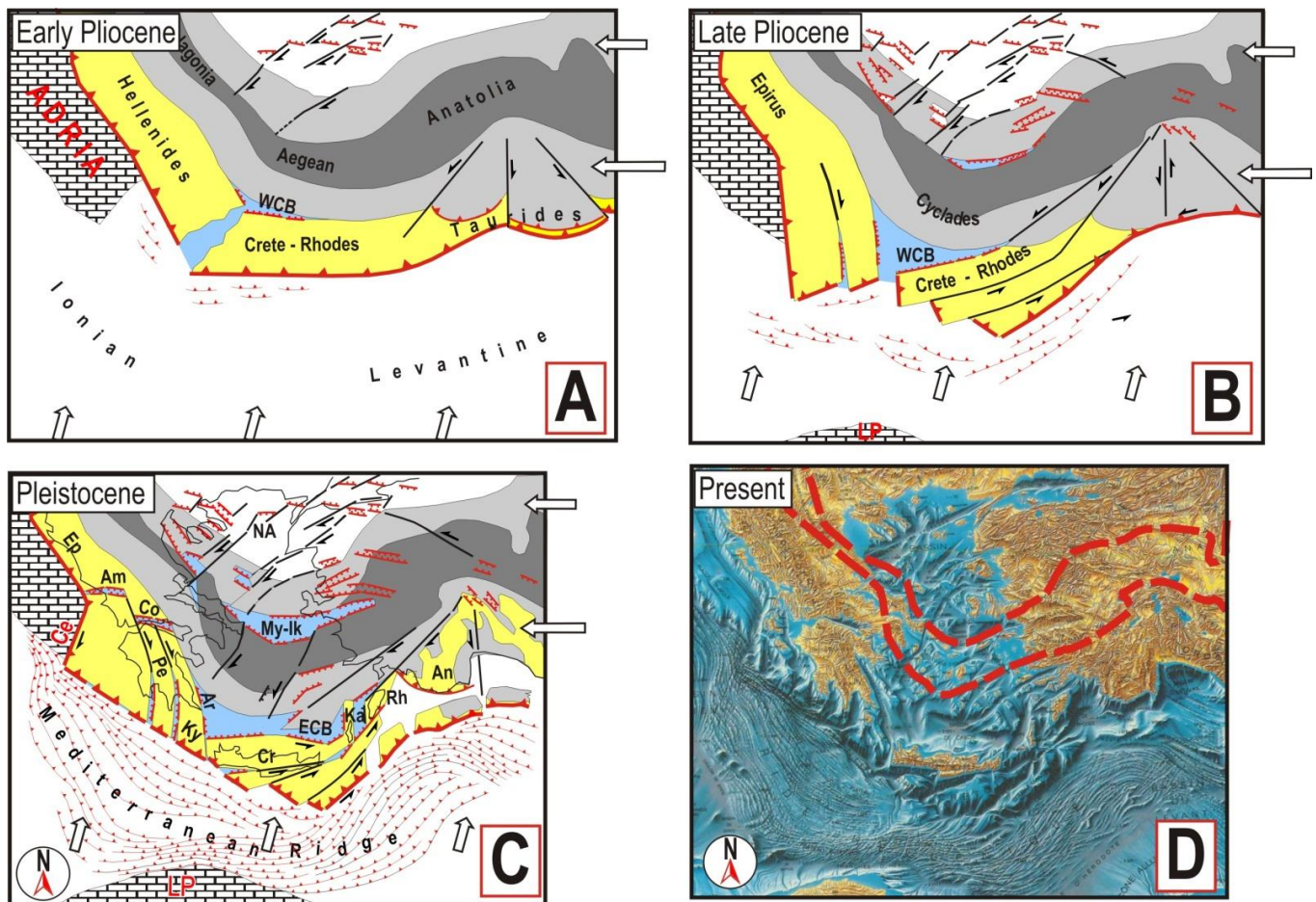




**Figure 10.** Geodynamic context that led to the detachment of the Morocco–Atlantic block with respect to Africa (modified from [171]). **(A)** About 25 My, the Atlantic–Iberia plate moved independently from Africa and Eurasia. Motion vectors (coloured arrows) were computed using the Euler poles indicated by relevant coloured points (Atlantic–Iberian (ATL-IBE): 50.0° N, 16.2° W,  $w = 0.10^\circ/\text{My}$ ; Africa (AFR): 36.2° N, 18.0° W,  $w = 0.24^\circ/\text{My}$ ). Empty arrows close to the southern plate border indicate relative motions predicted between the Atlantic–Iberia plate and Africa. EAF—eastern Azores fracture zone. **(B)** About 20 My, the collision between Africa and Iberia, with the interposed Alboran wedge, forced the Iberia microplate to decouple from the Atlantic domain and the Morocco–Atlantic microplate to decouple from Africa. Euler poles with respect to Eurasia: Iberia (IBE): 43.5 N, 14.2 W,  $w = 0.12^\circ/\text{My}$ ; Morocco–Atlantic (MOR): 28.5 N, 21.0 W,  $w = 0.13^\circ/\text{My}$ ; Africa: 36.2 N, 18.0 W,  $w = 0.16^\circ/\text{My}$ . The Atlantic microplate is tentatively assumed to be still moving independently from Eurasia (around the Euler pole ATL located as the ATL-IBE pole in **(A)**), with a velocity lower than in the first phase ( $w = 0.05^\circ/\text{My}$ ). The easternmost corner of the Morocco–Atlantic microplate might not move in complete connection with its main body (see discussion in [66]). The decoupling zone of this possible fragment could correspond to the NW-SE fault zone located offshore western Morocco, at the site of a recent strong earthquake (e.g., [189,190]). Empty arrows indicate relative motion between Morocco–Atlantic and Atlantic (along the Azores–Gibraltar border), between Iberia and Africa (in the Alboran zone) and between Morocco–Atlantic and Africa (along the trans-Moroccan–Canary Islands border). Ca—Canary Islands; PFS—Portugal fault system; TFS—trans-Moroccan–Canary Islands fault system. **(C)** The former Atlantic block is part of Europe. Other Euler poles as in **(B)**. Empty arrows indicate relative motions between the Morocco–Atlantic plate and Eurasia (along the Azores–Gibraltar border), between Morocco–Atlantic and Africa (at the trans-Moroccan–Canary Islands fault system) and between Iberia and Africa (at the western Algerian basin). Go—Gorringe thrust; CS—Corsica–Sardinia block.

## 5. Conclusions

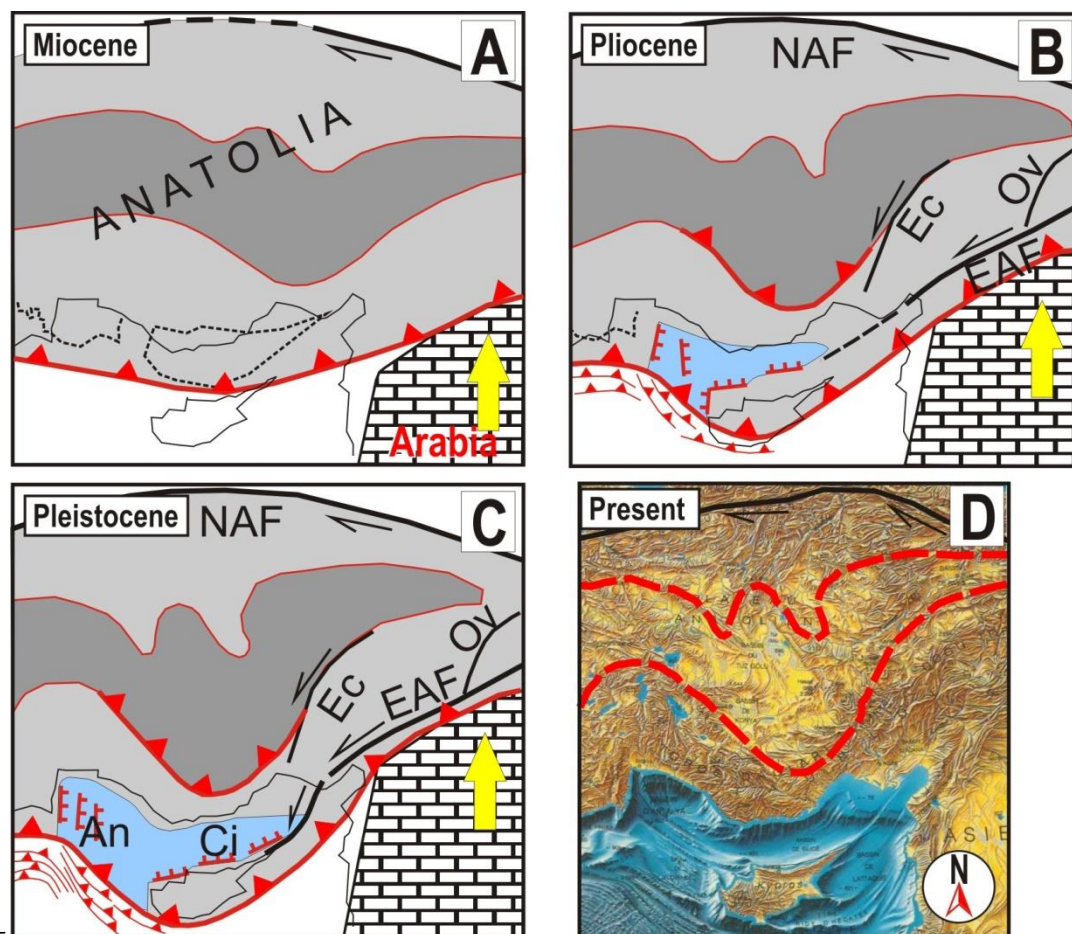
