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Extension of thickened continental crust, from brittle to ductile deformation: examples from Alpine Corsica and Aegean Sea

Laurent Jolivet
Département de géologie, Ecole Normale Supérieure, Paris, France

Abstract
Collapsed mountain belts are useful to understand the geometry of crustal extension, because metamorphic core complexes usually show how extension proceeds at depth, in the middle of lower crusts. The deformation seen in such particular contexts might however not entirely represent the behaviour of a normal continental crust subjected to rifting. Still they are the only places where deep ductile extension is accessible to direct observation by geologists, and we use them to describe how extension proceeds at various depths, based on regional examples with various finite rates of extension. The Alpine nappe stack of the Cyclades (Aegean Sea) has been collapsing since the early Miocene. Large rates of extension have produced several basins now floored with continental crust less than 20 km thick (Cretan Sea and North Aegean Trough). A HT metamorphism which dates back to the early Miocene is superimposed to the HP metamorphism which resulted from the Eocene crustal thickening. The gradient ranges from greenschists to amphibolites and migmatites in the center of the Cyclades (Paros and Naxos). The sense of shear seen along the major detachement in Naxos is the same as the one imposed by block tilting in the active domain, in Evia and Attica. So that the ductile deformation seen in Naxos might represent what is presently occurring in the middle and lower crust below the tilted blocks of Evia and Attica. In Corsica, the Alpine nappe stack collapsed when Corsica was separated from the main belt of the Alps and Apennines in the Late Oligocene. The HP metamorphism is here also reworked by a retrogression in greenschist conditions contemporaneous with extension. Alpine Corsica shows extension proceeding in the brittle-ductile transition zone. These two examples can be used to construct a synthetic cross-section of a continental crust undergoing extension. The upper crust accommodates extension by normal faulting and block tilting. Sense of tilt is consistent over the entire width of the rift zone. The transition to the lower crust is made through flat-lying shear zones which are the prolongation at depth of normal faults. In the brittle-ductile transition, strain is partitioned between zones of high simple shear strain, synthetic of the slip along normal faults, and zones of predominant flattening. This departs from the classical simple shear model.

1. Introduction

In the simple shear model of crustal extension, often used to describe the formation of Atlantic-type passive margins, one assumes that a large flat-lying extensional shear zone cuts through the entire crust and upper mantle (Wernicke, 1985; Boillot *et al.*, 1987). The model is derived from the observation of metamorphic core complexes where a flat-lying detachment divides an upper plate which shows only brittle deformation from a lower plate where ductile deformation is predominant (Wernicke, 1981; Seranne and Seguret, 1987; Lister and Davis, 1989; Andersen and Jamveit, 1990). This geometry can conveniently explain the frequent asymmetry of crustal structure of intra-continental rifts or passive margins where all normal faults dip in the same direction. But several facts go against this model: 1) deep seismic profiles in continental rift area never show reflectors cutting the entire crust, normal faults rooting usely at depth in flat-lying shear zones within the lower crust (Brewer and Smythe, 1984; Cheadle *et al.*, 1987). 2) Active normal faults with well-characterized focal mechanisms dip steeply and are plane down to the base of the upper crust (King *et al.*, 1985, Jackson and White, 1989). 3) Ductile extensional deformation is observed where extension occurs subsequently to crustal thickening, never in classical passive margin context. The reason is straightforward: in a classical intra-continental rift extension is set on a crust of a normal thickness, (30÷35) km, and extension is due to boundary forces. The upper half of the crust , down to (10÷15) km, is in the brittle field, the lower half being ductile. In the case of a thickened crust the ductile layer is expected to be much thicker than the brittle one. Extension can be either due to extensional boundary conditions or/and to internal body forces due to abnormal thickness. This has two major effects: 1) the deformation more widely distributed in the case of the thickened crust and 2) soon after the onset of extension, for a small amount of extension, a normal continental crust will be under sea level and sedimented, while a thickened crust would remain unsedimented until larger amounts of extension. One can thus never observe ductile rocks cropping out in a passive margin environment because it would require such large amounts of extension which cannot be attained before the incoming of sea water in the rift zone.

