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Postcollisional contractional and extensional deformation in the Aegean region

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ABSTRACT

In the Aegean area, distinct fault patterns with their associated stress regimes are evidenced along a curved convergent plate boundary. In this article, we analyze and review late collisional and extensional structures in five structural regions along the External Hellenides orogenic belt in order to define, primarily, the evolution of onshore basins and, secondarily, the evolution of off-shore basins and the role of the inherited structures within the present geotectonic framework. We also evaluate how these structures act on seismicity as well as present-day motion and magmatism in the Aegean area.

Northwestern Greece, which corresponds to our Region I, represents an area of active continental collision in which a previously overthickened crust collapsed mainly parallel to the structural grain of the orogen. At present, the most active structures in this region are the northwest-trending thrusts and the northeast-trending normal faults. Strong coupling and the transmission of horizontal forces from the collision front appear to explain the deformation within this region.

Central Greece (Region II) displays a mixed type of contractional-extensional deformation. Mesozoic inherited transverse structures are reactivated as WNW-trending faults and appear to accommodate most of the active north-south-trending present-day extension. Deformation in this area appears to be controlled both by rollback of the subducting slab and by the lateral extrusion of the Anatolia plate.

The areas that spread along the more curved parts of the Hellenic arc (Regions III and V) well emphasize the control exerted by pre-existing northeast-trending structures as well as their long-lived activity during the evolution of the arc. Present deformation within these regions possibly reflects the oblique convergence process, which is occurring in both areas in different degrees.

Finally, analysis of deformation patterns in the central part of the Hellenic arc and the Aegean Sea (Region IV) suggests that almost all pre-existing structures have remained active until the present and accommodate extensional deformation and rapid motion through a nonorthogonal fault system that crosscuts a thin and thermally weakened continental crust. Deformation north of our Region IV is accommodated

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primarily by the North Aegean trough. Basin-bounding faults within the North Aegean trough are dominantly characterized by apparently normal components as well as strike-slip components.

Plate boundary conditions, as well as pre-existing structures with a favorable orientation relative to the current crustal motion, control and variably affect the present-day deformation along the Hellenic arc and trench system.

Keywords: Hellenic arc, pre-existing structures, Aegean area, reactivation, active deformation, plate margins, overriding plate deformation, fault pattern

INTRODUCTION

Fault patterns and subsequent strain variability have recently been reported either from areas located at plate margins (Rebai et al., 1992; Andeweg et al., 1999) or from areas with an inherited crustal anisotropy (Sassi and Faure, 1996; Tikoff and Wojtal, 1999). In both structural settings, the pre-existing structures appear to play an important role in active tectonic deformation processes within the overriding plate, occurring consecutively over a long period of time on plate convergent margins (Braun and Beaumont, 1995; Sutherland et al., 2000; Claypool et al., 2002).

The back-arc basins of many continent-based island arcs are often affected by indentation processes occurring in adjacent colliding plates, as is the case in the Japan Sea (Tapponnier et al., 1986; Fournier et al., 1994) and the Pannonian basin (Ratschbacher et al., 1991). However, the degree of lateral extrusion related to large transverse structures (*sensu* Duebendorfer and Black, 1992) and their kinematic role in the opening of the back-arc basins remains a controversial issue (Coward et al., 1988). A similar setting, albeit on a smaller scale, is also suggested for the eastern Mediterranean region (Fig. 1, inset).

In this article, we review the main syncollisional and extensional structures in order to draw conclusions about the neotectonic evolution along the active Hellenic arc (Fig. 1). Our data set includes fault pattern and stress analysis results (Table 1; Doutsos and Kokkalas, 2001) from almost 1000 mesoscopic and large-scale faults within five structural regions (Table 1, Fig. 1). Chronological control of deformation was based on stratigraphic data of Tertiary-Quaternary basins and granitic intrusions in the Cycladic area (Boronkay and Doutsos, 1994; Doutsos and Kokkalas, 2001, and references therein).

GENERAL TECTONIC FRAMEWORK

The tectonic framework of the eastern Mediterranean is dominated by the collision of the Arabia and Africa plates with Eurasia (Fig. 1, inset; Jackson and McKenzie, 1984; Jackson et al., 1992). The Arabia plate is moving in a north-northwest direction at a rate of ~18–25 mm/yr, while the Africa plate is moving in a northerly direction at a rate of 6 ± 2 mm/yr, both relative to stable Eurasia (De Mets et al., 1990). The differential motion between Africa and Arabia (~10–15 mm/yr) is responsible for the westward extrusion and counterclockwise rotation of the Anatolia plate, which is driven westward between the North and East Anatolia faults (Fig. 1, NAF and EAF). The North Anatolia fault displays a dextral slip with rates of 24 ± 2 mm/yr while the East Anatolia fault shows a sinistral slip of 9 ± 2 mm/yr (Kahle et al., 1998; McClusky et al., 2000). The leading edge of the Africa plate is being subducted along the Hellenic arc at a higher rate than the relative northward motion of the Africa plate itself. Therefore, a further motion of the overriding Aegean plate toward the trench is required. Recent GPS data indicate that the central and southern Aegean area is characterized by coherent motion toward the southwest at 30 ± 2 mm/yr relative to Eurasia (Fig. 1; Clarke et al., 1998; McClusky et al., 2000).

The orogenic belt of the External Hellenides, which can be followed along the Hellenic arc (Bonneau, 1984; Garfunkel, 2004), comprises a sequence of Mesozoic and Tertiary sediments deposited on a series of platforms (Pre-Apulian and Gavrovo isopic zones) and basins (Ionian and Pindos isopic zones) that formed the rifted eastern margin of the Apulian microcontinent, bordering the Pindos Ocean (Smith, 1977; Robertson et al., 1991). The External Hellenides were developed during Tertiary times, following the closure of the Pindos Ocean and the consequent continent-continent collision between the Apulian and Pelagonian microcontinents to the east (Mountrakis, 1986; Doutsos et al., 1993).

The late orogenic evolution of the Aegean is associated with rotations on various scales. Extensive paleomagnetic research has been carried out suggesting clockwise rotations of 45°–50° in the western mainland of Greece, from Albania to Evia island and the northern Peloponnesus, in two phases of ~25° each, one during the middle to late Miocene and a younger one during Plio-Pleistocene times (Laj et al., 1982; Duermeijer et al., 2000; Kissel et al., 2003). Walcott and White (1998) also confirmed rotations of that order based on spatial variation of stretching lineations from the crystalline basement of the Cyclades area. Rotations on a local scale on the order of 20°–30°, both clockwise and counterclockwise, affected the same area, probably during or after the cooling of granites, around 10 Ma (Morris and Anderson, 1996; Avigad et al., 1998; Koukouvelas and Kokkalas,

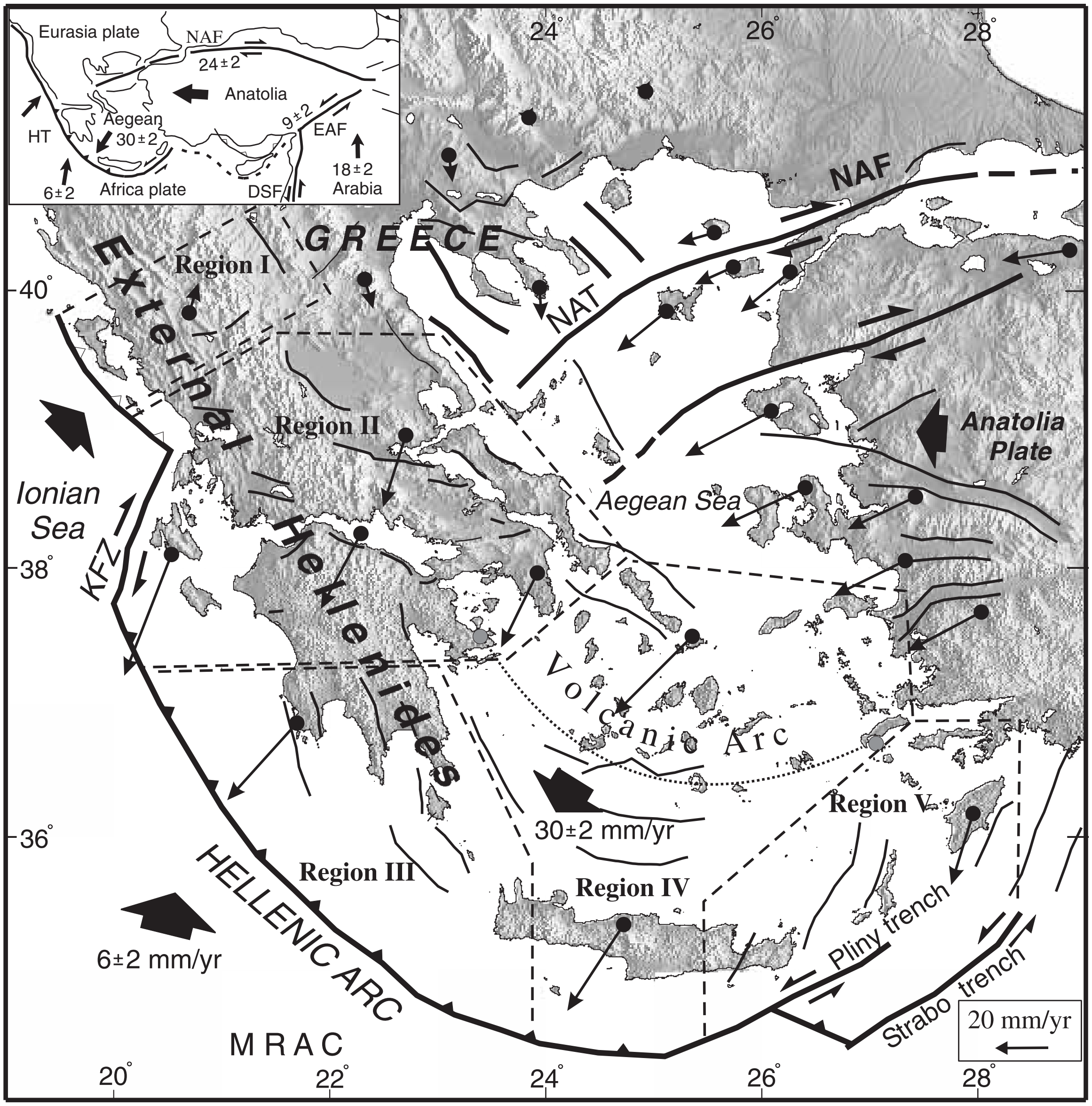

Figure 1. Simplified map showing the main structural features along the Hellenic arc and trench system, as well as the main active structures in the Aegean area. The mean GPS horizontal velocities in the Aegean plate, with respect to a Eurasia-fixed reference frame, are shown (after Kahle et al., 1998; McClusky et al., 2000). The lengths of vectors are proportional to the amount of movement. The thick black arrows indicate the mean motion vectors of the plates. The polygonal areas on the map (dashed lines) define the approximate borders of the five different structural regions discussed in the text. The borders between structural regions are not straightforward, and wide transitional zones probably exist between them. The inset shows a schematic map with the geodynamic framework in the eastern Mediterranean area (modified from McClusky et al., 2000). DSF—Dead Sea fault; EAF—East Anatolia fault; HT—Hellenic trench; KFZ—Kefallonia fault zone; MRAC—Mediterranean Ridge accretionary complex; NAF—North Anatolia fault; NAT—North Aegean trough.

TABLE 1. REVIEW OF STRESS ANALYSIS RESULTS IN THE STRUCTURAL REGIONS STUDIED

Stress net	Stratigraphic age	Location	*n*	Stress type
Region I: Northwest Greece				
I_A	Miocene	Mesohellenic trough	21	Strike-slip
I_B	Pliocene–Quaternary	Ptolemaida basin	46	Pure extension
I_C	Pliocene–Quaternary	Ptolemaida basin	17	Radial extension
I_D	Miocene	Mesohellenic trough	53	Transtension
Region II: Central Greece				
Stress net	Stratigraphic age	Location	*n*	Stress type
II_A	Pliocene–Quaternary	Thessalian basins	28	Pure extension
II_B	Pliocene–Quaternary	Atalanti graben	10	Pure extension
II_C	Pliocene	Paleros basin	23	Compression
II_D	Pliocene–Quaternary	West Corinth graben	23	Pure extension
II_E	Pliocene–Quaternary	Kymi basin, Evia	56	Pure extension-Transtension
II_F	Pliocene–Quaternary	Abelon graben	17	Radial extension
II_G	Pliocene–Quaternary	Pyrgos graben	27	Pure extension
II_H	Pliocene–Quaternary	Nedas graben	15	Pure extension-Transtension
II_I	Pliocene–Quaternary	Rio graben	30	Pure extension
II_J	Pliocene–Quaternary	Corinth area	38	Pure extension
Region III: Southwest Hellenic arc				
Stress net	Stratigraphic age	Location	*n*	Stress type
III_A	Pliocene–Quaternary	Kalamata graben	21	Pure extension
III_B	Miocene–Quaternary	Kythira island	12	Pure extension
III_C	Pliocene–Quaternary	Kythira island	20	Pure extension
III_D	Pliocene–Quaternary	Antikythira island	18	Pure extension
III_E	Miocene–Quaternary	Gramvusa peninsula, Crete	15	Pure extension to transtension
Region IV: Central part of the Hellenic arc				
Stress net	Stratigraphic age	Location	*n*	Stress type
IV_A	Miocene	Samos	24	Strike-slip to transpression
IV_B	Miocene	Naxos	15	Strike-slip
IV_C	Miocene	Milos	9	Strike-slip to transpression
IV_D	Miocene–Quaternary	Topolia basin, Crete	9	Pure extension
IV_E	Miocene–Quaternary	Episkopi rift zone, Crete	15	Transtension
IV_F	Miocene	Anafi	15	Radial extension
IV_G	Pliocene–Quaternary	Paleochora, Crete	8	Strike-slip
IV_H	Miocene	Koufonisia	12	Strike-slip
IV_I	Pliocene–Quaternary	Koufonisia	13	Transtension
IV_J	Miocene	Kymi basin, Evia	10	Strike-slip to transpression
IV_K	Upper Miocene	Kymi basin, Evia	35	Transtension
IV_L	Miocene	Paros	33	Strike-slip to transpression
IV_M	Pliocene–Quaternary	Paros	8	Transtension
Region V: Southeast Hellenic arc				
Stress net	Stratigraphic age	Location	*n*	Stress type
V_A	Miocene	Kos	8	Strike-slip
V_B	Pliocene–Quaternary	Kos	11	Transtension
V_C	Pliocene–Quaternary	Rhodes island	32	Pure extension
V_D	Pliocene–Quaternary	South Karpathos	17	Transtension
V_E	Miocene	Kasos island	10	Pure extension
V_F	Miocene	Ierapetra basin, Crete	18	Transtension
V_G	Miocene	Ierapetra basin, Crete	32	Pure extension
V_H	Pliocene–Quaternary	Sitia graben (south), Crete	65	Transtension
V_J	Miocene	Fothia basin, Crete	35	Transtension
V_K	Pliocene–Quaternary	Ierapetra basin, Crete	12	Strike-slip
V_L	Miocene	Grantes Bay, Crete	15	Strike-slip to transpression

Note: n—number of fault-slip data used in stress analysis.

