EXTENSIONAL COLLAPSE OF THE HELLENIDES: A REVIEW

A. A. Kilias[1], M. D. Tranos[1], M. Orozco[2], F. M. Alonso-Chaves[3] y J.I. Soto[2]

[1] *Department of Geology and Palaeontology, University of Thessaloniki, GR-54006 Thessaloniki, Greece.*
[2] *Instituto Andaluz de Ciencias de la Tierra and Departamento de Geodinámica, C.S.I.C.-Universidad de Granada, Campus Fuentenueva s/n, 18002 Granada, Spain.*
[3] *Departamento de Geodinámica y Paleontología, Universidad de Huelva Campus El Carmen, 21071 Huelva, Spain.*

Abstract: In this paper we summarise the geometry and kinematics of late-orogenic extension and orogen collapse in the Hellenic arc-shaped orogen during the Tertiary. The Hellenic orogen shows both symmetric and asymmetric late-orogenic extension giving rise to the exhumation of deep-seated crustal rocks. Double-vergent extension took place in the cold accretionary prism (Olympos, Ossa, Pelion, and Crete) without significant crustal heating, whereas asymmetric extensional collapse occurred in the back-arc areas (Rhodope and Cyclades), associated with significant crustal heating. Extension started in the internal Hellenides during the Eocene and extended progressively to the external Hellenides during the Miocene. Tertiary late-orogenic extension of the Hellenides occurred concomitantly behind a south-to-southwest migrating compressional front, finally forming an arched orogenic system. Late-orogenic extension in the Hellenides could have been driven by changes in the rate and direction of convergence between the African and Eurasian plates during the Tertiary.

Key words: Hellenide Orogen, late-orogenic extension, Tertiary exhumation, accretionary prism, Mediterranean back-arc basins.

Resumen: El Orógeno Helénide como parte del cinturón alpino mediterráneo, está estrechamente relacionado con la convergencia entre las placas africana y euroasiática durante el Mesozoico y el Cenozoico. En este contexto, durante el Terciario temprano a medio se produjo la colisión de bloques menores de estos continentes, como las placas Apulia y Pelagonia, dando lugar a la formación de la mayor parte del mencionado Orógeno Helénide. Este estadio se caracteriza por la formación de "pilas de mantos" producidas por cabalgamientos dirigidos fundamentalmente hacia el suroeste (respecto a la orientación actual) y el consiguiente engrosamiento de la corteza continental. Esta tectónica terciaria está asociada con la formación de dos cinturones metamórficos principales de alta P / baja T, alineados a lo largo de dos arcos concéntricos, uno interno y otro externo, que reflejan la evolución de procesos de subducción sucesivos. El cinturón interno se desarrolló durante el Paleoceno-Eoceno y está caracterizado por la zona de sutura entre Helénides Internos (placa superior) y Helénides Externos (placa inferior) que aflora en los montes Olimpo, Ossa y Pelión, en la isla de Evia y en las islas Cíclades. El cinturón externo de alta P / baja T se formó en el Oligoceno-Mioceno y está caracterizado por una zona de "collage" situada en el interior de los Helénides Externos que aflora en el Peloponeso meridional y en Creta. La tectónica extensional ha jugado un papel muy importante en la configuración final del Orógeno Helénide. En este artículo se investiga la geometría y la cinemática de la extensión tardi-orogénica y del colapso orogénico en los Helénides durante el Terciario. El estudio del Orógeno refleja la existencia de extensión tardi-orogénica, tanto de tipo simétrico como asimétrico, que dio lugar al colapso del Orógeno y al levantamiento y exhumación de rocas de niveles corticales más profundos. Por encima del prisma de acreción frío (Olimpo, Ossa, Pelión, Creta) se produce extensión divergente (simétrica) sin calentamiento cortical significativo. En el área trasera del arco (Rodopo, Cíclades) tiene lugar colapso extensional asimétrico asociado a un significativo calentamiento cortical. La extensión tardi-orogénica terciaria en los Helénides junto con la compresión en el frente de deformación de la placa litosférica estirada, constituyen un sistema orogénico dinámico que migró hacia el suroeste. La extensión comenzó en el Eoceno en los Helénides Internos y alcanzó progresivamente los Helénides Externos durante el Mioceno. La geometría y la cinemática de la extensión tardi-orogénica en los Helénides podrían ser atribuidas a cambios en la velocidad de convergencia entre las placas africana y euroasiática durante el Terciario.

Palabras clave: Orógeno Helénide, extensión tardi-orogénica, exhumación terciaria, prisma de acreción, cuencas retro-arco Mediterráneas.

Kilias, A.A., Tranos, M.D., Orozco, M., Alonso-Chaves, F.M. y Soto, J.I. (2002): Extensional collapse of the Hellenides: A review. *Rev. Soc. Geol. España,* 15 (3-4): 129-139.

During the last two decades, late-orogenic extension, that is, extension following orogenic stacking and lithosphere thickening, has been intensively investigated by many geoscientists, thus lending a new understanding to orogenic processes and the incidence of plate-convergence boundary conditions (*e.g.*, Wernicke 1981, 1985; Platt, 1986; Dewey,1988; Ruppel, 1995). The most influential result of this process is that horizontal extension through large-scale, low-angle extensional shear zones and normal faults leads to the uplift and exhumation of crustal rocks from deeper tectonic levels.

Taking into account the geometry and kinematics of these extensional shear zones, two end-member modes of exhumation of deep crustal rocks have been proposed (*e.g.*, McKenzie, 1978; Wernicke, 1985; Lister *et al.*, 1986; Hamilton, 1987; Davis and Lister, 1988; Malavieille, 1993): (1) The asymmetric type, which is characterised by simple-shear or non-coaxial deformation, and (2) the symmetric type, which is related to coaxial or pure-shear deformation. The first type is characterised by the development of a master low-angle extensional system, with listric normal faults in the upper crust, coalescing within a deeper ductile shear zone. Extension along the basal detachment fault drives upward doming and uplift of the underlying crust. The second type is characterised by the development of two conjugate sets of low-angle extensional shear zones with opposite sense-of-shear on the two sides of the exhumed dome.

This paper summarises the geometry and kinematics of the late-orogenic extension evidenced by Tertiary, large-scale, low-angle detachment systems recognised in different parts of the Hellenic orogen. This summary provides a thorough overview of the evolution of this extension during the Tertiary and a better understanding of the mechanisms for the Hellenic orogen collapse during the Alpine orogeny.

The nappe pile of the Hellenic orogen

The Hellenic orogen (Fig. 1), as a part of the Alpine orogen in Eurasia, is closely related to the convergence between the African and Eurasian plates during the Mesozoic and Tertiary and the closure of the Tethyan ocean basin or basins (Aubouin *et al.*, 1963; Godfriaux, 1968; Jacobshagen *et al.*, 1978; Seidel *et al.*, 1982; Robertson and Dixon, 1984; Vergely 1984; Mountrakis, 1986; Schermer *et al.*, 1989; Doutsos *et al.*, 1993; Dinter, 1998).

A complex nappe-pile structure dominates the Hellenides. In the Hellenic mainland and the island of Crete, the structural sequence comprises (Figs. 1 and 2), from top to bottom:

(1) The Mesozoic ophiolite unit (oph) of the Axios (Ax) and Subpelagonian (Sp) zones overlain by transgressive Cretaceous limestones terminating in an upper Cretaceous-Palaeocene flysch.

(2) The Pelagonian nappe (Pl), consisting of a pre-Alpine Pelagonian continental basement, is formed of gneisses and schists intruded by upper Palaeozoic granitoids. These nappes are overlain by an upper Palaeozoic-Triassic, volcano-sedimentary sequence and Triassic-Jurassic, platform carbonate rocks.