The tectonic evolution of the eastern Mediterranean region has been considerably influenced by the deformations of the Tethyan belt, driven by the indentation of the Arabian promontory. After decoupling from the Iranian sector by a SE-NW dextral fault system, the Anatolian–Aegean–Pelagonian sector was pushed roughly NW-ward, at the expense of the surrounding thinned domains. During this Miocene phase, the belt started its SW-ward bending at the expense of the Levantine and Ionian domains, to reach an almost E-W orientation. When the thinned domains interposed between the Arabian indenter and continental Eurasia were almost completely consumed or/and shortened and uplifted, the compression in eastern Anatolia considerably increased, starting its westward escape. This new context and the collision of the Aegean Tethyan sector with the Adriatic continental domain accelerated the southward bending of the Aegean arc, around the late Miocene-early Pliocene. Oroclinal bending was accommodated, with different deformation patterns, by the various belts constituting the Tethyan system (Figure 11). The central, more ductile metamorphic core (Cyclades) deformed without undergoing significant fragmentations, whereas the most curved (and stressed) sector of the outer brittle belt was affected by a major break, which led to the separation between the eastern (Crete–Rhodes) and western (Peloponnesus) arcs (Figure 11B). The divergence between these arcs and the Cycladic core led to the formation of the western Cretan basin. Since then, the western arc, stressed by the roughly E-W convergence between the Cycladic arc and the Adria domain, underwent crustal thickening, an uplift and southward escape, accompanied by a clockwise torsion. The divergence between the escaping Peloponnesus wedge and mainland Greece formed E-W troughs, such as the Corinth and Ambracique ones, while the torsion formed a series of curved transtensional faults in southern Peloponnesus (Figure 11C).