2003). This suggests the fragmentation of the Cycladic massif to several microplates and fault blocks and their incorporation into the regional rotational regime. The southwestern Hellenic arc from Kythira to central Crete did not rotate significantly on a regional scale after 10 Ma (Duermeijer et al., 2000). For the region extending between eastern Crete and Rhodes island, counterclockwise rotations of 20°–30° have been determined in post-Messinian sediments, while rotations older than 10 Ma have not been recorded (Duermeijer et al., 2000).

All these rotational data may indicate that a pre-existing orocline was formed before the synextensional phase of deformation in the Hellenides, which is considered to have been initiated in the middle Miocene (Drooger and Meulenkamp, 1973; Le Pichon and Angelier, 1979). This orocline contributed to the arcuate geometry of the External Hellenides. At least part of the rotational deformation can be associated with contractional deformation in the middle Miocene (Boronkay and Doutsos, 1994; Kokkalas and Doutsos, 2001, 2004; van Hinsbergen et al., 2005).

Seismicity and GPS Results in the Hellenic Peninsula

Pronounced seismic activity in the Aegean area concentrates along specific areas like the North Anatolia fault, with the North Aegean trough as its westward prolongation (Fig. 2; Koukouvelas and Aydin, 2002). High seismicity also characterizes the areas along the Hellenic arc, from the Kefallonia Fault zone to the Pliny and Strabo trenches, in a way defining the outline of the Aegean, Africa, and Anatolia plate boundaries (Fig. 1; Papazachos et al., 1984; Jackson et al., 1992; Huguen et al., 2001). In the last decade, various tectonic models have been proposed to unravel the evolution of the Aegean region and the tectonic processes that better explain its current deformation (McKenzie, 1972; Dewey and Şengör, 1979; Taymaz et al., 1991; Jackson, 1994; Le Pichon et al., 1995; Armijo et al., 1996; Hatzfeld et al., 1997; Meijer and Wortel, 1997; Doutsos and Kokkalas, 2001; Kreemer and Chamot-Rooke, 2004; Nyst and Thatcher, 2004). However, the role of specific structures in the current deformation still remains an open question.

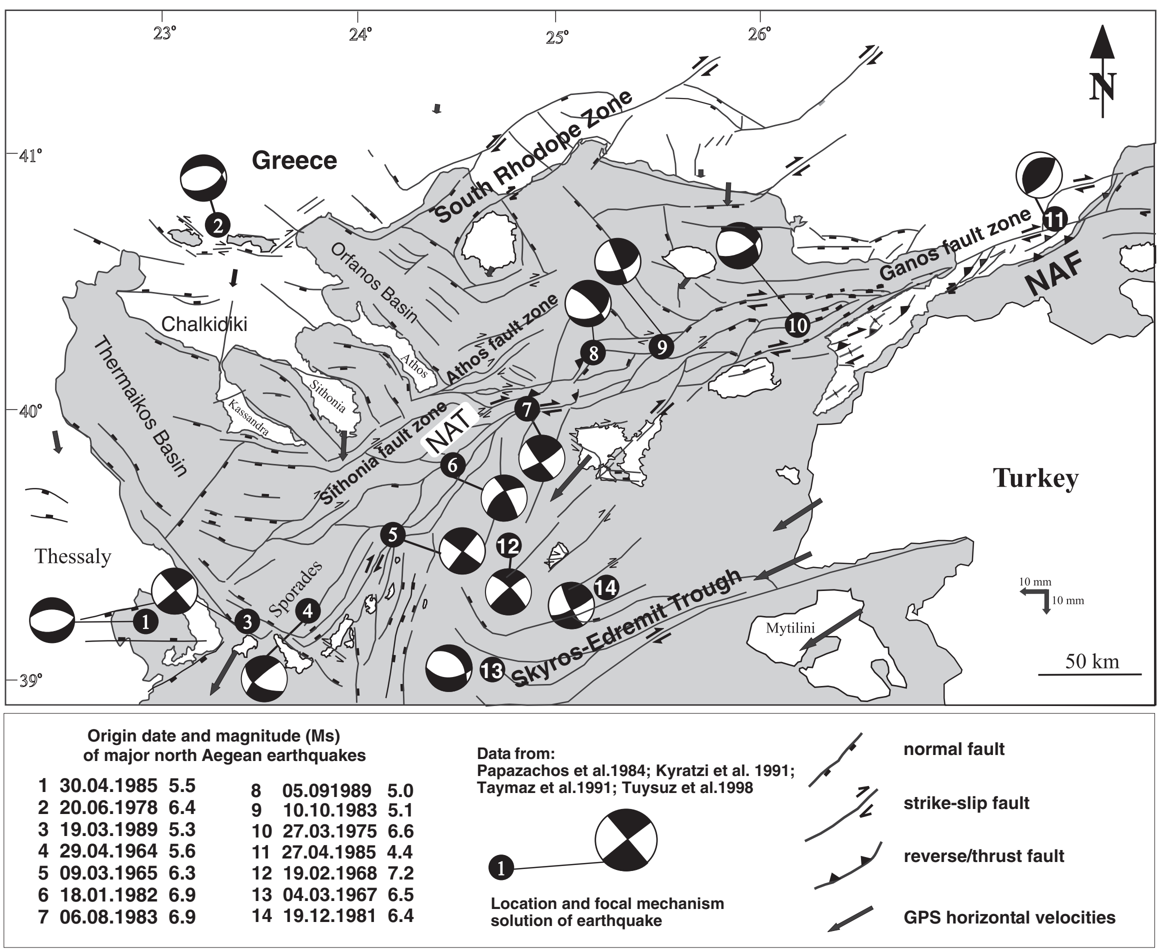

Figure 2. Tectonic map of the north Aegean area showing the main fault zones, the location and magnitude of major earthquakes in the north Aegean, and the mean GPS horizontal velocities in the area. NAF—North Anatolia fault; NAT—North Aegean trough. Map modified from Koukouvelas and Aydin (2002).

The GPS results from the Dalmatian-Albanian coastal region to northwest Greece indicate a belt affected by northeast-southwest shortening. GPS strain rates within this area are low, while focal mechanisms of moderate to strong shallow earthquakes indicate prominent thrust faulting (Mantovani et al., 1992). The zone of thrusting is possibly associated with the collision of the Adriatic block with the Eurasia plate. Seismicity data also provide no evidence of subduction along this part of the plate boundary. Thrusting continues southward to the Kefallonia fault zone, while farther to the south intermediate to oceanic subduction of the Africa plate occurs (Underhill, 1989). GPS results from western Greece show a small amount of motion north of the Kefallonia fault zone and homogenous southwest-directed rapid motion that reaches rates of 30–35 mm/yr in the western parts of the Peloponnesus and Crete (Kahle et al., 1998; Cocard et al., 1999). The dextral kinematics of the Kefallonia fault zone is clearly documented from the shear strain rates (Kahle et al., 1998), and it is associated with M > 7 strong earthquakes. Farther east, the seismicity resumes in the Greek mainland, but with normal fault mechanisms in the overriding plate.

For central Greece and especially for the Gulf of Corinth, GPS data show that the rift is currently extending at rates ranging between 14 mm/yr in the west and 10 mm/yr in the east (Briole et al., 2000). Deformation appears to be localized in a narrow deforming zone, especially in the west, where major faults transect the south coast of the gulf and the off-shore area (Koukouvelas et al., 2001; Stefatos et al., 2002). Microearthquake studies located a cluster of earthquakes at a depth of 10–12 km beneath the gulf (Rigo et al., 1996).

A GPS-determined velocity pattern similar to that of the western coasts of the Peloponnesus and Crete is displayed in the Aegean Sea area, where southwest-directed motion reaches rates of 28–33 mm/yr for the central and southern area. This pattern changes in the area north of the North Aegean trough and the Chalkidiki Peninsula, where negligible motion is recorded (Figs. 1 and 2; McClusky et al., 2000). The North Aegean trough marks the transition zone and separates a northern area of slow southwestward crustal motion from a southern area of rapid southwestward crustal motion (Figs. 1 and 2). The region, which includes mainly the Cycladic islands, has a very small amount of present and historical seismicity, and there is a clear deficit of seismic strain with respect to geodetic strain (Jackson et al., 1992; Jackson, 1994). This may be related to the high temperatures of the crust along the volcanic arc.

In the southeastern part of the Hellenic arc, GPS rates indicate an increasing component of sinistral strike-slip motion that correlates well with motion along the Pliny and Strabo trenches (Kahle et al., 1998).

The North Aegean Region

The north Aegean region and especially the North Aegean trough is a key structural area for understanding the evolution of the broader Aegean region (Fig. 2). Stratigraphic, structural, and geodetic data suggest that the North Anatolia fault propagated westward into the Aegean Sea during the early Pliocene (Barka, 1992; Dinter, 1998; Armijo et al., 1999). The crust underlying the North Aegean trough is the thinnest over the entire north Aegean area (Makris, 1977) and is characterized by Moho upwelling along the trough axis during the last 10 m.y. (Le Pichon et al., 1984).

The north Aegean area represents a domain with highly diffused deformation (Mascle and Martin, 1990; Pavlides and Caputo, 1994). This complex structural domain comprises three main ENE- to northeast-trending right lateral strike-slip fault zones with significant normal components (Fig. 2). These fault zones are, from south to north, (1) the Skyros-Edremit trough, (2) the North Aegean trough, and (3) the South Rhodope zone, all of which appear to be branches of the North Anatolia fault and probably follow pre-existing zones of weakness in the crust (Fig. 2). The right lateral strike-slip faults often terminate in well-defined extensional basins in their northwest quadrants (Koukouvelas and Aydin, 2002). These basins are controlled by northwest-trending normal faults (e.g., the off-shore Pelion area), with some of them displaying left slip components (e.g., faults from Limnos and Alonissos islands), as is confirmed by focal mechanisms of earthquakes in the area off-shore Chalkidiki Peninsula (Hatzfeld, 1999) and more recently by the Skyros (Mw = 6.4) earthquake (Ganas et al., 2005).

However, although we cannot ignore the significance of the deformation in this area, the general lack of specific data, including shallow coring and 3-D seismics within the North Aegean trough, limits our understanding in the area. Thus, our study concentrates mainly on the area along the Hellenic arc and trench system, primarily for two reasons: first, because the stratigraphy and the palinspastic evolution within the External Hellenides are well known, and second, because our data are more coherent regarding the syncollisional and extensional structures in this region.

In this context, our study aims to contribute to an understanding of how faulting in the upper crust accommodates this present-day velocity field, mainly along the Hellenic arc and trench system, and how orogenic crustal anisotropies control and affect the orientation of faulting and deformation of the overriding Aegean plate, passing from collision to extension.

STRUCTURAL REGION I: NORTHWEST GREECE

This collision belt is situated in the northwestern part of the Hellenic mountain chain between Corfu island and western Macedonia, where the Adria plate collides with the External Hellenides (Fig. 3; Mercier et al., 1972; Anderson and Jackson, 1987; Doutsos et al., 1987; Hatzfeld et al., 1995). This collision zone is separated from the west Hellenic subduction zone to the south by a large dextral strike-slip fault zone, the Kefallonia fault zone (Fig. 1, KFZ; Papazachos and Kiratzi, 1996; Tselentis et al., 1997; Cocard et al., 1999).