(3) The internal blueschist unit with high P-low T blueschist facies metamorphism of uncertain age.

(4) The weakly- to non-metamorphosed sedimentary sequences of the External Hellenides constitute the bottom of the Hellenic nappe-pile edifice. The External Hellenides are composed of tectonic sheets with neritic and pelagic limestones and flysch deposits of Triassic to Miocene age, and were deposited along the Apulian continental margin. The Parnasos (Pa), Pindos (Pi), Gavrovo-Tripoli (G), Ionian (I) and Paxos (Px) zones belong to the External Hellenides.

(5) The external blueschist unit of the External Hellenides.

The Serbomacedonian (Sm) and Rhodope (Rh) massifs in the NE Hellenides have been considered as the relative hinterland of the Hellenic orogenic belt. They comprise a structurally complex domain of tectonically intercalated high- and low-grade pre- and syn-Alpine metamorphosed rocks, intruded by syn- and post-kinematic Mesozoic and Tertiary igneous rocks/ bodies. Recent studies have distinguished in the Rhodope massif a Tertiary metamorphic core complex (Dinter and Royden, 1993; Dinter, 1998; Kilias *et al.*, 1999). The Serbomacedonian massif, together with the Circum Rhodope Belt (CRB), with a volcano-sedimentary series of Triassic-Jurassic age, overthrust the Axios zone towards the SW (Tranos *et al.*, 1999). The Serbomacedonian and Rhodope massifs, the Circum Rhodope Belt, the internal blueschist unit and the Pelagonian nappe constitute the main part of the Internal Hellenides. Furthermore, the Attico-Cycladic massif (AC) in the Cyclades and Evia islands and the Attica peninsula is mainly composed of high-pressure rocks of the internal HP belt, overlying a carbonate metamorphic sequence and underlying thinned Hercynian or older crystalline basement rocks. Remnants of ophiolitic rocks overlain by transgressive Cretaceous limestones occupy the tectonic top of the Cycladic structural sequence. Thus, in the Cyclades a nappe-stack order is exposed that is similar to that described for the Hellenic mainland. Some parts of the Attico-Cycladic massif have been interpreted as Tertiary metamorphic core complexes (Lister *et al.*, 1984; Gautier and Brun, 1994).

Nappe stacking started in the middle-upper Jurassic with the obduction of the ophiolitic unit onto the Pelagonian carbonate cover from one or more Tethyan ocean basins (Jacobshagen *et al.*, 1978; Vergely, 1984; Mountrakis, 1986). At the same time, syn-metamorphic deformation and basement-involved thrusting in greenschist to amphibolite facies conditions affected the internal parts of Hellenides (Yarwood and Dixon, 1977; Jacobshagen *et al.*, 1978; Vergely, 1984). Deformation progressed episodically throughout the

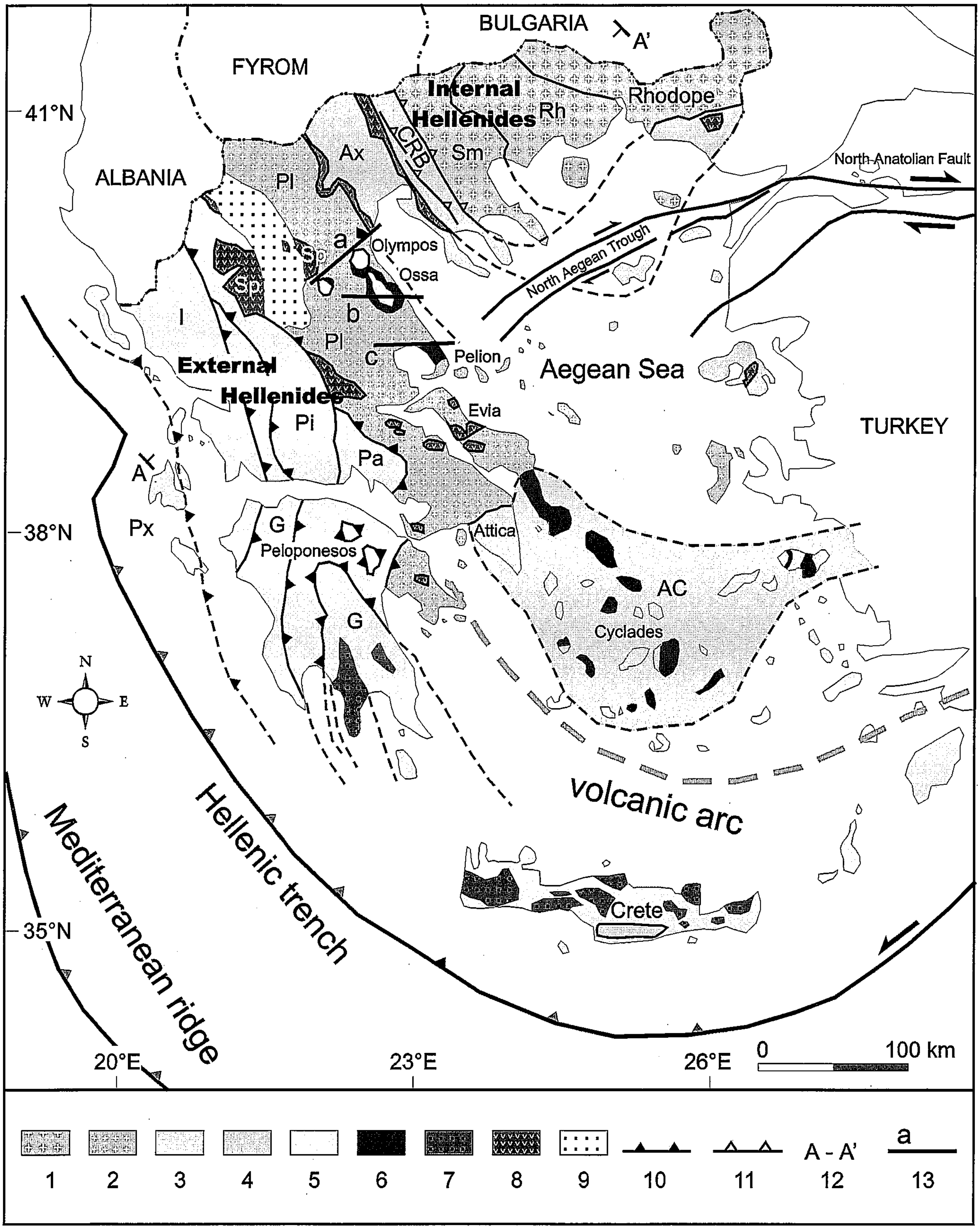

Figure 1.- Geological map showing the main structural domains of the Hellenides (modified from Aubouin *et al.*, 1963 and Mountrakis, 1986). 1. Rhodope (Rh) and Serbomacedonian (Sm) massifs, 2. Pelagonian nappe (Pl), 3. Attico-Cycladic massif (AC), 4. Axios zone (Ax) and Circum Rhodope Belt (CRB), 5. External Hellenides (Pa: Parnassos zone, Pi: Pindos zone, G: Gavrovo-Tripoli zone, I: Ionian zone, Px: Paxos zone), 6. Internal HP metamorphic belt, 7. External HP metamorphic belt, 8. Ophiolitic rocks, SP: Subpelagonian zone, 9. Mesohellenic Mollase Trough, 10. Eocene to Miocene thrusts, 11. Palaeocene-Eocene thrusts,12. A-A' cross-section of Fig. 2, 13. Cross-sections shown in Fig. 5.