**Figure 11.** Tentative reconstruction of Aegean arc's deformation pattern (A–C) and present morphological setting ((D), modified from [191]). The red dashed lines in (D) tentatively delineate the shape of the inner Tethyan belt. Am—Ambracique trough; An—Antalya; Ar—Argolides trough; Co—Corinth trough; Cr—Crete; ECB—eastern Cretan basin; Ep—Epirus; Ka—Karpathos; Ky—Kythira slice; LP—Libyan promontory; My-Ik—Mykonos–Ikaria basin; NA—North Aegean; Pe—Peloponnese; Rh—Rhodes; WCB—western Cretan basin. Symbols as in Figure 5.

The deformation of the Crete–Rhodes sector changed significantly around the early Pleistocene when its southwestern part collided with the Libyan promontory. Since then, the convergence between southwestern Anatolia and Nubia forced this structure to bend SE ward, at the expense of the Levantine domain. This driving mechanism also involved a sinistral shear that caused the slicing of the Cretan wedge, allowed by the Pliny and Strabo strike-slip fault systems. The divergence between the migrating Rhodes and Crete fragments with respect to the Cyclades massifs formed the eastern Cretan basin and the Karpathos trough, accompanied by the land interruption between Rhodes and Crete (Figure 11C).

The strong increase in E–W compression that affected the Tethyan belt since the late Miocene–early Pliocene also enhanced the bending of the inner core in eastern Anatolia, causing the detachment of the Cyprus arc from the main Anatolian body (Figure 12). The consequent divergence between this wedge and Anatolia formed the Antalya and Cilicia basins.



**Figure 12.** Tentative reconstruction of the tectonic mechanism that generated the Cyprus arc and the Antalya (An) and Cilicia (Ci) basins (A–C) and present morphological setting ((D), modified from [191]). EAF—East Anatolian fault; Ec—Ecemis fault; NAF—North Anatolian fault; Ov—Ovacek fault. Symbols as in Figure 5; Figure 11.

The evolutionary reconstruction here proposed **relies** on some basic concepts:

- (a) The deformations observed in the Mediterranean region were driven by the convergence between the surrounding plates (Nubia, Arabia and Eurasia), without any additional driving force, as provided by Plate Tectonics.
- (b) A SSW–NNE Nubia–Eurasia convergence was adopted, as suggested by numerous pieces of evidence in the Mediterranean region [66,170,171].
- (c) The very complex distribution of the shortening processes which accommodated such boundary conditions in each evolutionary phase was controlled by the minimum-action principle. At the end of a shortening process in a given zone (due, for instance, to the suture of a consuming boundary), the tectonic framework changed in order to activate the most convenient shortening pattern. This concept is very useful to understand the numerous tectonic reorganisations that have occurred in the study area since the Oligocene [52,66,67,128,129,172]. To complete the above view, it must be taken into account that the subduction of an oceanic lithosphere cannot simply be induced by plate collision. The triggering of that process requires the effects of the lateral escape of orogenic wedges. In fact, the consumption of the remnant oceanic domains in the Mediterranean region took place in front of migrating arcs.
- (d) Old belts, with an upper crust modified by the intrusion of lower crustal material, may deform in a dominant ductile way. This behaviour is suggested by the deformation pattern of the inner part of the Anatolian–Aegean–Pelagonian Tethyan belt. A strain field which can produce the crustal modification mentioned above is the

contemporaneous occurrence of extension and perpendicular shortening, as was clearly recognized in the Cycladic arc [36].

- (e) The mainly brittle behaviour of the outer parts of the Tethyan belt was clearly evidenced by the major break of the Hellenic arc, which led to the separation between the western and eastern branches of this arc, and by the following strong fracturation of the resulting fragments (Crete–Rhodes and Peloponnesus). Similar considerations were suggested by the deformation pattern of the Cyprus arc.

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