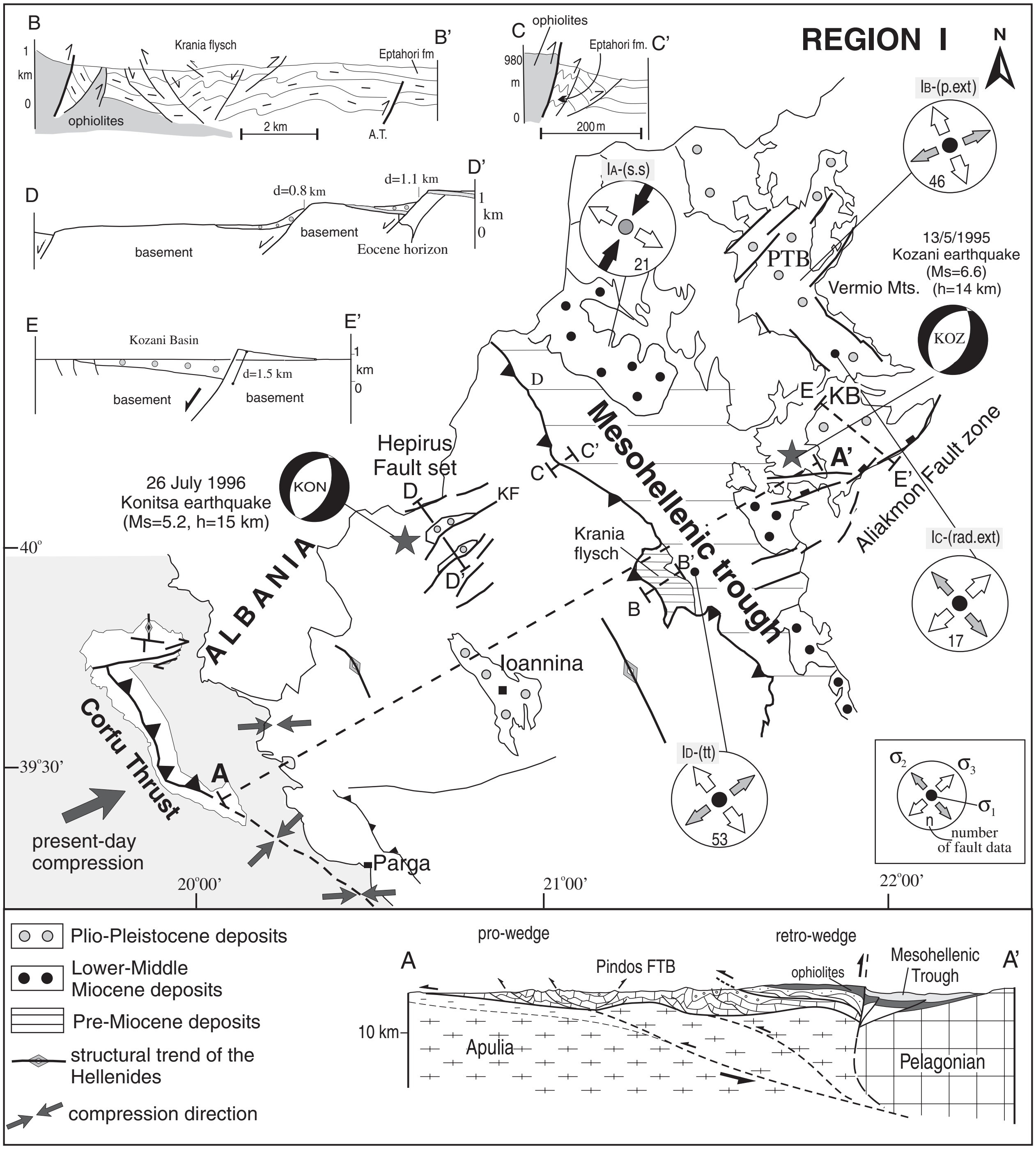

Figure 3. Simplified structural map of northwestern Greece (Region I), with cross-sections indicating the geometry and kinematics of the Hellenides in the area (section A-A′), the evolution of the Mesohellenic trough (sections B-B′ and C-C′), and the evolution of Plio-Pleistocene basins (sections D-D′ and E-E′). The orientation of principal stress axes in the synorogenic and younger deposits is also shown. In the stress nets, white outward double arrows indicate the σ_3 orientation, black inward double arrows the σ_1 orientation, and gray arrows the σ_2 orientation. Similarly, the vertical stress is shown by a solid circle with white color indicating a compressional regime, gray color a strike-slip regime, and black color an extensional regime. The numbers on the stress nets show the number of fault-slip data used for the determination of the stress regime, and the abbreviations in parentheses indicate the type of stress regime (p.ext—pure extensive; rad.ext—radial extensive; tt—transtensive; tp—transpressive; s.s.—strike-slip; comp.—compressive). KB—Kozani basin; KF—Konitsa fault; FTB—fold and thrust belt; PTB—Ptolemaida basin.

Contractional Structures

The northern part of the External Hellenides represents an asymmetric double-vergent orogen formed by the northeastward underthrusting of the Apulian microcontinent beneath the remnants of the Pindos Ocean and the Pelagonian microcontinent during the early Eocene. Subsequently, this margin was indented by the Pelagonian microcontinent and was thickened and uplifted, forming a bivergent fan thrust system (Doutsos et al., 1994) separating two domains: a retro wedge to the east from a pro wedge to the west (Fig. 3, section A-A′; Doutsos et al., 2004).

The retro wedge province is represented by the Mesohellenic trough, which is formed along the Apulian-Pelagonian suture zone as a piggyback basin ahead of three hinterland propagating thrust faults (Fig. 3, section A-A′). About 4 km of flysch and molassic sediments filled this basin in the time between the upper Eocene and the middle Miocene (Barbieri, 1992; Doutsos et al., 1994). Contraction is evidenced mainly along the trough's western margin, where the upper Eocene Krania flysch and the Oligocene Eptachori formation were overthrust by ophiolitic rocks along a dense pattern of upthrusts that are associated with very strong synthetic layer tilting of the downthrown block (Fig. 3, sections B-B′ and C-C′). Mesoscopic east-directed folds and fault-related fold structures are dominant close to the western border of the basin, declining progressively toward the east (Fig. 3, section B-B′). The sediments that are caught in the core of mesoscopic anticlines within the Krania flysch have a well-defined axial planar solution cleavage that dips steeply to the southwest.

The pro wedge province is represented by the Pindos fold and thrust belt and the Ionian basin filled with more than 3000 m of flysch sediments that were deposited in a piggyback style coeval with the southwest prograding folding and thrusting (Fig. 3, section A-A′). The fault system in the Ionian zone is characterized by broadly spaced arrays of both forward- and backward-verging thrust faults and Mesozoic reactivated faults that are sole-out in a gently dipping detachment at the base of the Triassic evaporites. Evidence for the age of this contraction is given by the deposition of upper Eocene flysch sealing major thrusts, as well as the presence of lower Miocene sediments in the core of synclines in front of the thrusts (IGRS-IFP, 1966; Skourlis and Doutsos, 2003). Recent seismological studies in the area showed a concentration of microseismic events at a depth of 5–10 km, implying that this detachment is active at the frontal part of the Ionian zone (Fig. 3, section A-A′; Martakis, 2003).

In general, the compressional structures are propagating toward the foreland and along the Ionian coast. At present, recent earthquakes located along the Albanian coast and extending to the Parga area are associated with horizontal compressional stress, and the slip vectors indicate shortening in a northeast-southwest to east-west direction, mainly along low-angle northwest-trending and hinterland-dipping thrust faults (Fig. 3; Louvari et al., 2001). Farther west, the Corfu thrust carries evaporites toward the west above a steeply dipping post–middle Miocene sequence (Doutsos and Frydas, 1994). The recent activity of the central segment of the Corfu thrust is demonstrated by a cluster of shallow seismic epicenters as well as by folded Tyrrhenian marine sediments (Doutsos and Frydas, 1994).

Extensional Structures

In the Middle Miocene, first-order northwest-trending faults and second-order northeast-trending faults accommodated extensional deformation parallel and perpendicular, respectively, to the axis of the Mesohellenic trough (Fig. 3 map, section B-B′, stress nets I_A, I_C, and I_D). The northwest-trending normal faults have strongly rotated the youngest molassic series of middle to late Miocene age (Brunn, 1956; Doutsos et al., 1994). Northwest-striking normal faults also controlled the early stages of formation of the Ptolemaida basin (Pavlides and Mountrakis, 1987) and caused the deposition of a late Miocene conglomerate sandstone sequence (Gregor and Velitzelos, 1995). The minimum-compression σ_3 axis determined from fault-slip data for the area has a northeast-southwest orientation (Pavlides and Mountrakis, 1987; Doutsos and Kokkalas, 2001; Fig. 3, stress net I_C).

In the Pliocene–Quaternary, the stress field in Region I changed significantly. Activity in this period was concentrated along two first-order structures: the Hepirus fault set in the northwest and the Aliakmon fault zone in the east (fault location and names as in Fig. 3 of Doutsos and Koukouvelas, 1998). Both fault systems comprise northeast-southwest-trending zones of primarily normal faulting. The Hepirus fault set truncates the fold and thrust belt in the northernmost part of Region I and comprises two northeast-trending asymmetric grabens spaced 8–10 km apart (Fig. 3, section D-D′). At the central part of the faults, maximum downthrow movements attain 1 km, while footwall uplift is on the order of 0.4 km (Doutsos and Koukouvelas, 1998; Fig. 3, section D-D′). Considering the 4 km cumulative heave of the three faults that comprise the Hepirus fault set, we calculated that total extension is on the order of 16%. The Konitsa fault (Fig. 3, KF) is likely linked to the 26 July 1996 (Ms = 5.2) Konitsa earthquake, which caused damage in the southern part of Konitsa village (Fig. 3, focal mechanism solution kon; Doutsos and Koukouvelas, 1998). Slip vector azimuths of 290° to 330° are calculated for the Konitsa fault along the central and western sections of the fault, respectively (Goldsworthy et al., 2002).

The 70-km-long, northeast-trending Aliakmon fault zone is marked by large, well-preserved fault scarps (Pavlides et al., 1995). The fault zone borders the Kozani basin, which is comprised of lacustrine and alluvial deposits of middle Pleistocene to Holocene age. These deposits make up the upper part of a 600 m southward-thickening sedimentary wedge (Fig. 3, section E-E′) that began to form in middle Pliocene time (Brunn, 1956; Koufos et al., 1991). The fault has a cumulative displacement on the order of ~1.5 km (Fig. 3, section E-E′; Doutsos and Kokkalas, 2001). Activity along a 25-km-long segment of this fault was evidenced during the 13 May 1995 (Ms = 6.6) Kozani earthquake (Fig. 3, focal mechanism solution koz), which caused a maximum coseismic slip of 0.5 m at a depth of 10 km (Pavlides et al., 1995; Resor et al., 2005). The interpreted fault network, which better explains the observed deformation on the surface of this area, consists of five significant segments (Resor et al., 2005). Furthermore, a conjugate system of nearly northeast-southwest-trending normal faults has internally deformed the sedimentary fill of the Ptolemaida basin (Fig. 3). Most of these faults are synsedimentary to the deposition of the early Pliocene lignite horizons (Schneider and Velitzelos, 1973; van de Weerd, 1979). These early Pliocene faults are compatible with an extensional stress field characterized by a σ_3 axis in a roughly northwest-southeast direction (Fig. 3, stress net I_B). This stress regime was also confirmed by the 13 May 1995 Kozani and the 26 July 1996 Konitsa earthquakes (Fig. 3, focal mechanism solutions kon and koz).

STRUCTURAL REGION II: CENTRAL GREECE

Contractional Structures

Contractional structures within the central part of the External Hellenides along the northern coast of the Gulf of Corinth show a pronounced orogenic polarity to the west (Fig. 4, section A-A′; Skourlis and Doutsos, 2003; Sotiropoulos et al., 2003). In the early Eocene, after the closure of a small oceanic strand of the Pindos Ocean, the Pindos zone was detached along the oceanic crust–sediments interface by the Pindos thrust (Degnan and Robertson, 1998) and overthrust westward onto the Tripolitsa zone. Fault-bend folds, duplexes, and imbricate thrust sheets characterize the internal deformation within the Pindos fold and thrust belt. Farther to the west, Mesozoic normal faults bordering the Ionian from the Tripolitsa zone were reactivated throughout the Oligocene (Fig. 4, section A-A′; Sotiropoulos et al., 2003).

In the Peloponnesus, convergence commenced in the middle Eocene with the eastward subduction of the Pindos Ocean beneath the Pelagonian microcontinent. In the early Oligocene, continental subduction along the Pelagonian margin was hindered and subduction migrated farther to the west along the weak crustal zone between Plattenkalk and the Tripolitsa zone. In the course of this intracontinental subduction within Apulia, the Plattenkalk and the Phyllite-Quartzite units were buried and suffered high-pressure, low-temperature (H*P*-L*T*) metamorphism (Seidel et al., 1982; Bassias and Triboulet, 1993). The exhumation history of the H*P*-L*T* rocks began at the Oligocene-Miocene boundary with the progressive entrance of continental crust in the subduction channel (Doutsos et al., 2000; Xypolias and Koukouvelas, 2001), resulting in the upward ductile extrusion of the Phyllite-Quartzite unit, derived from the deepest underthrust parts (Doutsos et al., 2000). An additional stage of deformation in the Peloponnesus involved gravitational nappe gliding, which was associated with lower Miocene orogenic uplift and final exhumation of the metamorphic units (Xypolias and Doutsos, 2000).