We study here the geometry of brittle-ductile coupling and uplifting of metamorphic rocks based on field examples in Corsica and Aegean Sea. Various amounts of extension during the Miocene after Paleogene crustal thickening led to the uplift of metamorphic complexes formed at various depths. Using those two examples we describe a geometry for extension and uplift of deep crustal material both in time and space in the context of a collapsing thick crust. The model describes both the geometry of brittle-ductile transition in space and the progressive incorporation of ductile material in the upper crust with time.

2. Aegean Sea and the Cyclades

2.1. *Tectonic context*

The Aegean Sea is situated above the African slab which has subducted northward since at least 30-40 Ma. It is the site of distributed extension in a back-arc environment (fig. 1). The distribution of seismic deformation, its kinematics and mechanics have been extensively studied in the last 20 years (McKenzie, 1972; Le Pichon and Angelier, 1979; Le Pichon, 1981; Jackson *et al.*., 1982; Angelier *et al.*, 1982; Mercier *et al.*, 1987; Taymaz *et al.*, 1991). A very high crustal seismicity results from extensional deformation which was initiated in the early Miocene as attested by the oldest marine sediments found in Naxos and Evia as well as by radiometric ages of high-temperature metamorphics related to extension in Naxos (Robert, 1982; Wijbrans and McDougall, 1988; Gautier *et al.*, 1990). Active extension is taken up by normal and strike-slip faults as well as block rotations clearly demonstrated in continental Greece (Laj *et al.*, 1982; Kissel *et al.*, 1986).

Extension was set on a thickened crust which can still be seen either in continental Greece or in Turkey. Large rates of extension have produced several basins now floored with continental crust less than 20 km thick (Cretan Sea and North Aegean Trough) (Makris, 1978; Le Pichon and Angelier, 1979). Most of the islands where ductile extension can be observed rest on a 30 km thick crust, which corresponds to presumably half the initial thickness before extension started.

A gradient of extension and crustal thickness is observed along the arc from the 45 km thick crust in northern Greece to (30÷20) km in the central Cyclades. Parallel to this gradient of crustal thickness, a gradient of HT metamorphism is observed. It reworks an older HP-LT event related to Eocene crustal thickening (Blake *et al.*, 1981; Bonneau and Kienast, 1982). A gradient of pressure is also seen toward the Cyclades during this stage. Avigad and Garfunkel (1991) proposed that the gradient is due to progressively greater finite extension southeastward, and unroofing of progressively deeper parts of the crust.

Crustal thickening was active in the Eocene in the central Aegean. It was still active in the Early Miocene in the most external zones in Crete while extension was already active in the Cyclades. Two parallel high-pressure belts have been described: the Cycladic blueschists have radiometric ages which cluster about 45 Ma, and in Crete and Peloponnesus high-pressure nappes yield ages around 20 Ma. In both regions pressure conditions reached 10 to 12 kbar. According to