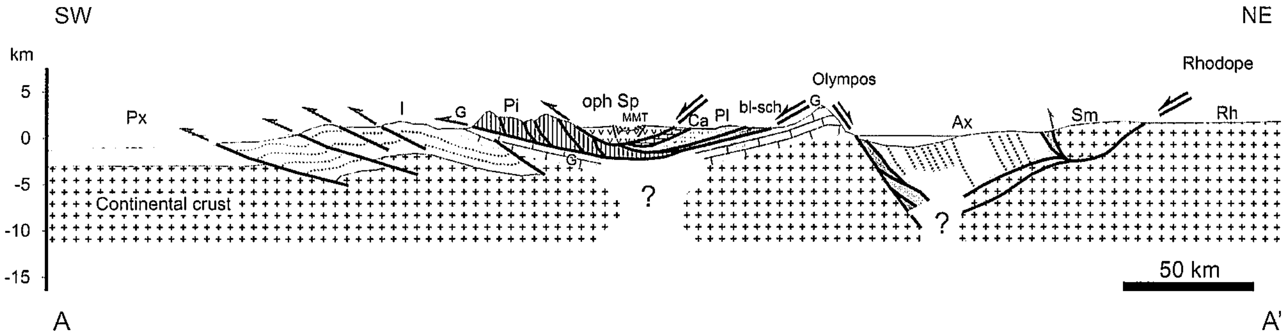

Figure 2.- Simplified geological cross-section through the Hellenides (modified from Nance, 1981). (Abbreviations as in Fig. 1; bl-sch: blueschists, Ca: Pelagonian carbonate cover, MMT: Mesohellenic Molasse Trough, grey arrows: thrusts, black arrows: extensional faults.

Cretaceous and Tertiary. During the late Eocene, the Pelagonian nappe and the internal blueschist unit were transported SW-wards onto the External Hellenides, thus ending the last stage of the Tertiary collision (Godfriaux, 1968; Schermer *et al.* 1989; Schermer 1990; Kilias *et al.* 1991). However, in contrast, some researchers suggest a NE-ward emplacement of the Pelagonian nappe (Barton, 1976; Doutsos *et al.*, 1993; Lips *et al.*, 1998).

The initial Tertiary tectonics were characterised by extensive nappe stacking, mostly directed towards the SW (with respect to the present-day orientation), and continental crustal thickening. Nappe stacking occurred throughout the Tertiary, with a migration of the deformation and compressional front towards the SW (Aubouin *et al.*, 1963; Schermer *et al.*, 1989; Robertson *et al.*, 1996). Nowadays, active compression is recognised along the Hellenic subduction trench, south of Crete, in relation with the subduction of the African plate beneath the Hellenides (Figs. 1 and 3) (Le Pichon and Angelier, 1979; Spakman *et al.*, 1988; Meulenkamp *et al.*, 1988).

Tertiary collisional tectonics and crustal thickening is associated with the formation of the two main high-pressure (HP) metamorphic belts shaping two concentric arcs, the internal and the external one, reflecting the evolution of successive subduction processes (Figs. 3 and 4a):

(1) The internal HP belt, which consists of meta-sediments and meta-volcanic rocks deriving from a continental margin, reached peak pressures during the Palaeocene-Eocene (Godriaux, 1968; Wijbrans and McDougal, 1988; Schermer *et al.*, 1989; Schermer, 1990; Lips *et al.*, 1998). This HP belt represents the suture zone between the Pelagonian nappe (upper plate) and the External Hellenides (lower plate); this suture zone is exposed in the Olympos, Ossa, and Pelion mountains as well as in the Evia and Cyclades islands (Godfriaux, 1968; Schliestetd *et al.*, 1987; Schermer, 1990; Kilias *et al.*, 1991). HP metamorphism reached blueschist and eclogite facies. Estimated peak PT conditions for this event are *ca.* 15 kbar and 450°-500°C in the Cyclades (Fig. 4b) (Altherr *et al.*, 1972; Schliestedt *et al.*, 1987). However, in the Olympos, Ossa, Pelion and Evia areas, P-T conditions are *ca.* 5-8 kbar and 300-350°C (Fig. 4c) (Schermer *et al.*, 1989; Kilias *et al.* 1991).

(2) The external HP belt lies within the External Hellenides and reached peak PT conditions during the Oligocene to Miocene (Seidel *et al.*, 1982). It crops out in southern Peloponesos and Crete; the Phyllite-Quartzite and Plattenkalk units comprise the external HP belt in the latter area (Seidel *et al.*, 1982). Maximum P-T conditions of *ca.* 10 kbar and 450°C have been estimated for these rocks (Fig. 4d) (Thomson *et al.*, 1998).

Another HP metamorphic event has recently been reported in the internal-most part of the Hellenic orogen, in the Rhodope crystalline massif (Liati, 1986). However, the age of this metamorphism is still under discussion, some authors considering this early subduction event to have taken place during the Cretaceous (Wawrzenitz and Mposkos, 1997), while others date it as Eocene (Liati and Gebauer, 1999).

Tertiary late-orogenic extension

Extensional tectonics played a significant role in the Tertiary-to-Recent tectonic evolution and the final configuration of the Hellenic orogen. Late-orogenic extension occurred both in the internal Hellenides and in the two HP belts, causing orogenic collapse and crustal thinning. Extensional exhumation of the deeper tectonic units occurred at the same time, thus giving rise to a series of tectonic windows and metamorphic core complexes (Lister *et al.*, 1984; Schermer *et al.*, 1989; Kilias *et al.*, 1991, 1994, 1995; Dinter and Royden, 1993).

The extensional shear zones and extensional detachments which accommodated the collapse of the Hellenic orogen have a diverse geometry and kinematics, as detailed below (Figs. 3, 4a and 5):

(1) In the crystalline Rhodope massif of the internal Hellenides (Kilias and Mountrakis, 1990; Dinter and Royden, 1993) and in the Cyclades area (Gautier and Brun, 1994), the collapse took place along low-angle extensional systems and associated extensional shear zones. The main sense of shear of these systems is top to the SW in the Rhodope massif, and top to the NNE in the Cyclades area (Figs. 6c and 6d). The upwarping of the normal shear zone surfaces caused by tectonic denudation can lead to the exhumation of lower plate rocks which usually crop out in the core of domal structures.