A transition between east-west shortening and north-south extension occurs in the area between the Ionian islands and the Gulf of Patras, respectively. Thrust faults with moderate eastward dip and elongated anticlines, which were formed from the upper Miocene to Pliocene–Quaternary time, comprise the compressional structures identified by field data and deep seismic profiles (Underhill, 1989; Kamberis et al., 1996). Evaporitic layers in the area of contact between the Ionian and Pre-Apulian zones were thrust upon Pliocene sediments in the off-shore area between Zakynthos Island and Peloponnese. The sedimentary sequence of early Miocene to Quaternary, which extends over most of the area, has a significant thickness of ~6 km (Monopolis and Bruneton, 1982). Furthermore, on Lefkada island the Ionian thrust was active during the late Miocene–early Pliocene, causing folding in the Pre-Apulian zone sediments in the footwall of the thrust. Field observations made after the 14 August 2003 (Ms = 6.4) Lefkada earthquake showed indications of recent reactivation along the NNE-SSW-trending faults parallel to the Ionian thrust, with a sinistral sense of motion (Kokkalas et al., 2003).

On the mainland coast opposite Lefkada island, the Mesozoic basement overthrust the upper Miocene–lower Pliocene sediments along the margins of Paleros basin (Fig. 4, section B-B″, stress net II_C; Doutsos et al., 1987). The sediments at the footwall of these thrusts, which display a roughly dip-slip motion, have suffered major flexural folding with NNW- to NNE-trending axes accompanied by NNW-trending cleavage planes and low-angle thrusts. Farther to the south, the compressional zone is submerged, and all of the western coast of the Peloponnesus is affected by extensional structures (Koukouvelas et al., 1996).

Extensional Structures

Extension in Region II is accommodated by two main fault trends: ENE- to northeast- and WNW-trending faults (Fig. 4). The first fault system prevails in the western Peloponnesus and began to develop early, as in the case of the Pyrgos basin area, where the Lapithas and Vounargon faults (Fig. 4, LF and VF, respectively) caused the deposition of post–late Miocene sediments (Koukouvelas et al., 1996). Farther to the north, the ENE- to northeast-trending Rio graben, which has a length of ~30 km and a width of ~10 km, displays a maximum subsidence on the order of 1.8 km. Although faults of this direction remain active today, they display slip rates that decrease with time (Koukouvelas et al., 1996).

In the eastern Peloponnesus and in south Sterea Hellas, this ENE-trending fault system seems to be weaker and accommodates only a small amount of the extension in the area, as is the case at the eastern end of Corinth graben (Fig. 4, Klenia fault zone [KF]; Koukouvelas et al., 1999). East of the Gulf of Corinth, faulting continues in an ENE direction along a zone of diffuse

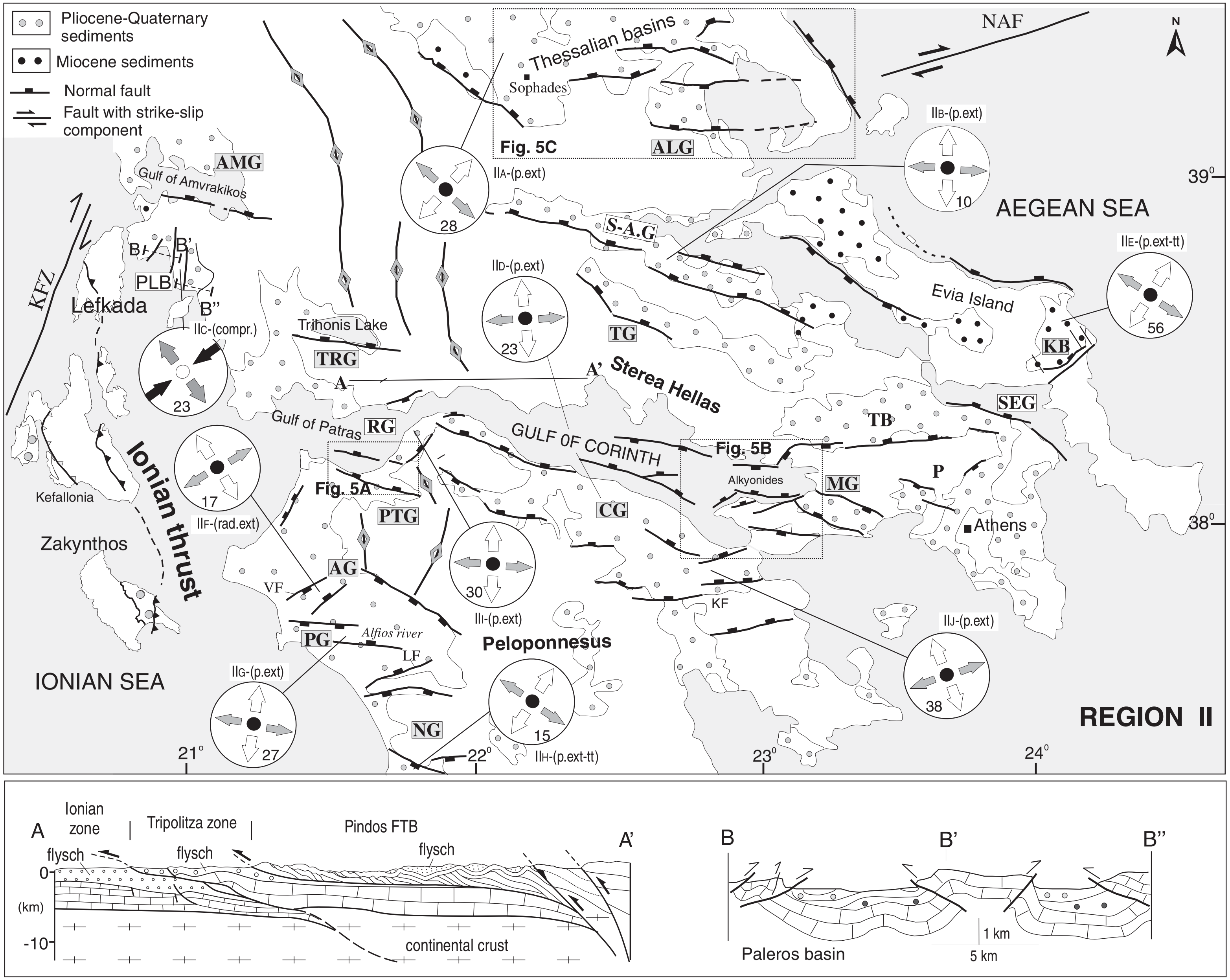

Figure 4. Schematic structural map of the central Hellenic Peninsula (Region II), with stress nets showing the orientation of principal stress axes. Stress net explanation as for Figure 3. Also included are cross-sections showing the geometry and kinematics of the External Hellenides in the area (A-A′) and the evolution of the synorogenic basin in the Paleros area (B-B′-B″). AG—Abelon graben; ALG—Almyros graben; AMG—Amvrakikos graben; CG—Corinth graben; KB—Kymi basin; KF—Klenia fault zone; KFZ—Kefalonia fault zone; LF—Lapithas fault; MG—Megara graben; NG—Nedas graben; P—Parnitha area; PG—Pyrgos graben; PLB—Paleros basin; PTG—Patras graben; RG—Rio graben; S-A.G—Sperchios-Atalanti graben; SEG—South Evoikos graben; TB—Thiva basin; TG—Tithorea graben; TRG—Trihonis graben; VF—Vounargos fault.

deformation through Thiva basin to the South Evia graben, but also in a WNW direction, as the recently activated Parnitha area (Fig. 4, P; Ganas et al., 2004). The South Evia graben is a WNW-trending graben ~50 km long and 15 km wide that it is thought, based on off-shore stratigraphy, was developed in the last 1 m.y. (Perissoratis and van Andel, 1991). The South Evia graben marginal fault is segmented into five prominent segments with lengths that vary from 5 to 15 km (Goldsworthy et al., 2002).

In western Greece, the WNW-trending faults controlled the deposition of Pleistocene sediments and accommodate part of the general extension, as is indicated by the slightly rotated beds and displacements of up to 500 m. This WNW-trending faults are presently active, as was revealed by recent seismicity in the area, and exert strong geomorphologic control on the present landscape, including river courses (e.g., Alfios River), lakes (e.g., Trihonis Lake), and shallow gulfs (e.g., Patras and Amvrakikos Gulfs; Figs. 4 and Fig. 5A).

Farther east, the area most representative of active deformation is the Gulf of Corinth (Fig. 4). Extension has been largely accommodated by WNW-trending normal faults since the upper Miocene. In the Gulf of Corinth and adjacent areas, two major phases of rifting are recognized (Doutsos and Piper, 1990). The older, during the Pliocene, formed shallow basins, although much more rapid fault displacement took place in the Pleistocene–Holocene. The latter started perhaps as early as the early Pleistocene in the Gulf of Corinth, but as late as ca. 0.5 Ma in the Gulf of Argolis (van Andel et al., 1993) and in the Gulf of Patras (Piper et al., 1990). The grabens propagated gradually westward (Doutsos et al., 1988) through the northern Peloponnesus and Sterea Hellas in the Quaternary, perhaps as a result of basin deformation in a "pull-apart" style between the stable Europe and the rapidly southwest-moving south Aegean area.

The WNW-trending faults control, from south to north, four typical grabens: the Corinth, Tithorea, Sperchios-Atalanti, and Almyros grabens (Fig. 4, CG, TC, S-A.G, and ALG). The asymmetry of the grabens is largely induced by north-dipping basin-bounding faults that are segmented along their strike. In the western Corinth graben, a series of north-dipping normal faults form a half graben that shows an increase of deformation toward the graben axis (Koukouvelas and Doutsos, 1996; Avallone et al., 2004; McNeill and Collier, 2004). The faults reach a depth of 10–12 km and show displacements of up to 1000 m associated with wedge-shaped sedimentary prisms deposited in the Quaternary (Doutsos and Piper, 1990; Micarelli et al., 2003). The sediments in the Gulf of Corinth have a thickness of ~1.7 km (Clement et al., 2004). The maximum structural relief between the lowest bedrock areas under the gulf and the adjacent mountains is ~4 km. Farther east, marine terraces on the southern coast of the gulf were uplifted 600 m above sea level during the last 500,000 yr, indicating uplift rates on the order of 1 mm/yr (Jackson and McKenzie, 1984; Keraudren and Sorel, 1987). However, a large number of smaller faults control sedimentation and the characteristic morphological steps of these marine terraces (Doutsos and Piper, 1990; Xypolias and Koukouvelas, 2004).

Stress tensor analysis in Region II shows a consistent σ_3–axis in a roughly north-south direction (Fig. 4, stress nets

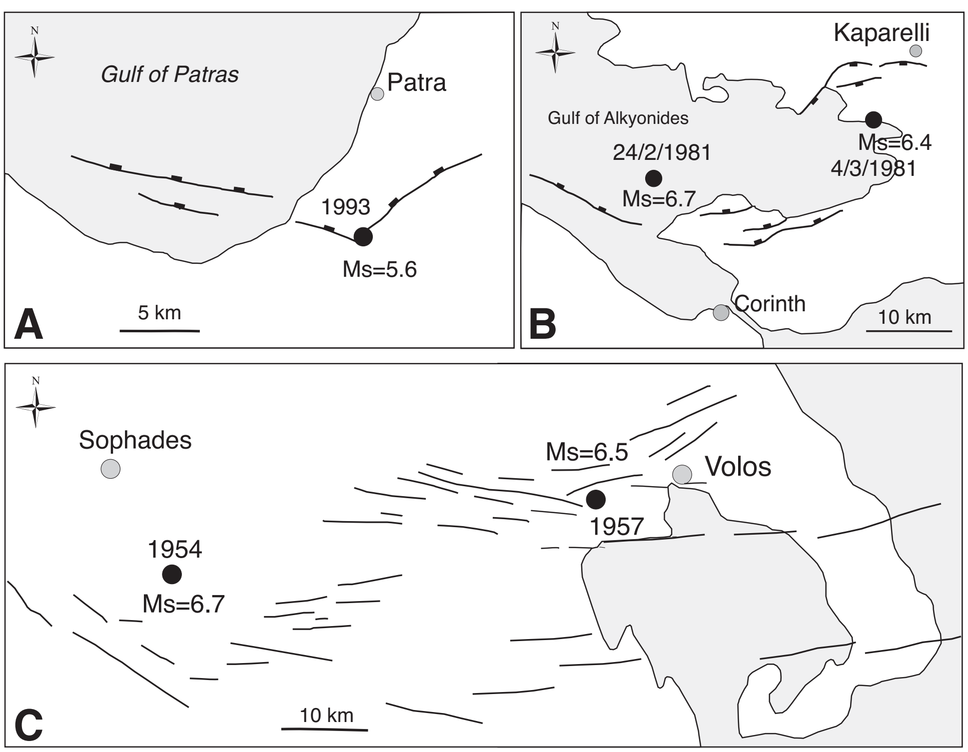

Figure 5. Schematic maps of (A) the Gulf of Patras; (B) the Alkyonides area, eastern Corinth Gulf; and (C) the Volos area, showing the interaction of active WNW- and ENE-trending normal faults and the associated seismicity (modified from Poulimenos and Doutsos, 1996).

II_B, II_D, II_I, and II_G). However, deviations from this general direction may be due to the presence of ENE-trending faults and their interaction with the WNW-trending fault system (Fig. 4, stress nets II_A, II_F, II_H, and II_J). In cases in which one fault set is dominant, the σ_3 trends northeast (Fig. 4, i.e., stress nets II_A and II_H), corresponding to northwest-oriented grabens (e.g., Nedas, southwest Thessalian basin), or northwest (Fig. 4, i.e., stress nets II_F and II_J), as in the case of the ENE-trending Ambelon and east Corinth grabens, respectively. However, in many cases, such as in the Pyrgos and Patras graben areas, where both sets of faults are active, σ_3 trends nearly north-south, bisecting the obtuse angle of the two fault sets (Fig. 4, stress nets II_I and II_G).