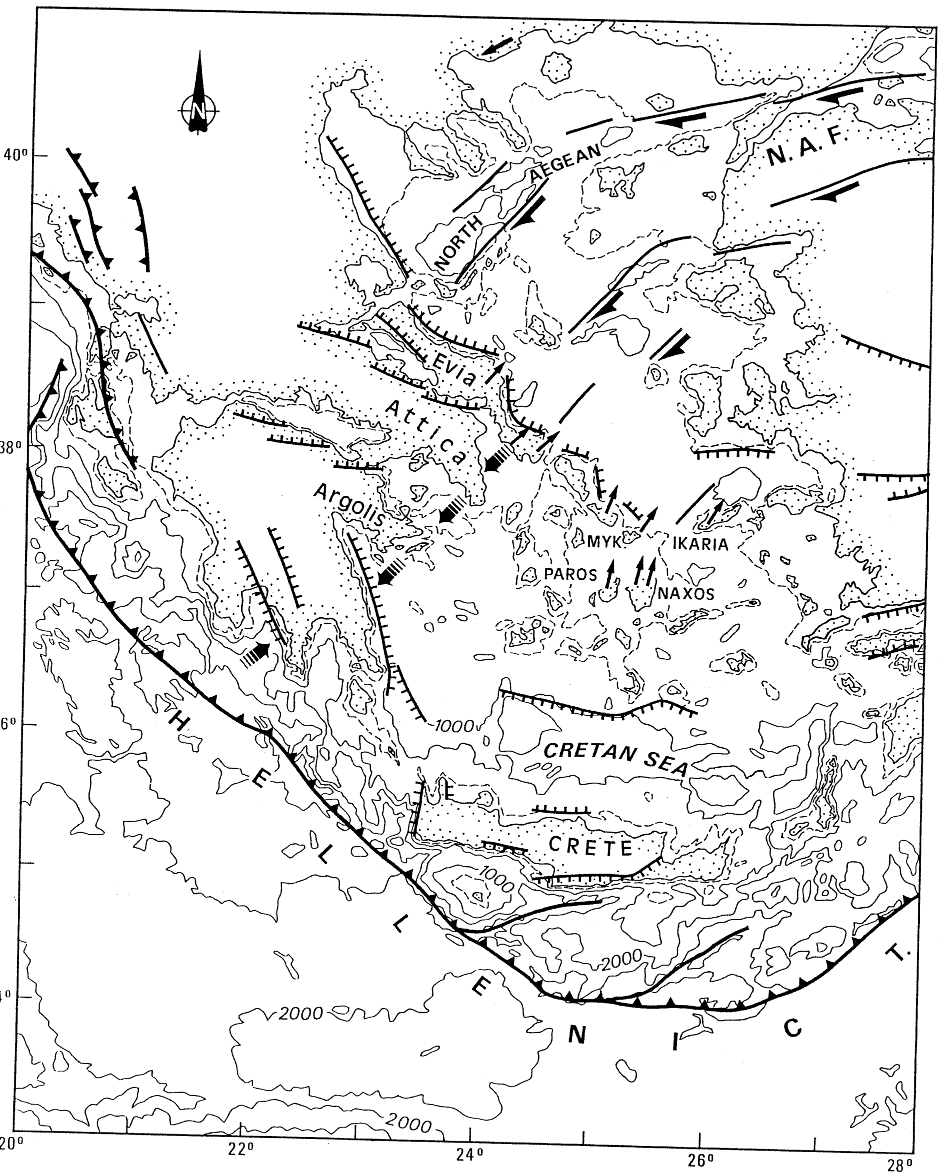

Fig. 1. Tectonic map of the Aegean region showing the major faults and sense of ductile shear. The northern part of the stretched domain is invaded by satellites of the North Anatolian Fault which accommodates the westward extrusion of Anatolia. This strike-slip regime is observed up to the western end of the North Aegean trough where it is transtensional. Evia, Attica, Argolis and the southern peninsula in the Peloponnesus are tilted blocks separated by large normal faults. Only the major north-dipping faults which root in flat-lying detachement at depth, are shown. Small black arrows represent the direction and sense of Miocene shear in the metamorphic core complexes of the Cyclades. Half large arrows represent the sense of slip along strike-slip faults. Large black arrows show the sense of dip of sediment infilling between tilted blocks.

Bonneau (1984) the Cycladic blueschists results from the closure of the small Pindos oceanic basin, and the blueschists themselves represent the subducted oceanic crust of that basin.