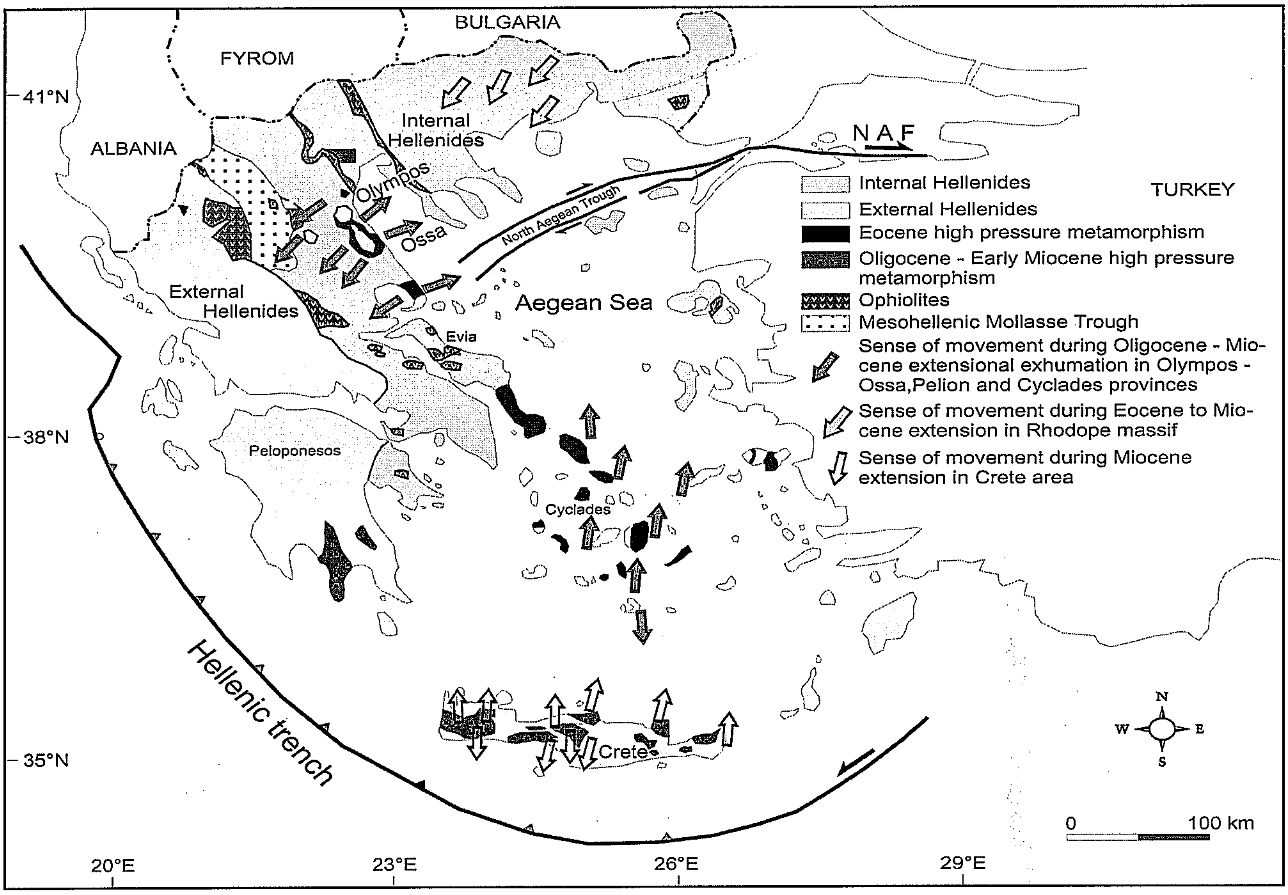

Figure 3.- The two HP suture belts and the kinematics of the Tertiary late-orogenic extension of the Hellenides. NAF: North Anatolian fault.

(2) In the Olympos, Ossa, and Pelion areas (Fig. 5) (Kilias, 1991; Kilias *et al.*, 1991, 1995), as well as on the island of Crete (Kilias *et al.*, 1994; Fassoulas *et al.*, 1994), a double-vergent extensional regime has been documented. In this area orogenic collapse took place throughout two different systems of low-angle extensional shear zones, showing opposite sense-of-shear in the flanks of the lower-plate dome (Figs. 6a and 6b). These antithetic extensional shear zones developed either coevally, suggesting bulk coaxial deformation, or almost coevally in time, the result being a double-vergent extensional regime and symmetrical domes. The dominant sense of displacement of these extensional shear zones is both top-to-the-SW and to-the-NE in the Olympos, Ossa, and Pelion mountains, and both top-to-the-N and to-the-S in Crete.

In the areas of Rhodope and Cyclades, extensional exhumation occurred associated with significant heating under low P conditions, leading to greenschist and amphibolite facies metamorphism (HT metamorphism) and granite intrusions (Liati, 1986; Wijbrans and McDougall, 1988), so that the HP mineral assemblages have been preserved only as relics (Fig. 4b; Liati, 1986; Altherr *et al.*,1982). Granite intrusions followed the HT event in these areas (Liati, 1986; Schliestedt *et al.*, 1987; Dinter, 1998; Kilias and Mountrakis, 1998). In contrast, in the Olympos, Ossa, Pelion, and Crete regions, extensional exhumation took place under isothermal decompression conditions in a relatively cold environment (Figs. 4b, c). As a result, and in association with the rapid orogenic exhumation, the initial HP mineral assemblages have been preserved in these regions (Schermer *et al.*, 1989; Kilias *et al.*, 1991, 1995; Thomson *et al.*, 1998).

Extensional collapse in the Rhodope and Cyclades areas took place behind the orogenic arc in the back-arc area, whereas in Olympos, Ossa, Pelion, and Crete, extensional collapse took place immediately above the cold accretionary prism of the Hellenic orogen (Fig. 4a) (Kilias, 1991; Jolivet *et al.*, 1994; Kilias *et al.*, 1994; Fassoulas *et al.*, 1994).

Footwall rocks from the crystalline Rhodope massif have progressively cooled from the Eocene/Oligocene up to the Miocene. Cooling to near-surface temperatures becomes progressively younger towards the SW, that is, parallel to the sense of displacement of the detachment. (Liati, 1986; Dinter and Royden, 1993; Wawrzenitz, 1997; Dinter, 1998). Kilias *et al.* (1999) related the gradual uplift of the Rhodope massif to extensional collapse of the orogen. Moreover, in Cyclades the extension began during the Oligocene-Miocene, as indicated by the age of the syn-extensional metamorphism (Altherr *et al.*, 1982; Lister *et al.*, 1984). In Olympos, Ossa, and Pelion, extensional collapse started during the Oligocene-Miocene (Schermer *et al.*, 1989; Sfeikos *et al.*, 1991; Kilias *et*