Furthermore, the diffused zone at the junction between the ENE- and WNW-trending faults seems to be particularly important in terms of seismicity, for differential displacement along these faults causes strong earthquakes (Fig. 5A, B, and C). The concept that the bend of fault zones acts as initiator of large earthquakes appears to be confirmed in this case. Typical examples are the 14 July 1993 (Ms = 5.6) earthquake in the Patras area (Fig. 5A), the 24 February 1981 (Ms = 6.7) earthquake in the Gulf of Alkyonides (Fig. 5B), and the 30 April 1954 (Ms = 6.7) Sophades earthquake (Fig. 5C; Poulimenos and Doutsos, 1996). More specifically, the Sophades earthquake caused well-exposed ruptures near the junction of WNW- and ENE-trending normal faults, while its epicenter occurred near the bend of the two fault orientations (Fig. 5C; Pavlides, 1993).

A basin particularly interesting for the structural evolution of the eastern part of Region II is the Kymi basin (Evia island), because it has had a much longer evolution, from the lower Miocene to the present, than other basins in the area and is affected by both contraction and extension (Fig. 4, KB). Its Plio-Pleistocene extensional deformation better resembles that of Region II, where mesoscopic WNW-trending faults were formed under tension in a northeast direction, while the ENE-trending faults were reactivated, displaying oblique normal motion along them (Fig. 4, stress net II_E; Kokkalas, 2001). In contrast, the Miocene evolution of the Kymi basin better resembles that of the molassic basins of the central part of the Hellenic arc (Region IV), and therefore it will also be discussed in the section on that region.

STRUCTURAL REGION III: SOUTHWEST HELLENIC ARC

Contractional Structures

The contractional structures discussed in the section on Region II continue farther to the south in this region and show a similar deformation style. These structures were developed as the end result of the intracontinental subduction of the Phyllite-Quartzite and Plattenkalk units (H*P*-L*T* belt), which continued southward to the south Peloponnesus, and their buoyant uplift and exhumation during the early Miocene (Fig. 6, section A_1-A_2-A_3; Doutsos et al., 2000). Therefore, in discussing this region we focus only on the extensional structures.

Extensional Structures

An area with a distinct stress and deformation pattern is extending from the south Peloponnesus through Kythira island to western Crete. This area is located behind the Matapan trench (Fig. 6). At the north Matapan trench, an important change in the subduction process takes place, with subduction of intermediate to oceanic crust in the north and more oceanic crust to the south (Le Pichon and Angelier, 1979; Lyberis and Lallemant, 1985). Beneath this area, the subduction zone dips gently to the northeast, changing to a steeper dip farther northeast beneath the Argolikos Gulf (Hatzfeld, 1994; Papazachos et al., 2000).

The dominant fault directions have a NNW-SSE to northwest-southeast orientation in the south Peloponnesus and a northeast-southwest to NNE-SSW orientation in the westernmost part of Crete. The northwest-trending faults are aligned almost parallel to the Matapan trench. In the south Peloponnesus, northwest-trending faults control the accumulation of sediments of post–late Pliocene age (Frydas, 1990). They remain active today and divide the area into three peninsulas. The largest one, the Sparta fault (Fig. 6, SF), trends parallel to the eastern front of Taygetos Mountain (Fig. 6, section B-B′) and forms a steep, 700-m-high escarpment that has been reactivated in historical times (464 B.C.), and it is inferred to have hosted an earthquake event of magnitude Ms = 7 (Armijo et al., 1991). It is argued that the major factor contributing to the building of Taygetos Mountain was flexural uplift with viscoelastic relaxation resulting from the mechanical unloading of the crust during extension (Poulimenos and Doutsos, 1997; Doutsos et al., 2000). Estimated footwall uplift and maximum downthrow along the Sparta fault are ~1.2 km and ~2.5 km, respectively (Doutsos et al., 2000). To the south, on Kythira island, these grabens are filled with early upper Miocene marine deposits (Meulenkamp et al., 1988) and remain active today, as is indicated by prominent and fresh fault scarps (Fig. 6, Kythira Isl., stress net III_B).

The north-NNE-trending normal and oblique normal faults form young scarps in the southern Peloponnesus and the offshore Kythira area. At the northernmost end of the Kalamata basin, small north-south-trending fault scarps that dip steeply to the west controlled the deposition of Pliocene–Pleistocene sediments. Normal slip sense along the NNE-trending faults is also documented by the Kalamata earthquake on 13 September 1986 (Ms = 6.2), which produced a 6-km-long NNE-trending seismic fault on the western flank of Taygetos Mountain (Lyon-Caen et al., 1988). These faults are compatible with a pure extensive stress regime, with the σ_3 axis trending almost east-west (Fig. 6, stress nets III_A, III_C, and III_D).

Farther southward, in western Crete, northwest-trending faults are scarce, whereas NNE-trending faults become more important and have controlled, since the late Miocene, the evolution of off-shore (Mascle et al., 1982) and on-shore basins (Fig. 6).

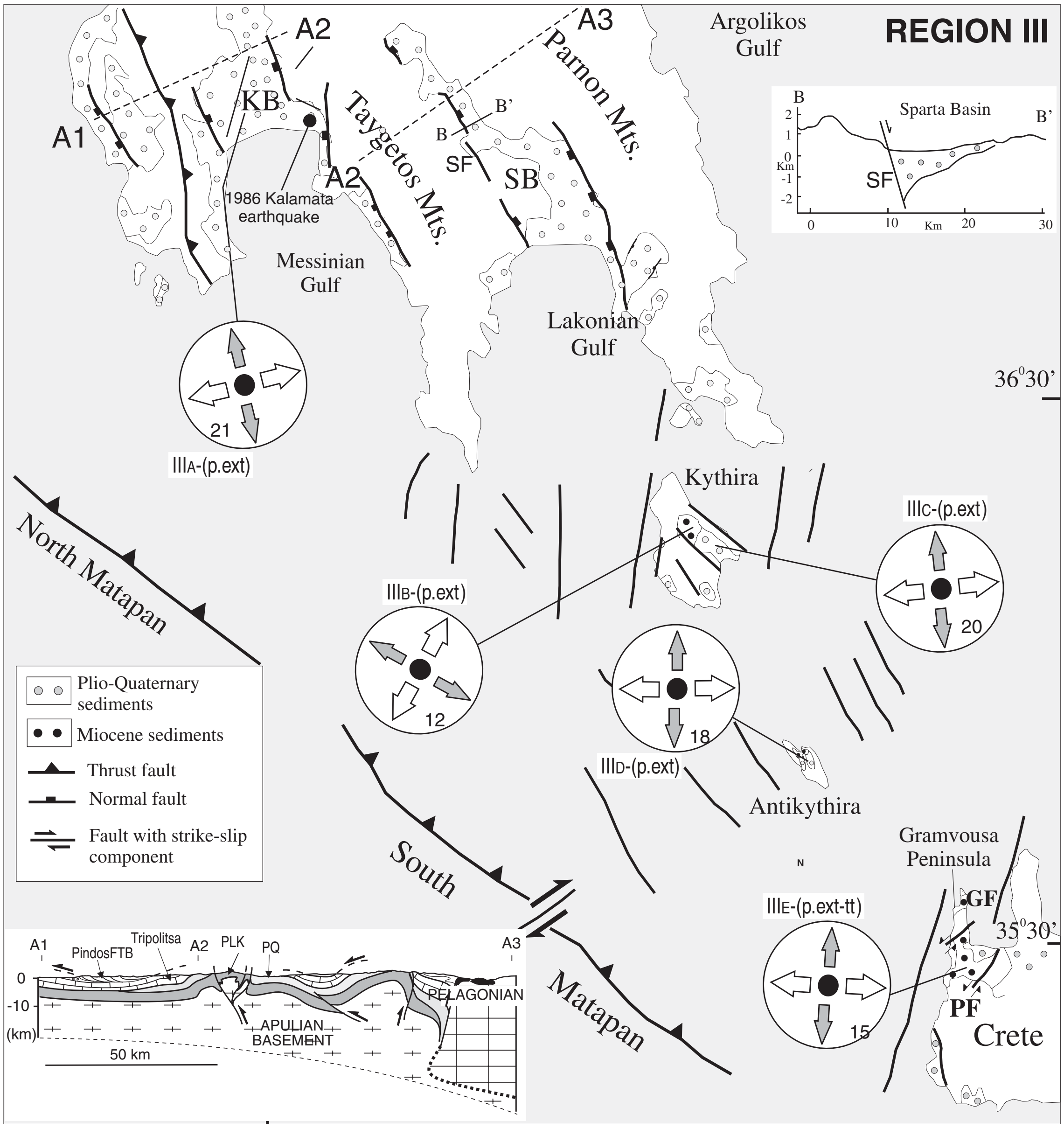

Figure 6. Simplified map of the southwestern Aegean area (Region III). Synthetic cross-section A1-A3 in the panel at bottom left shows the geometry and kinematics of the External Hellenides and tectonic windows formed in the area. Cross-section B-B′ displays the geometry of the Sparta basin. Stress net explanation as for Figure 3. GF—Gramvousa fault; KB—Kalamata basin; PF—Platanos fault; SB—Sparta basin; SF—Sparta fault.

In western Crete, sinistral oblique normal NNE-trending and steeply dipping faults control the deposition of Tortonian sediments (Fig. 6, PF and GF; Frydas and Keupp, 1996). These NNE-trending faults show both oblique normal and normal components of motion. The σ_3 axis deduced from fault-slip analysis is trending east-west (Fig. 6, stress net III_E).

The majority of the earthquake events in the brittle upper crust in the western part of Crete, for which fault plane solutions were determined, show a horizontal east-west-trending T axis. The P axes of these events are oriented essentially north-south, corresponding to the regional stress field of the subduction process, and vary from horizontal plunges indicating strike-slip events to nearly vertical plunges indicating normal faulting (Taymaz et al., 1990; De Chabalier et al., 1992).

Taymaz et al. (1990) were the first to propose a stratification of earthquake mechanisms, including extension with an east-west-oriented T axis in the overriding crust, low-angle thrusts with north-dipping planes interpreted as being the contact surface of subduction at ~35–41 km depth, and slab-pull events with T-axes in a north-south orientation deeper within the Africa plate. This heterogeneity of fault plane solutions in a subduction zone may have been caused in part by ruptures on pre-existing zones of weakness not favorably oriented to the current stress field. Geological evidence supports the existence of variously oriented pre-existing zones of weakness in the uppermost crust of western Crete (e.g., Angelier et al., 1982). Thus, the variability of focal mechanisms indicates a plate-slab decoupling in the vertical crustal scale that can be explained by the presence of critically stressed fault zones throughout the brittle crust (see also Zoback and Harjes, 1997).

Deformation along the NNW- to northwest-trending faults seems to have started in the upper Pliocene in the Kalamata basin and the Messinian Gulf (Zelilidis and Doutsos, 1992), while farther to the east, in the Argolikos Gulf, a late Quaternary age of deformation is suggested (Piper and Perissoratis, 2003). This may indicate a diachronous onset of deformation along the NNW and northwest-trending faults, linked to the process of slab retreat below the south Peloponnesus. The north- to NNE-trending fault system is widespread and can be interpreted either as the result of incipient collision with buoyant African crust in this part of the arc (Lyon-Caen et al., 1988) or as a consequence of strain partitioning during oblique subduction (Le Pichon et al., 1995).

STRUCTURAL REGION IV: CENTRAL PART OF THE HELLENIC ARC

This area includes the southern part of Evia island and the Cycladic islands (the Attico-Cycladic massif), the Cretan Sea, and the central part of Crete (Figs. 1 and 7A). It lies opposite the central part of the Mediterranean Ridge accretionary complex (see Fig. 1, MRAC), which collided after the lower Pliocene with the African continental margin (Le Pichon et al., 1995). Seismic surveys (Wortel and Spakman, 2000; Bohnhoff et al., 2001) indicate a continental crust with a thickness of 32 km below central Crete, while to the north the crustal thickness decreases significantly, to 15 km below the Cretan Sea. Lateral variations in crustal and sedimentary thickness were also identified in an east-west direction (Bohnhoff et al., 2001). The slab is slightly inclined beneath Crete and steepens abruptly beneath the Cretan Sea (Wortel and Spakman, 2000).

Several models have been proposed for the orogenic evolution of the Hellenides, with the majority of these focusing mainly on the exhumation mechanisms of deep-seated metamorphic rocks. Up to now, various mechanisms have been proposed, such as (1) syn- and/or post-orogenic extension, (2) extension related to subduction retreat (or roll-back), and (3) ductile extrusion occurring in an overall compressional setting.

The first model suggests that rapid uplift and exhumation of the H*P* rocks is occurring along the footwall of shallow-dipping extensional detachment faults. This mechanism was first proposed by Lister et al. (1984) for the Cyclades blueschist belt, and due to its broad acceptance, it was consequently applied in many regions along the Hellenides, from the Rhodope area in the north (Dinter and Royden, 1993; Sokoutis et al., 1993) through the central Aegean area and its surroundings (Avigad and Garfunkel, 1989; Gautier and Brun, 1994; Avigad et al., 1997; Ring et al., 1999) to Crete in the south (Kilias et al., 1994; Ring et al., 2001).

The second model also accepts the importance of major extensional detachments in the exhumation of H*P*-L*T* rocks but gives more emphasis to the role of progressive or continuous slab retreat (roll-back) of the subducting plate during the collision and exhumation process (Thomson et al., 1998; Ring and Layer, 2003).