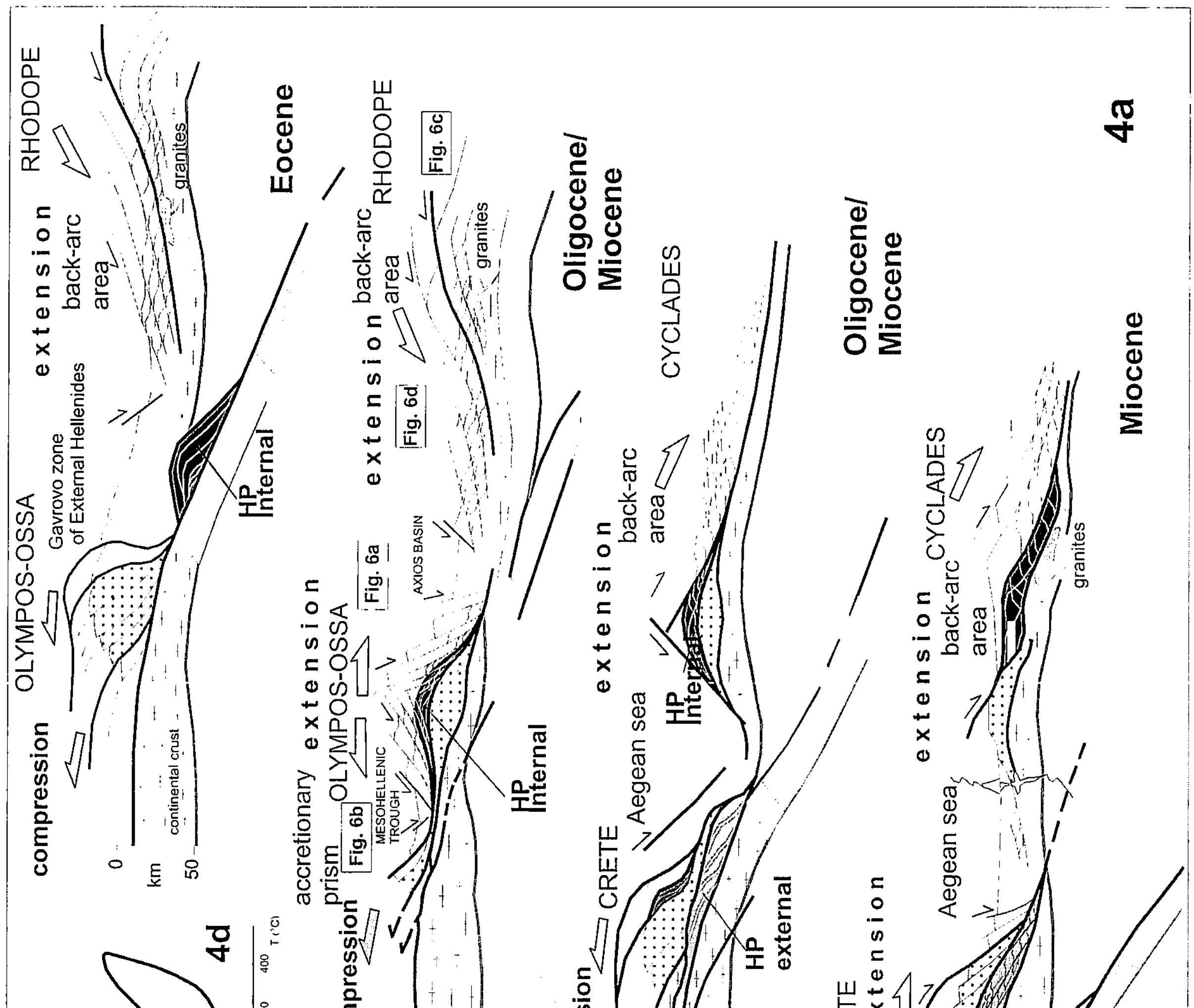

Figure 4.- (a) Schematic cross-sections illustrating the geometry and kinematics of the late-orogenic extension and associated compression in the Hellenides and for different ages, (b), (c), (d) indicate the representative P-T paths of the HP metamorphic belts in the Cyclades (taken from Wijbrans and McDougall, 1988), Olympos-Ossa (Schermer, 1990) and Crete (Thomson *et al.*, 1998) areas respectively, (e) the SSW-wards propagation of the subduction processes and the associated compression in the Hellenides.

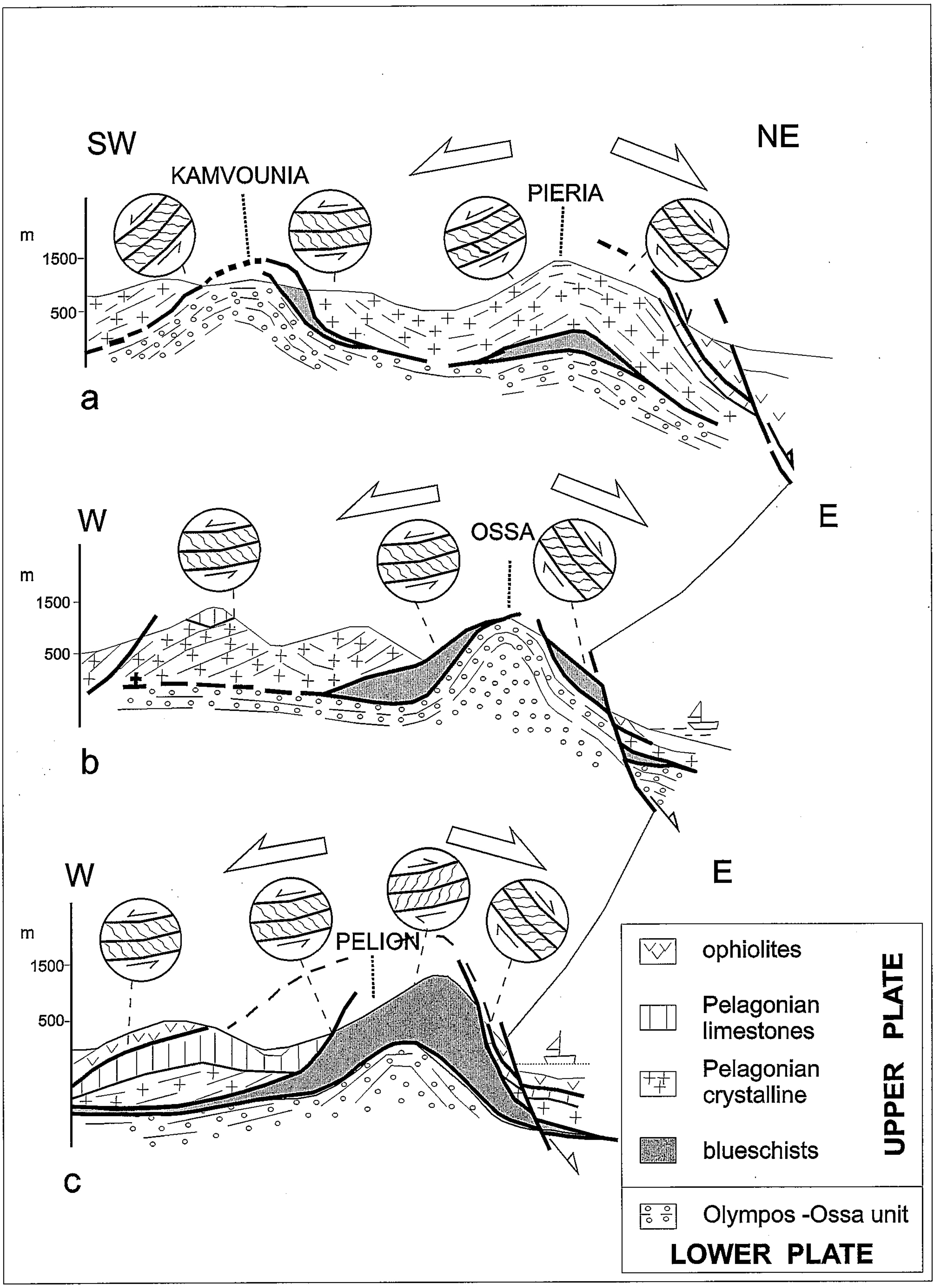

Figure 5.- Geological cross-sections illustrating the double-vergent sense of movement of the orogenic collapse of the Hellenides at the Olympos, Ossa and Pelion areas (from Kilias, 2001).

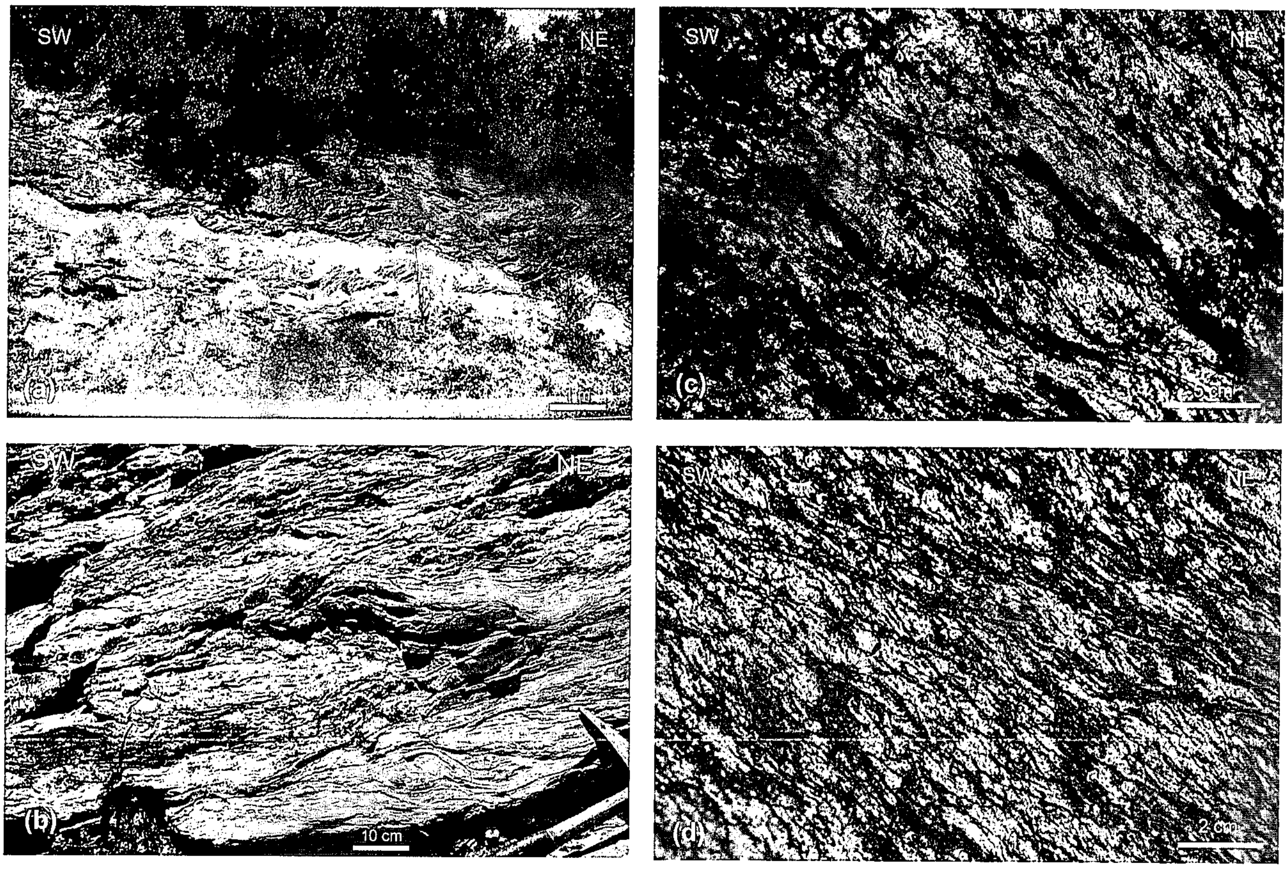

Figure 6.- Extensional shear zones in the Hellenic orogen. (a, b) Antithetic top to the NE and SW sense of movement in the Olympos area, (c, d) Top-to-the-SW sense of movement in the Rhodope province. Their location is shown in Fig. 4a.

al., 1995), whereas in Crete it initiated during the middle Miocene (Kilias *et al.*, 1994; Fassoulas *et al.*, 1994; Thomson *et al.*, 1998).

During the early stages of the Tertiary late-orogenic extension, pervasive shallow-dipping main schistosity (S_2) associated with stretching lineation (L_2) formed in the stretched deeper crustal parts and the HP metamorphic belts. Simultaneously with the D_2 ductile fabric, normal faults with the same kinematic symmetry formed in the higher crustal levels, leading to the development of extended basins. According to our above description, both S_2 fabric and L_2 stretching lineation indicate a different age and P-T metamorphic conditions as well as orientation at the several deformed parts of Hellenides.

The S_2 schistosity partly or totally overprints a previous fabric that formed during the HP deformational stage (S_1 schistosity). Usually, due to intense transposition of the pre-existing schistosity along the S_2 planes, both the S_1 and S_2 schistosities trace almost parallel, whereas in other cases very distinct crenulation occurs. Occasionally, isoclinal folded elongate quartz lenses and/or white micas are included as intrafolial rootless fold within the S_2 microlithons.

In the Rhodope and Cyclades back-arc areas, ductile D_2 shear zones indicate a thickness of at least 1-2 km. They occurred simultaneously with or outlasted the thermal peak of the HT metamorphism and continued to be active as the temperature decreased (Lister *et al.*, 1984; Schliestedt *et al.*, 1987; Kilias *et al.*, 1999). In the Olympos - Ossa accretionary prism area, ductile extensional shear zones reveal a thickness of some hundred metres associated with intense mylonitization and very low-grade retrogression of the HP mineral assemblages (Schermer *et al.*, 1989; Kilias *et al.*, 1991; Sfeikos *et al.*, 1991). Finally, in the Crete accretionary prism area, extensional deformation is localised at its early stages in ductile thinner shear zones also associated with very low-grade retrogression of the HP mineral assemblages (Fassoulas *et al.*, 1995; Thomson *et al.*, 1998).

Discrete brittle-ductile to brittle low-angle extensional fault zones (S_3) have progressively formed under lower P-T conditions, thus indicating the progressive uplift of the orogen. Extensional processes still continue to the present, forming high-angle extensional faults with analogous geometry and kinematics.

Discussion and Conclusions

Taking into account the aforementioned description of the characteristic elements and timing of the Tertiary late-orogenic extension in the Hellenides, it can be established that it started in the more internal parts of the orogen and progressively proceeded towards the SSW to the more

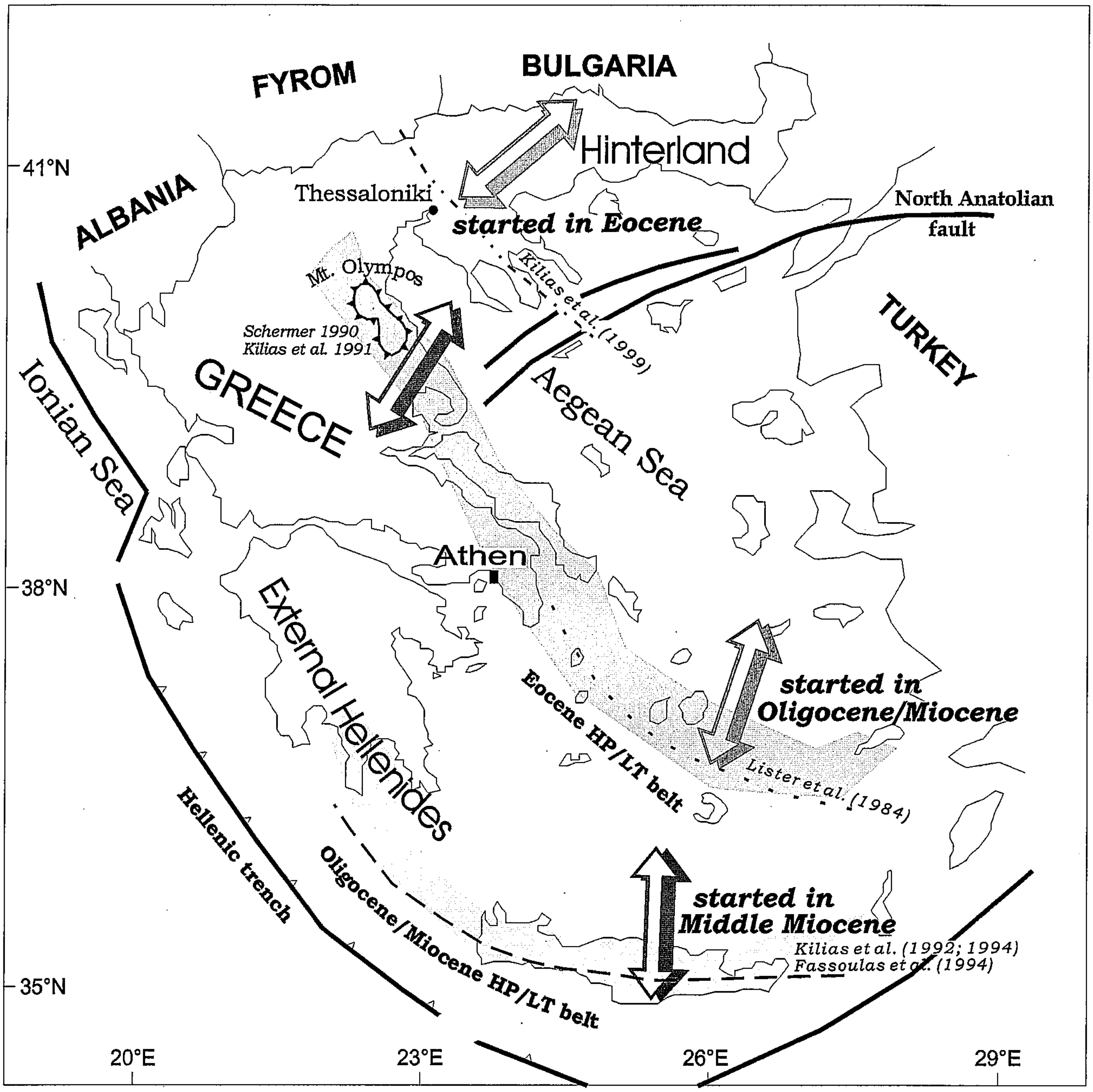

Figure 7.- SSW progressive migration of the extension in the Hellenides during the Tertiary. The ages correspond to the onset of extension (Kilias *et al.*, 1999).

external ones: (1) In the crystalline Rhodope massif, the onset of extension and collapse took place during the Eocene / Oligocene; (2) more externally, along the internal HP metamorphic belt (Olympos, Ossa, Pelion, and Cyclades), extensional collapse started during the Oligocene to Miocene transition, and (3) in the outermost parts, along the external HP metamorphic belt (in Crete) extension initiated during the middle Miocene.