The third model, which constitutes an alternative view, suggests that the H*P*-L*T* rocks in the External Hellenides (from Thessaly to the Peloponnesus, Evia, and Crete) exhumed with a mechanism comprising solid-state ductile extrusion in an overall compressional setting. Backthrusting and/or transpression during a middle Miocene late orogenic stage completed this process (Doutsos et al., 1993, for Thessaly; Xypolias and Doutsos, 2000, and Xypolias and Koukouvelas, 2001, for the Peloponnesus; Xypolias et al., 2003, for Evia; Kokkalas and Doutsos, 2000 and 2004, for Crete).

As was emphasized by Thomson et al. (1998), convincing exposures of the major extensional detachments are either obliterated by recent deformation or covered by the sea and thus are difficult to find. The presence of these detachments relies mainly on the combination of indirect metamorphic and geochronological data and is not typically supported by field structural data. Besides our work, other recent studies in Crete have questioned the presence and kinematics of such extensional detachments (Zulauf et al., 2002; Campell et al., 2003). Oblique contractional deformation in the Aegean is also supported by (1) recent block rotations around vertical axes (Morris and Anderson, 1996), (2) folded middle Miocene sediments along northeast-trending strike-slip faults recognized in off-shore seismic data (Mascle and Martin, 1990), and (3) transpressional or strike-slip faults that control the evolution of middle Miocene molassic basins, as well as the ascent and emplacement of granite intrusions (Boronkay and Doutsos, 1994; Koukouvelas and Kokkalas, 2003). Finally, Avigad et al. (1998) argue that strike-slip faults within the Aegean Sea are underestimated.

Contractional Structures

In the Aegean region, the collision of several microplates with irregular margins occurred from the early Eocene to the early Miocene (Boronkay and Doutsos, 1994; Shaked et al., 2000; Avigad et al., 2001; Xypolias et al., 2003). A regional ENE-trending stretching lineation is identified in the units of the Aegean metamorphic complex (Fig. 7A). Even in the southern Cyclades, where an almost north-south-trending stretching lin-

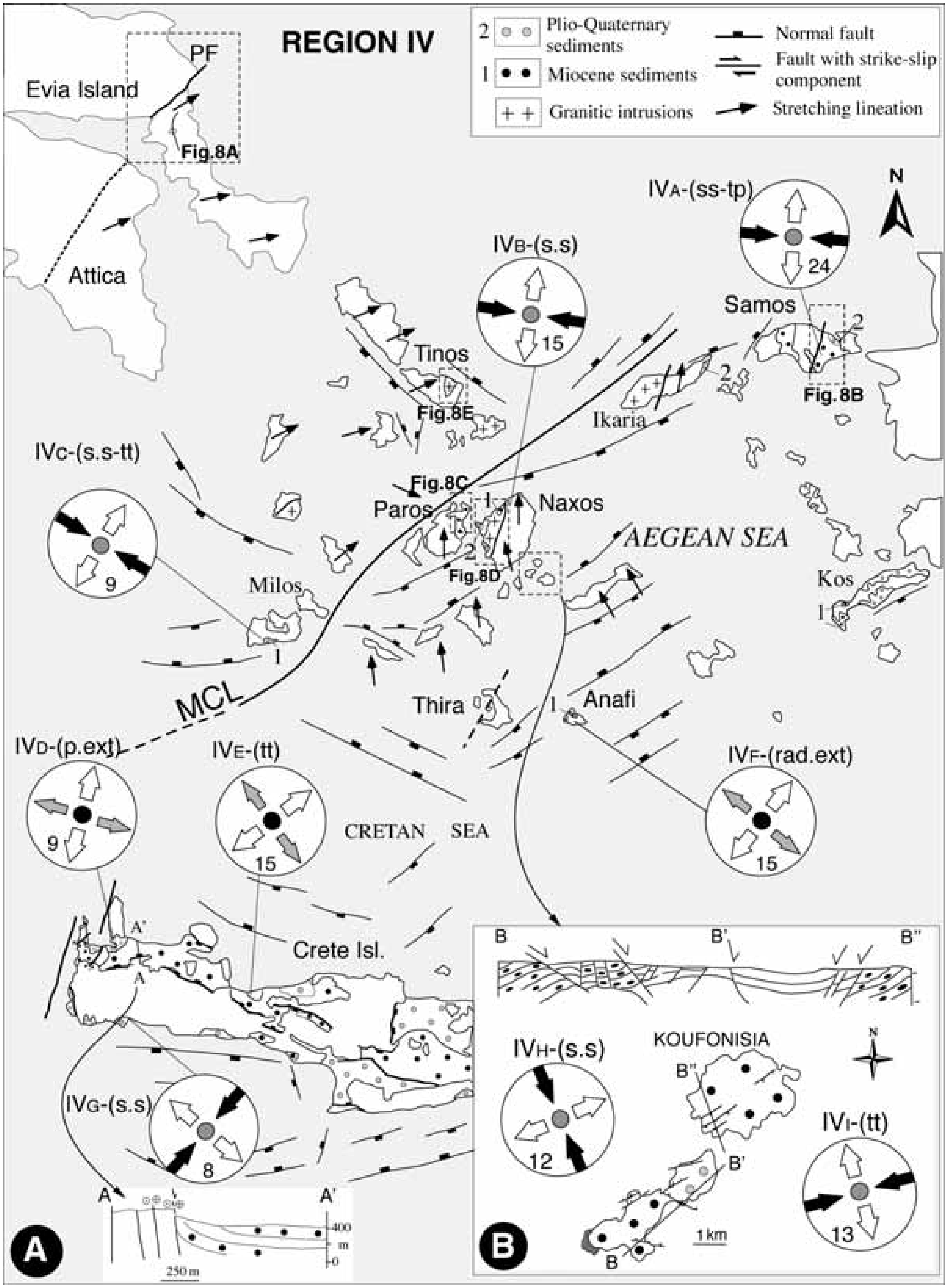

Figure 7. (A) Simplified structural map and cross-sections in the central part of the Hellenic arc (Region IV). Inset (B) shows the deformation and stress regimes in the Koufonisia islands. Stress net explanation as for Figure 3. Seafloor lineaments after Mascle and Martin (1990); Piper and Perissoratis (2003). MCL—Mid-Cycladic lineament; PF—Pelagonian fault.

eation is prominent, there are places where the earlier ENE-orientation is also identified. The two domains with contrasting orientations of stretching lineation are separated by a significant crustal lineament, the Mid-Cycladic lineament (Fig. 7A, MCL; Walcott and White, 1998). Supporting the existence of the Mid-Cycladic lineament as a significant crustal structure are the palaeomagnetic data from middle and late Miocene sediments and intrusions, which show remarkable amounts of block rotation within the Cyclades and differential movement along both sides of the Mid-Cycladic lineament, as well as the presence of magmatic intrusions along strike (Avigad et al., 1998; Walcott and White, 1998). The northeast-southwest-trending Mid-Cycladic lineament marks the boundary between a western block that experienced 20°-30° of clockwise rotation and an eastern block that rotated ~19° counterclockwise during the Miocene and ceased to be active, probably in the late Miocene (Le Pichon et al., 1995).

Because the ENE direction of convergence was nearly parallel or slightly oblique to the trend of microplate margins, lateral ramps have been formed that gave rise to a dextral transpressional setting, as is recorded within Miocene synorogenic granitic intrusions and late orogenic molassic basins (Figs. 7A and 8; Boronkay and Doutsos, 1994; Koukouvelas and Kokkalas, 2003). The uplift and exhumation of H*P*-L*T* rocks was achieved in the hangingwall of these ramps at steeply dipping, highly strained shear zones, in which crustal material escaped both laterally and vertically (Fig. 8A; Boronkay and Doutsos, 1994; Xypolias et al., 2003). On Evia island, intercontinental collision began during the Eocene involving the subduction of the protolith of the Blueschist unit beneath the Pelagonian microcontinent. During the Oligocene, a slowdown or eventual cessation of subduction occurred as continental material entered the subduction channel. This resulted in the detachment of the Blueschist unit from its basement and the subsequent extrusion of H*P*-L*T* rocks (Fig. 8A, evolutionary sections A-A′-A″). At the Oligocene-Miocene boundary, the Blueschist unit was tectonically emplaced over the mildly metamorphosed carbonate rocks of the Almyropotamos unit (Fig. 8A).

From the early Miocene to the middle Miocene, the nappe pile in the northern part of central Aegean was affected by transpressional deformation, which led to the development of the Pelagonian fault. Dextral transpressional shearing along this crustal scale fault caused further uplift and doming of the H*P* rocks (Fig. 8A; Xypolias et al., 2003). This northeast-southwest-trending faulted Pelagonian margin continues farther to the southwest in the Attica Peninsula (Fig. 7A, dashed line). Transpressional movements caused the formation of the synorogenic Kymi basin in the early Miocene and controlled the accumulation of the Lower Formation (Aquitanian to lower Serravalian), comprising ~500-m-thick syntectonic sediments (Fig. 8A, lithostratigraphic section). The western margin of the basin is marked by a NNE-trending arcuate right lateral transpressional fault, the Kymi fault (Kokkalas, 2001), resulting in a vertical structural relief of more than 1300 m (Fig. 8A). The westward thickness increase of the Lower Formation within the Kymi basin implies a synsedimentary origin for this fault. Transpressional faults controlled and also deformed the 13 Ma Oxylithos volcanics (Pe-Piper and Piper, 1994). Slip data on mesoscopic faults are consistent with a northeast-southwest direction of compression (Fig. 8A, stress net IV_J).

On Samos island, a highly segmented transpressional fault, the Mavradzei fault, which has a northeast strike, carried crystalline basement rocks above lower Tortonian sediments (Fig. 8B; Boronkay and Doutsos, 1994). The Mavradzei fault is a 16-km-long dextral oblique thrust that produced in its footwall a 1.3-km-wide fault zone internally deformed by thrusts and mesoscopic folds. Its central part rotated clockwise into a north-south position by means of four ENE-trending transcurrent faults with right lateral displacement (Fig. 8B). On Paros island to the west, a NNE-trending molassic basin lies uncomformably above an ophiolitic basement. It consists of conglomerates, sands, and clays of Burdigalian age underlain by upper Miocene calcareous sandstones and conglomerates (Dermitzakis and Papanikolaou, 1980). The faults that affect the sediments show a transpressional character, while folding is confined to narrow north-south-trending zones near the faults (Fig. 8C, section B-B′). Two coeval fault sets comprising northwest-trending dextral and ENE-trending sinistral oblique thrust were identified on the island related to a transpressional regime characterized by an ESE compression direction and bisecting the acute angle of the two fault sets (Fig. 8C, stress net IV_L).

The lower parts of the upper crust have also been affected by transpression, as is suggested by structural analysis of plutons. The transpressional structures comprise a conjugate system of NNE-trending dextral faults and northwest-trending sinistral faults. The NNE-trending fault system is very prominent in the central Aegean, controlling the intrusion of granitic plutons during the middle–late Miocene (13–10 Ma), as is clearly observed on Naxos island (Figs. 7A and 8D; Boronkay and Doutsos, 1994; Pe-Piper et al., 1997; Koukouvelas and Kokkalas, 2003). On Naxos, kinematic and strain analysis suggest that the NNE-trending dextral Naxos fault zone controlled the emplacement and deformation of the granodiorite under a transpressional regime. Additionally, the northeast-trending faults were active during pluton formation, as is evidenced by the deflection of both synmagmatic foliation trajectories and flow lines along these faults (Fig. 8D; Koukouvelas and Kokkalas, 2003). Another fault system comprising northwest-trending oblique thrusts and left lateral faults deformed and offset pluton exposures on Serifos, Tinos, and Mykonos islands under brittle ductile conditions (Boronkay and Doutsos, 1994). On Tinos, a granitic body intruded basement rocks in the middle Miocene (15–14 Ma; Altherr et al., 1982), with northeast-trending faults controlling its contact zone (Fig. 8E). The northwestern border of the granitic body intruded along a thick, highly strained zone that also included WNW-trending mylo-

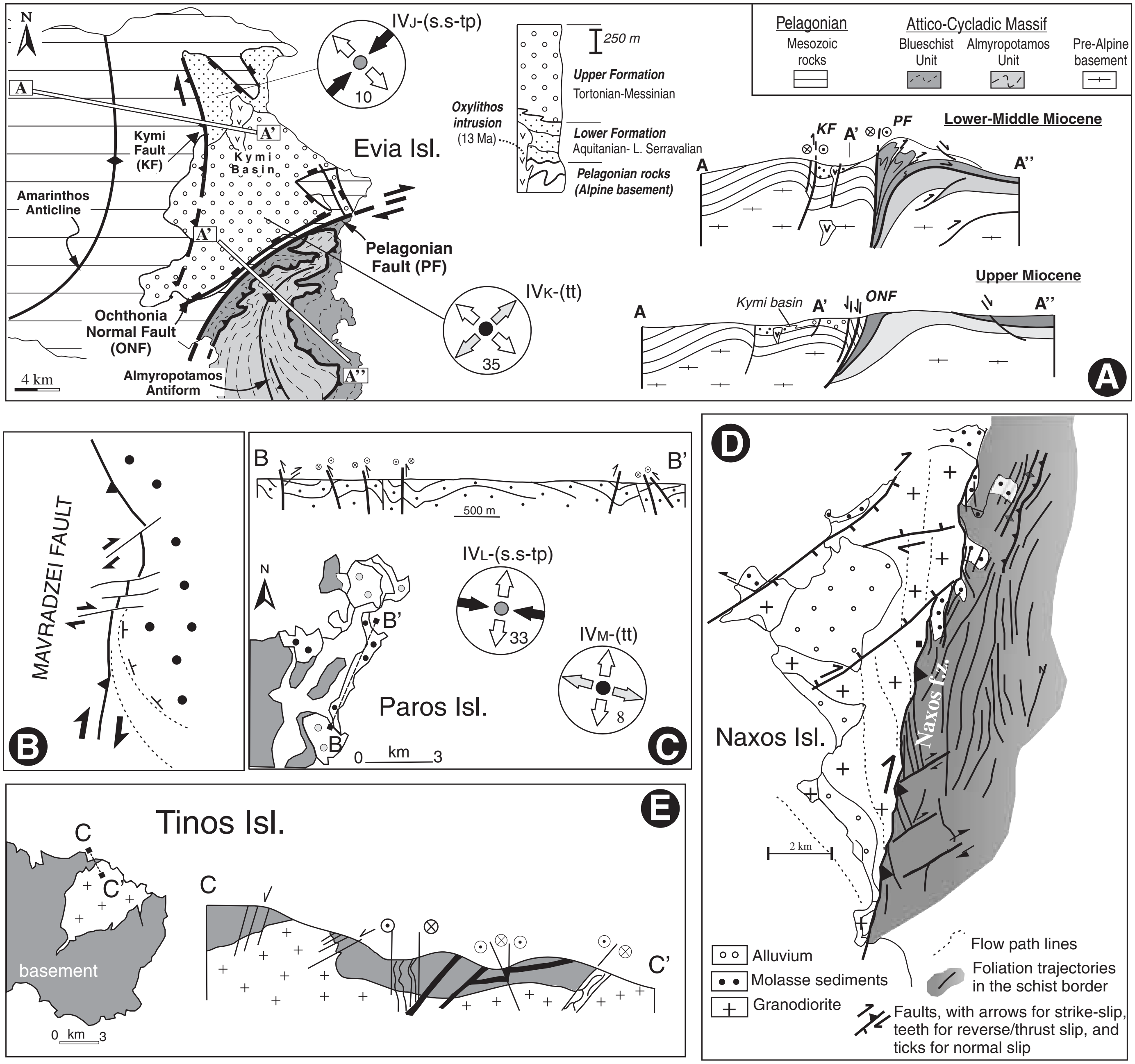

Figure 8. Key structural areas in Region IV (for locations see also Fig. 7) (A) Simplified structural map of central Evia island with stratigraphic column and stress nets from the Kymi basin. The orogenic evolution of the area is shown along a synthetic cross-section (after Xypolias et al., 2003). (B) Schematic sketch map of the transpressional Mavradzei fault, which borders the Miocene mollasic basin (Samos island). (C) Simplified map, cross-section, and stress nets (IV_L—older phase; IV_M—younger phase) in the western part of Paros island showing the deformation style in the molassic basin. (D) Structural map of the western area of Naxos island showing the granodiorite intrusion and flow lines deflection along the Naxos fault (after Koukouvelas and Kokkalas, 2003). (E) Geological map and cross-section in the eastern part of Tinos island showing the strike-slip deformation pattern related to the intrusion of the Miocene granitic body.

nites (Fig. 8E, section C-C′). Similar NE-trending faults also control the Ikaria (Fig. 7A, map) and Delos plutons (Boronkay and Doutsos, 1994; Pe-Piper et al., 2002).

Contractional movements in the Cycladic islands probably ceased during the late upper Miocene. For this period, the computed σ_1 stress axis varied between an ENE and an ESE orientation and caused dextral slip on the dominant NNE-trending fault system and sinistral slip on the northwest-trending system (Fig. 7A, stress nets IV_A, IV_B, and IV_I, and Fig. 8C, stress net IV_L).

Extensional Structures

After this contractional phase, a transtensional stress regime was established (Fig. 7A, i.e., the upper Miocene stress net Milos IV_C, and Fig. 8C, stress net Paros IV_M). Structures typical of this stage are also recognized on Evia Island (Fig. 8A, Kymi basin, stress net IV_K). During this period (the upper Miocene and after the Oxylithos intrusion at 13 Ma), the Kymi basin was strongly modified by an orthogonal system of ENE- and NNW-trending normal faults (Kokkalas, 2001), which was associated with the deposition of 800-m-thick upper Miocene sequence consisting of alluvial fan deltas and alluvial fan deposits (Fig. 8A, stratigraphic section). The synrift sediments thicken toward the southern margin of the basin, which is marked by an ENE-trending oblique normal fault, the Ochthonia normal fault (Fig. 8A, section A-A″, upper Miocene). Slip data taken along mesoscopic faults are consistent with a northwest-southeast-directed extension (Fig. 8A, stress net IV_K).

Similarly, on Samos sediments deposited from 7 Ma to the present are deformed by northeast- and northwest-trending oblique normal faults (Boronkay and Doutsos, 1994). Synsedimentary tectonism is evidenced by the rapid changes of the sedimentary facies and the presence of large wedge-shaped sedimentary prisms in the hangingwall of the faults. A non-orthogonal extensional system comprising northwest- to WNW- and northeast- to ENE-trending faults appears to have been established during this period accommodating the extensional deformation in the central Aegean Sea (Fig. 7A, stress net IV_D, and Fig. 7B, stress net IV_I; Doutsos and Kokkalas, 2001; Piper and Perissoratis, 2003). Analysis of slip data indicates that these faults were formed in the frame of a transtensional regime with an σ_3 axis trending around a northeast-southwest orientation (Fig. 7A, stress nets IV_E and IV_F). Small deviations from this direction are probably caused by variation in the shear magnitude acting along the ENE- and northeast-trending faults.

To summarize, we can say that syntranspressional molassic basins and magma intrusions along the ENE- or NNE-trending crustal scale faults marked the end effect of contraction and wrenching in the area around 10 Ma. Postorogenic collapse followed these pre-existing crustal discontinuities, which were established during the main orogeny. Uplift and thinning of the crust was mainly accomplished by footwall uplift along crustal-scale extensional faults.

STRUCTURAL REGION V: SOUTHEAST HELLENIC ARC

This region includes the eastern part of Crete and the Dodecanese islands (Fig. 9A). The crust in this part of the Hellenic arc is continental and attains a maximum thickness of 35 km below central Crete and then thins out to 24 km toward eastern Crete. Recent seismic tomography identified a NNE-ward dipping oceanic crust layer that is decoupled from the overlying continental crust at approximately central Crete (Bohnhoff et al., 2001).

Contractional Structures

The External Hellenides extend all the way across the southern Aegean area, although the correlation is not straightforward (Bonneau, 1984; Garfunkel, 2004; Kokkalas and Doutsos, 2004). The southeast Aegean region is an area where the overall orientation and geometry of the main tectonic structures reflect the strong control exerted by the pre-existing geometry of plate margins in the area. During the early Mesozoic, Apulia was separated from north Africa by WNW-trending rift zones and northeast-trending transfer faults (Robertson et al., 1991). These transfer faults probably caused a reorientation of the WNW-trending margins locally to an ENE orientation (Fig. 9, inset [i]). Oblique to pure contraction along the WNW- to ENE-trending margins resulted in the uplift and exhumation of H*P*-L*T* metamorphic rocks in Crete by a mechanism involving syncompressional uplift aided by vertical buoyancy forces (Kokkalas and Doutsos, 2004). Late orogenic basins were originated in association with a transpressional strike-slip stress regime consistent with the regional compression in a NNW-SSE direction, trending parallel to the nappe transport direction in the area (Fig. 9B, stress net V_L). Deformation accommodated along the pre-existing WNW-ESE- and NNE-SSW-trending faults formed during ductile transpression in the main orogenic phase. Typical examples can be given from eastern Crete and western Kos island. In eastern Crete, basement rocks were carried above middle–upper Miocene sediments (Fortuin, 1978; Frydas, 1988) along WNW-trending transpressional faults forming cataclasites up to 7 m thick, while the hangingwalls of the faults were internally deformed by mesoscopic reverse faults and a closely spaced cleavage (Fig. 9B, section D-D′; Kokkalas and Doutsos, 2004). A small-scale basin was downflexed by two WNW-trending faults that show a high oblique reverse character of motion (Fig. 9B, section C-C′). Oblique striae and offset of enechelon mesoscopic faults indicate the highly oblique reverse character of movements along the marginal faults. The stress field that was generated during transpressional deformation is characterized by a regional strike-slip or transpressional stress tensor with subhorizontal σ_1 trending in a NNW direction (Fig. 9B, stress net V_L).

On Kos Island, an ENE right lateral oblique reverse fault bounds basement rocks from middle–upper Miocene marine marls (Boger et al., 1974; Fig. 9B, section B-B′). In its footwall

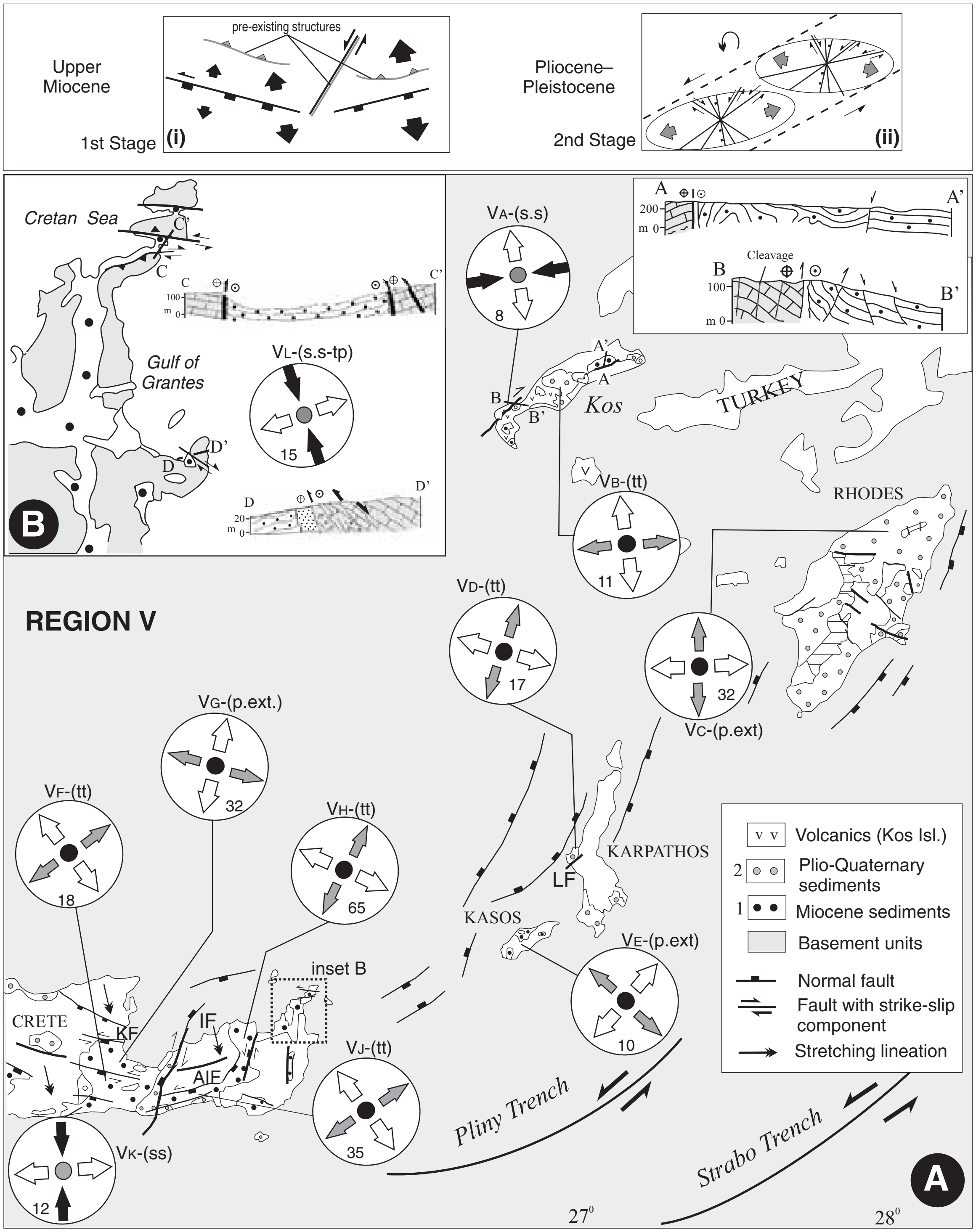

Figure 9. (A) Simplified map of the southeastern Aegean area (Region V). Stress net explanation as for Figure 3. The cross-sections (A-A′, B-B′) illustrate the style of deformation in the Miocene sediments of Kos island. The kinematic interpretation of the observed fault patterns and the resulting local stress fields during the forearc evolution are shown with insets for the two stages: (i) Upper Miocene and (ii) Plio-Pleistocene to present. (B) Simplified geological map and cross-section at the northeastern edge of Crete showing contractional structures related to the deposition of middle–upper Miocene deposits. Seafloor lineaments after Mascle and Martin (1990); Piper and Perissoratis (2003). AIF—Agios Ioannis fault; IF—Ierapetra fault; KF—Kritsa fault; LF—Lefko fault.

the fault produced a 20-m-wide zone internally deformed by oblique thrusts and mesoscopic folds (Kokkalas and Doutsos, 2001). To the east, an ENE-trending fault is also associated with small-scale flexures in the middle Miocene marls (Fig. 9B, section A-A′). The mesoscopic folds have a NE axis orientation, are open in shape, and attenuate toward the northeast. A strike-slip stress tensor is also determined from fault population analysis but with σ_1 trending ENE-WSW (Fig. 9A, stress net V_A).