Late-orogenic extension evolved in a continuing plate-convergence regime, wherein compression and crustal thickening at the outermost parts of the orogen took places simultaneously with extension at the core of the orogen (Fig. 4) (e.g., Schermer *et al.*, 1989; Kilias *et al.*, 1999). On the other hand, the late-orogenic extension occurred always after the compression and lithospheric thickening associated with HP metamorphism and subduction processes, considerably affecting HP metamorphic belts.

Consequently, we can conclude that during the Tertiary and under a constant plate-convergence regime, extension and compression in the Hellenic orogen constituted a dynamic system migrating progressively to the SSW (Figs. 4 and 7). As a result, the deep crustal levels were exhumed during extensional tectonics simultaneously with the underplating of lithospheric material and nappe stacking in the deformational front of the orogen (Figs. 3 and 4) (Kilias *et al.*, 1999).

When late-orogenic extensional tectonics took place in the back-arc area (i.e. crystalline Rhodope massif and Cyclades), orogenic collapse evolved asymmetrically; whereas extensional collapse above the cold accretionary prism (i.e. Olympos, Ossa, Pelion, and Crete areas) seems to fit the symmetric mode of deformation (Figs. 3, 4 and 5).

The differences in geometry and kinematics of the extensional deformation along the collapsed orogen could be ascribed to changes in the convergence rate between the African and Eurasian plates during the Tertiary in combination with a retreat of the subduction zone towards the S (Dercourt *et al.*, 1986; Royden, 1993). We suggest that the continuing subduction of Africa counteracted the high potential energy of the over-thickened Hellenic accretionary prism so that asymmetrical collapse was favoured in the back-arc area. Extensional collapse above the cold accretionary prism, on the other hand, developed symmetrically, possibly due to an increase in the buoyant stresses resulting from the subducted African slab in combination with the dynamic instabilities of the thick accretionary prism.

Acknowledgements

Constructive reviews from Demosthenis Mountrakis and an anonymus referee on a previous version are greatly appreciated and have helped to improve the manuscript. Financial support of 2000GR000, PB98-1334 and REN2001-3868-C03 research projects is acknowledged. We thank also C. Laurin for her careful and detailed linguistic revision of the final manuscript.

References

Cycladic crystalline complex (SE Pelagonian, Greece). *Geologishes Jahrbuch*, E2E: 97-164.

Aubouin, J. (1959): Contribution a l'étude géologique de la Grèce septentrionale : les confins de l' Épire et de la Théssalie. *Annales Géologiques des Pays Helléniques*, 10: 1-525.

Barton, C.M. (1976): The tectonic vector and emplacement age of an allochthonous basement slice in the Olympos area, NE Greece. *Bulletin du la Societé Geologique du France*, 18: 253-258.

Bird, P. (1978): Initiation of intracontinental subduction in the Himalayas. *Journal of Geophysical Research*, 83: 4975-4987.

Davis, G.A. y Lister, G.S. (1988): Detachment faulting in continental extension: Perspectives from the southwestern U.S. Cordillera. En: *Processes in continental lithospheric deformation* (S.P. Clark, B.C. Burchfiel y J. Suppe, Eds.). *Geological Society of America Memoir*, 18: 133-159.

Dercourt J. y 18 others (1986): Geological evolution of the Tethys belt from the Atlantic to the Pamir since the Lias. *Tectonophysics*, 123: 241-315.

Dewey, J.F. (1988): Extensional collapse of orogens. *Tectonics*, 7: 123-1139.

Dinter, D.A. (1998): Late Cenozoic extension of the Alpine collisional orogen, northeastern Greece: Origin of the north Aegean basin. *Geological Society of America Bulletin*, 110: 1208-1226.

Dinter, D.A. y Royden, L. (1993): Late Cenozoic extension in north-eastern Greece: Strymon Valley detachment system and Rhodope metamorphic core complex. *Geology*, 21: 45-48.

Doutsos, T., Pepiper, G., Boronkay, K. y Koukouvelas, I. (1993): Kinematics of the central Hellenides *Tectonics*, 12: 936-953.

Fassoulas, C., Kilias, A. y Mountrakis, D. (1994): Postnappe stacking extension and exhumation of high-pressure/low-temperature rocks in the island of Crete, Greece. *Tectonics*, 13:127-138.

Gautier, P. y Brun, P. (1994): Ductile Crust exhumation and extensional detachments in the central Aegean (Cyclades and Evia Islands). *Geodinamica Acta*, 7: 57-85.

Godfriaux, I. (1968): Étude géologique de la region de l' Olympe (Grèce). *Annales Géologiques des Pays Helléniques*, 19: 1-280.

Hamilton, W. (1987): Mesozoic geology and tectonics of the Big Maria Mountains region, southeastern California. En: *Mesozoic rocks of southern Arizona and adjacent regions* (W.R. Dickinson y M.A. Klue, Eds.). *Arizona Geological Survey Dig.*, 18: 33-48.

Housemann, G. A., McKenzie, D.P. y Molnar, P. (1981): Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergence belts. *Journal of Geophysical Research*, 86: 3651-3663.

Jacobshagen, V., Dürr, S., Kockel, F., Kopp, K.O., Kowalczyk, G., Berckhemer, H. y Buttner, D. (1978): Structure and geodynamic evolution of the Aegean region. En: *Alps, Apennines, Hellenides* (H. Closs, D.Roeder y K. Schmidt, Eds.). *IUG Scientific Report*, 38: 537-564.

Jolivet, L., Daniel, J.M., Truffert, C. y Goffe, B. (1994): Exhumation of deep crustal metamorphic rocks and crustal extension in back arc region. *Lithos*, 33: 3-30.

Kilias, A. (1991): Transpressive Tektonik in den zentralen Helleniden. Aenderung der Translationspfade durch die Transpression (Nord-Zentral Griechenland). *Neues Jahrbuch für Geologie und Paläeontologie, Monatshefte*, 20: 291-306.

Kilias, A. (2001). Late orogenic extension in Hellenides. *Bulletin of the Geological Society of Greece*, 34: 149-156.

Kilias, A. y Mountrakis, D. (1990): Kinematics of the crystalline sequences in the western Rhodope massif. *Geologica Rhodopica*, 2: 100-116.

Kilias, A. y Mountrakis, D. (1998): Tertiary extension of the Rhodope massif associated with granite emplacement (Northern Greece). *Acta Vulcanologica*, 10: 331-337.

Kilias, A., Frisch, W., Ratschbacher, L. y Sfeikos, A. (1991): Structural evolution and P/T metamorphic conditions of blue schists of E. Thessaly. *Bulletin of the Geological Society of Greece*, 25: 81-99.