Extensional Structures

During the upper Miocene, contractional movements diminished and a progressive transtensional stress regime was established.

As a first stage, extension was accommodated along ENE- and WNW-trending normal and oblique normal faults. These faults often delimited basement highs from the basin infill, such as the Agios Ioannis and Kritsa faults on Crete and the Lefko fault on Karpathos island (Fig. 9A, AIF, KF, and LF). The WNW-trending faults controlled mainly the late Miocene subsidence in the Ierapetra basin on Crete (Fig. 9A, IB). The regional tension (σ_3) during the first stage in the upper Miocene was oriented perpendicular to the main ENE fault trend, the main axis of the tectonic windows formed during the main orogenic phase, as well as to the direction of the Aegean-Africa plate margin (Fig. 9A), suggesting that a mechanism involving arc-normal pull and postorogenic collapse was responsible for the fore-arc evolution during this time (Fig. 9A, stress nets V_B, V_F, and V_J; see also Kokkalas and Doutsos, 2001). However, during this period the WNW- and NNE-trending faults that comprise inherited structures from the previous orogenic phase were reactivated and caused local stress deviations (Fig. 9A, stress nets V_E, V_G). The NNE tension represents a more local stress field caused by oblique slip along these faults.

At a later stage during the Plio-Pleistocene, a complex fault pattern consisting of NNE-SSW, NNW-SSE, and ENE-WSW oblique normal faults was active (Fig. 9A, inset [ii]). The regional stress field is characterized by WNW-ESE- to east-west-oriented tension perpendicular to the main NNE-trending faults and is compatible with transtensional deformation (Fig. 9A, stress nets V_C, V_D, V_H, and V_K). These faults are often associated with mesoscopic NNW-trending strike-slip faults, which do not follow any older fault direction. Important for the recent deformation of the region are the ENE-trending faults that display left lateral oblique normal and strike-slip movements, transecting upper Tortonian and younger sediments on Crete and Plio-Pleistocene sediments on Karpathos and Kos (Buttner and Kowalczyk, 1978; Kokkalas and Doutsos, 2001). The ENE orientation is trending subparallel to the active sinistral transform system occurring along the southeast Aegean plate boundary (Fig. 9A, Pliny and Strabo trenches; Le Pichon and Angelier, 1979).

This complex fault pattern sufficiently suits an incremental strain pattern of associated secondary strike-slip (P and Riedel shears), and also pull-apart extensional structures produced by left lateral shearing along large-scale ENE-trending structures, such as the Pliny and Strabo trenches, which appear to have significantly affected the deformation in the area during this period (Fig. 9A, inset [ii]). The pre-existing NNE-trending fault set, represented mainly by the Ierapetra fault in eastern Crete (Fig. 9A, IF), was reactivated in this period, as is displayed by the recent seismic activity and the Quaternary fan breccias juxtaposed against the fault scarp, and has controlled local stress field evolution (Kokkalas and Doutsos, 2004).

DISCUSSION

The deformation of the Aegean region and its surroundings is a result of the interaction of the Africa and Aegean plates and the key role of the Anatolia plate. The latter has had direct implications for the North Aegean trough, at least since 5 Ma (Armijo et al., 1999). However, this work focuses on the post-collisional deformation along the External Hellenides. The External Hellenides orogenic belt formed an orocline with variability in the crustal thickness of the overriding plate along strike and more advanced stages in the plate convergence process along the Hellenic arc. These different structural styles along strike have been enhanced by the previously established crustal anisotropy of each region (Fig. 10 and insets).

In Region I, pure continental collision is occurring in a belt extending from the Ionian islands to the Ioannina area. An array of northwest-striking normal faults that trend parallel to the Hellenides are associated with the postorogenic collapse of the thickened crust during the upper Miocene (Fig. 10, inset Region I). Strong coupling and resistance of the colliding plates may have allowed the transmission of the horizontal compressive forces from the collision front to the overriding Aegean plate. As a result, northeast-trending grabens were formed parallel to the plate motion vector of the mechanically strong Adria plate in response to indentation tectonics similar to those of the Rhine graben (Laubscher, 2001; Hinzen, 2003).

Farther to the south, in Region II, a mixed type of contractional-extensional deformation is evidenced. This type of deformation is associated with the subduction of intermediate crust with variable slab dip angles below the overriding plate (Hirn et al., 1996; Laigle et al., 2002), similar to the geometry of the Andes plate margin (Ramos et al., 2002). Contraction affects the outer arc area (the Ionian islands), while in more distal areas above the flat and steep part of the subducted slab, extensional deformation dominates along both the ENE- and the WNW-trending faults (Fig. 10, inset Region II). These fault orientations appear to be reactivated Mesozoic transverse structures that controlled the width of the passive margin and the Mesozoic synrift sedimentation (Fig. 10, i.e., the Corinth rift transfer zone and the Sperchios transfer; see also Skourlis and Doutsos, 2003). The ENE-trending faults have been regarded as the result of the postorogenic uplift and extension behind the Hellenic trench after the middle Miocene (Doutsos et al., 1988). The whole area can

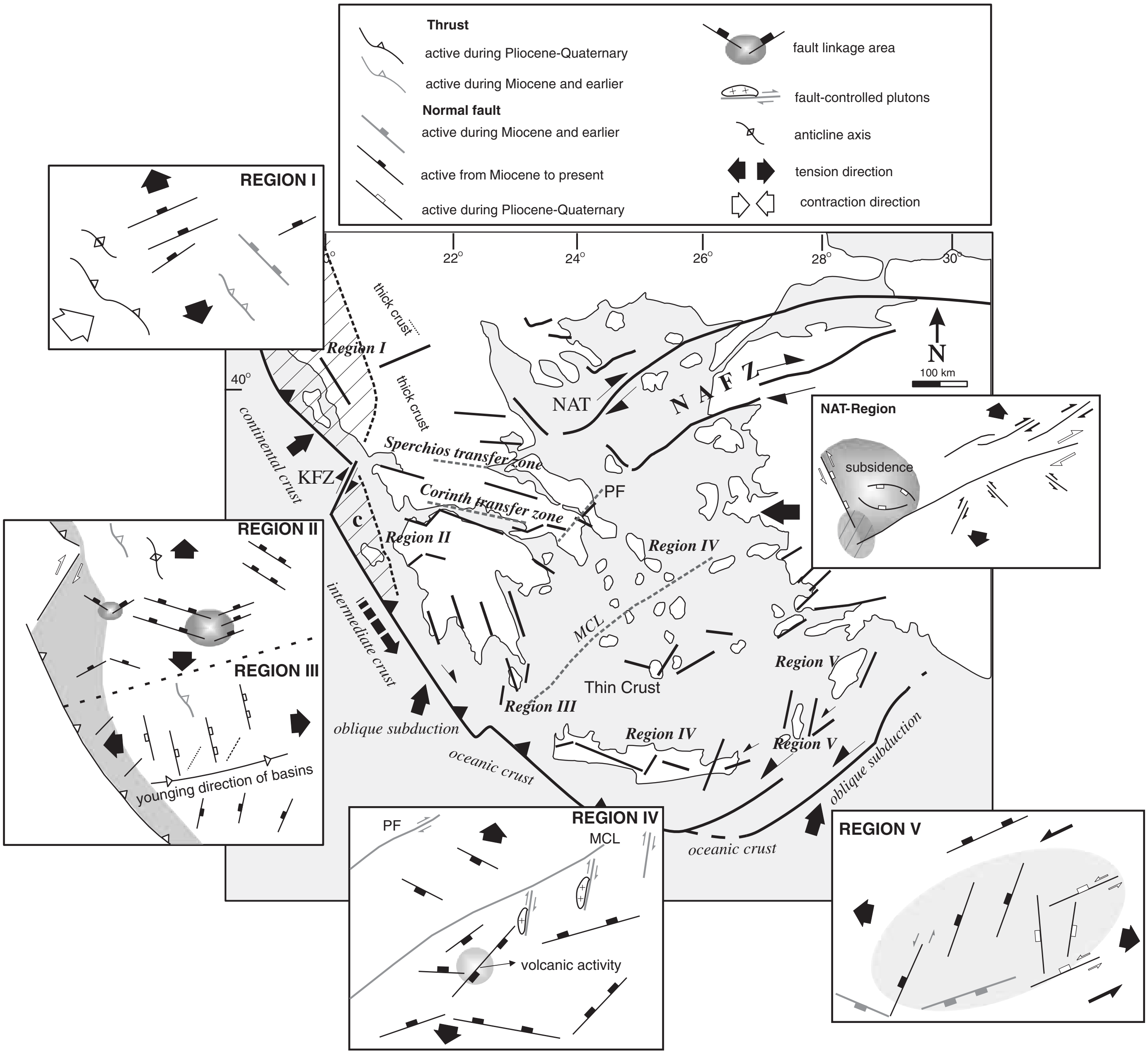

Figure 10. General simplified structural map of Greece showing the main currently active structures in the five structural regions along the Hellenic Arc, as well as some main pre-existing lineaments. Insets illustrate the main structural features of each region and the period of activity of these structures (for further details see discussion). KFZ—Kefallonia Fault zone; MCL—Mid-Cycladic lineament; NAFZ—North Anatolia fault zone; NAT—North Aegean trough; PF—Pelagonian fault.

be interpreted as one of diffuse deformation affected by both the propagating North Anatolia fault and the slab retreat process (Dewey and Şengör, 1979; Reuther et al., 1993; Doutsos and Kokkalas, 2001).

Regions III and V display similar deformation styles but with opposite shear senses, a less pronounced dextral shear sense in the west and a more prominent sinistral shear sense in the east, respectively (Fig. 10, inset Regions III and V). In both areas, oceanic crust is subducting. The NNW- to northwest-trending faults in the west and the WNW-trending faults in the east are aligned parallel to the structural grain of the External Hellenides and are also associated with postorogenic collapse during the upper Miocene. The NNE-trending pre-existing faults cause local deviations and redistribution of stress. During the Pliocene–Quaternary, a complex fault pattern of reactivated and newly formed NNW- to NW-, NNE-, and ENE-striking faults characterized the deformation of the overriding plate in the course of the oblique subduction process of the Africa plate. The time period for the increased activity of the plate slab retreat mechanism in Region III is constrained by the fact that sedimentary basins controlled by the NNW-trending faults become younger toward the east. Sediment accumulation within the basins started during the upper Pliocene in the southwest Peloponnesus and extended to the upper Quaternary in the Argolic Gulf (Zelilidis and Doutsos, 1992; Piper and Perissoratis, 2003). In the southeastern part of the arc (Region V), the ENE-striking faults that trend parallel to the Pliny and Strabo troughs display sinistral shearing, emphasizing in this way the high obliquity of convergence in the area (Fig. 10, inset Region V). Along the concave part of the arc (Regions III and V), intense east-west extension (Lyon-Caen et al., 1988; Kokkalas and Doutsos, 2001; Piper and Perissoratis, 2003) and radial GPS vectors are evidenced (McClusky et al., 2000). The process of the partitioning of oblique convergence in response to simultaneous drag from the downgoing plate can explain these characteristics as well as the slower convergence rates normal to the Hellenic arc (McCaffrey and Nabelek, 1998). Localized trench-parallel asthenospheric flow from rapid slab retreat (Wortel and Spakman, 2000) can also explain this extension (see also Russo and Silver, 1994).

Finally, the central part of the Hellenic arc (Region IV) is characterized by a thermally weakened crust and pre-existing NE- to ENE-trending crustal scale lineaments (Fig. 10, inset region IV, i.e., the Pelagonian fault and Mid-Cycladic lineament), while similar ENE-trending "weak" zones are also cutting the eastern Aegean area and the Anatolia plate (Dilek et al., 1999). The ENE-trending zones appear to play a significant role and to control the present deformation of the area (e.g., the 1999 Athens earthquake; Papadopoulos, 2002) and have made a great contribution to the present-day bathymetry of the North Aegean trough (Fig. 10, inset North Aegean trough). The effect of the North Anatolia fault in the north Aegean area may have started after 5 Ma (Armijo et al., 1999; Koukouvelas and Aydin, 2002; Flerit et al., 2004). In off-shore basins of the Aegean Sea, WNW- and ENE-trending faults accommodate the present north-south-directed extensional deformation (Doutsos and Kokkalas, 2001; Piper and Perissoratis, 2003), while their linkage area seems to concentrate both seismic activity (e.g., the 1956 Amorgos earthquake, Ms = 7.4; Papadopoulos and Pavlides, 1992) and volcanism (e.g., the Santorini volcanic center; Piper et al., this volume). The thermally weakened lower crust may have resulted in high degrees of vertical decoupling between the upper and the lower crust, facilitating the rapid advancement of the Aegean plate toward the southwest, as is indicated by GPS motions along the ENE- and northeast-trending pre-existing crustal scale faults.

The fault pattern and stress field in the overriding Aegean plate are variable along the Hellenic arc (Fig. 10), depending mainly on crustal thickness and mantle rheology, allowing a solely crustal or mantle signature to dominate in certain areas. Thus, the fault pattern and the associated stress field in the overriding plate may not reflect the state of strain of the entire plate but may be partially decoupled from the driving forces of the broader scale plate and reflect the inherited anisotropy of the crust.

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