Kilias, A., Fassoulas, C. y Mountrakis, D. (1994): Tertiary extension of continental crust and uplift of Psiloritis metamorphic core complex in the central part of the Hellenic arc (Crete, Greece). *Geologische Rundschau*, 83: 417-430

Kilias, A., Falalakis, G., Nastos, G. y Mountrakis, D. (1995): Tertiary extensional exhumation of the HP/LT Makrynitsa metamorphic complexes in Mt. Pelion (Eastern Thessaly). XV Crpatho-Balkan Congress, *Geological Society of Greece, Special Publication* 4: 48-52.

Kilias, A., Falalakis, G. y Mountrakis, D. (1999): Cretaceous-Tertiary structures and kinematics of the Serbomacedonian metamorphic rocks and their relation to the exhumation of the Hellenic Hinterland (Macedonia, Greece). *International Journal of Earth Sciences*, 88: 513-531.

Le Pichon, X. y Angelier, J. (1979): The Hellenic arc and trench system: A key to the neotectonic evolution of the

Eastern Mediterranean area. *Tectonophysics,* 60: 1-42.

Liati, A., (1986): Regional metamorphism and overprinting contact metamorphism of the Rhodope zone, near Xanthi, N. Greece: petrology, geochemistry, geochronology. *Ph.D. Dissert. Technische Universität Braunschweig,* Germany, 1-186.

Liati, A. y Gebauer, D. (1999): Constraining the prograde and retrograde P-T-t path of Eocene HP rocks by SHRIMP dating of different zircon domains: inferred rates of heating, burial cooling and exhumation for central Rhodope, northern Greece. *Contributions to Mineralogy and Petrology,* 135: 340-354.

Lips, A.L.W., White, SH. y Wijbrans, J.R. (1988): $^{40}Ar/^{39}Ar$ Laser prebe direct dating of discrete deformational events: a continuous record of early Alpine tectonics in the Pelagonian zone, NW Aegean area, Greece. *Tectonophysics*, 298: 133-153.

Lister, G.S., Banca, G. y Feenstra, A. (1984): Metamorphic core complexes of Cordilleran type in Cyclades, Aegean Sea, Greece. *Geology,* 12: 221-225.

Lister, G.S., Etheridge, M.A. y Symonds, P.A. (1986): Detachment faulting and evolution of passive continental margins. *Geology,* 14: 246-250.

Malavieille, J. (1993): Late orogenic extension in mountain belts: Insights from the Basin and Range and the late Paleozoic Variscan belt. *Tectonics,* 12: 1115-1130.

McKenzie, D. (1978): Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters,* 40: 23-32.

Meulenkamp, J.E., Wortel, M.J.R., Van Wamel, W.A., Spakman, W. y Hoogerduyn Strating, E. (1988): On the Hellenic subduction zone and the geodynamic evolution of Crete since the late middle Miocene. *Tectonophysics,* 146: 203-215.

Mountrakis D. (1986): The Pelagonian zone in Greece: A polyphase deformed fragment of the Cimmerian continent and its role in the geotectonic evolution of the Eastern Mediterranean. *Journal of Geology,* 94: 335-347.

Nance, D. (1981). Tectonic history of a segment of the Pelagonian zone, Northestern Greece. *Canadian Journal of Earth Sciences*, 18: 1111-1126.

Platt, J.P. (1986): Dynamics of orogenic Wedges and the uplift of high-pressure metamorphic rocks. *Geological Society of America Bulletin,* 97:1037-1053.

Robertson, A.H.F. y Dixon, J.E. (1984). Introduction: aspects of the geological evolution of the Eastern Mediterranean. En: *The geological evolution of the Eastern Mediterranean.* (Dixon, J.E. & Robertson, A.H.F. Eds.) Special Publication of the Geological Society, London. 17: 1-74.

Robertson, A.H.F., Dixon, J.E., Brown, S., Collins, A., Morris, A., Pickett, E., Sharp, I. y Ustaömer, T. (1996): Alternative tectonic models for the Late Paleozoic-Early Tertiary development of Tethys in the Eastern Mediterranean. En: *Palaeomagnetism and Tectonics of the Mediterranean Region* (A. Moris y E.H. Tarling, Eds.). Geological Society of London Special Publication, 105: 239-263.

Royden, L.H. (1993): Evolution of retreating subduction boundaries formed during continental collision. *Tectonics,* 12: 629-638.

Ruppel, C. (1995): Extensional processes in continental lithosphere. *Journal of Geophysical Research*, 100: 24187-24215.

Schermer, E. (1990): Mechanisms of blueschist creation and preservation in an A-Type subduction zone, Mount Olympos region, Greece. *Geology,* 18: 1130-1133.

Schermer, E., Lux, D. y Burchfiel, B. (1989): Age and tectonic significance of metamorphic events in the Mt. Olympos region (Greece). *Bulletin of the Geological Society of Greece,* 23: 13-27.

Schliestedt, M., Altherr, R. y Mathhews, A. (1987): Evolution of the Cycladic crystalline complex: petrology, isotope geochemistry and geochronology. En: *Chemical Transport in Metamorphic Processes.* (H.C. Helgeson, Ed.). NATO ASI Ser., 389-428.

Seidel, E., Kreuzner, H. y Harre, W. (1982): A Late Oligocene/ Early Miocene high-pressure belt in the External Hellenides. *Geologisches Jahrbuch,* 23: 165-206.

Sfeikos, A., Boehringer, C., Frisch, W., Kilias, A. y Ratschbacher, L. (1991): Kinematics of Pelagonian nappes in Kranea area, North Thessaly, Greece. *Bulletin of the Geological Society of Greece,* 25: 101-105.

Spakman, W., Wortel, M.J.R. y Vlaar, N.J. (1988): The Hellenic subduction zone: a tomographic image and its geodynamic implications. *Geophysical Research Letters,* 15: 60-63.

Thomson, S.N., Stoeckhert, B. y Brix, M.R. (1998): Thermochronology of the high-pressure metamorphic rocks of Crete, Greece: implications for the speed of tectonic processes. *Geology,* 26: 259-262.

Vergely, P. (1984): *Tectonique des ophiolites dans les Hellénides interes. Consequences sur l'evolution des regions Téthysiennes occidentals.* Tesis doctoral Univ., 1-650.

Wawrzenitz, N. (1997): *Mikrostrukturelle unterstützte Datierung von Deformations inkrementen in Myloniten: Dauer der Exhumierung und Aufdomung des metamorphen Kern-komplexes der Insel Thassos (Süd-Rhodope, Nordgriechenland).* Tesis doctoral Univ., Nuernberg, 1-180.

Wawrzenitz, N. y Mposkos, E. (1997): First evidence for Lower Cretaceous HP/LT metamoprhism in the Eastern Rhodope, North Aegean region, North-East Greece. *European Journal of Mineralogy,* 9: 659-664.

Wernicke, B.P. (1981): Low-angle normal faults in the Basin and Range Province: Nappe tectonics in a extending orogen. *Nature,* 291: 645-648.

Wernicke, B.P. (1985): Uniform-sense normal simple shear of the continental lithosphere. *Canadian Journal of Earth Sciences,* 22: 108-125.

Wijbrans, J.R. y McDougall, I. (1988): Metamorphic evolution of the Attic-Cycladic Metamorphic Belt on Naxos (Cyclades, Greece) utilizing $^{40}Ar/^{39}Ar$ age spectrum measurements. *Journal of Metamorphic Geology,* 6: 571-594.

Yarwood, G. y Dixon, J. (1977): lower Cretaceous and younger thrusting in the Pelagonian rocks of the high Pierie, Greece. *VI Coll. Geol. Aegean region, Athen*, 269-280.

Manuscrito recibido el 25 de enero de 2002
Aceptado el manuscrito revisado el 10 de julio de 2002