The related oceanic sediments escaped from being deeply buried and are now part of the superficial Pindos nappe. The oceanic units are sandwiched between continental units. Above, the Pelagonian nappes represent a continental basement and its sedimentary cover together with pieces of oceanic crust obducted in an earlier stage. The base of the Pelagonian units was metamorphosed under MP-LT conditions. Below the Cycladic blueschist and Pindos nappe, several continental units were stacked after the closure of the oceanic domain. Active convergence south of Crete gives rise to a large accretionary prism, the Mediterranean Ridge. Crustal seismicity related to extension is recognized far south of the Hellenic Trench s.s. which is a series of fore-arc grabens.

2.2. *Geometry of extension*

Lower crustal material is seen in the center of the Cyclades in the metamorphic core complex of Naxos first described by Lister *et al.* (1984). A typical flat-lying detachment separates the upper plate with only brittle deformation from the lower plate with ductile deformation associated with migmatites and amphibolites (fig. 2). The sense of ductile shear is top-to-the-north compatible with the direction of extension and with the sense of block tilting in the upper plate (Urai *et al.*, 1989; Buick, 1991; Gautier *et al.*, 1990).

Extension is active in the Aegean region north and west of Naxos. This active deformation is responsible for the formation of upper crustal blocks which are all tilted southward (King *et al.*, 1985; Papanikolaou *et al.*, 1988) (fig. 1 and 3).

Moving along the strike of these blocks one can observe the transition from brittle to ductile. Active extension proceeds by steeply dipping normal faults reaching the brittle-ductile transition at a depth of 10 km and connecting to flat-lying shear zones with a lower seismicity and top-to-the-north sense of shear as shown by King *et al.* (1985) (fig. 4). The sense of ductile shear seen in Naxos and everywhere in the Cyclades is-top-to-the-north as well (Faure *et al.*, 1991; Gautier *et al.*, 1990). We think that the deformation seen in Naxos is equivalent to that occurring at depth in the deep crust beneath the active deformation zone.

A metamorphic gradient can be observed below the detachement from poorly metamorphosed cataclasites, immediately below the upper plate to HT migmatites with ductile deformation in the center of the gneiss dome. High

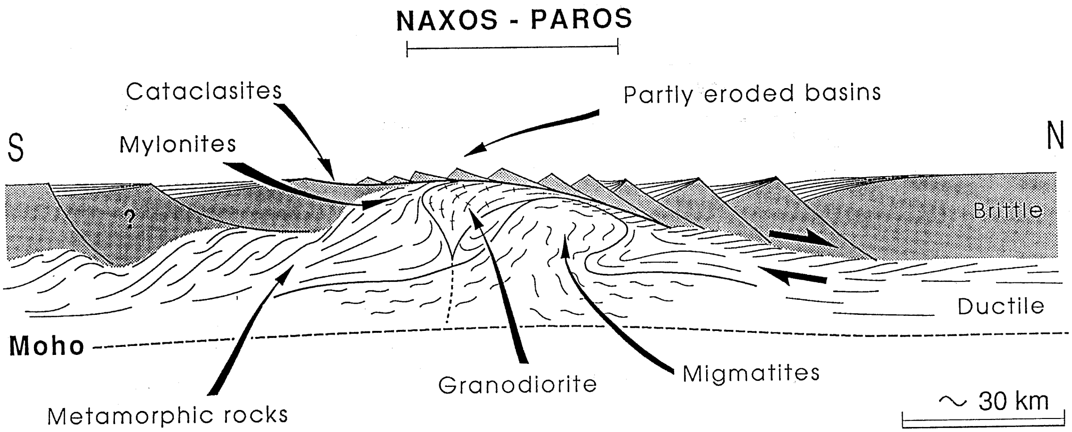

Fig. 2. Idealized cross-section in the Paros-Naxos metamorphic core complex after Gautier *et al.* (1990). Deep ductile metamorphism seen there corresponds to the uplift of lower crust below a major detachment which leads to the very large thinning of the upper crust. The sense of shear along the detachment is coeval of the sense of slip along brittle normal faults in the upper plate.

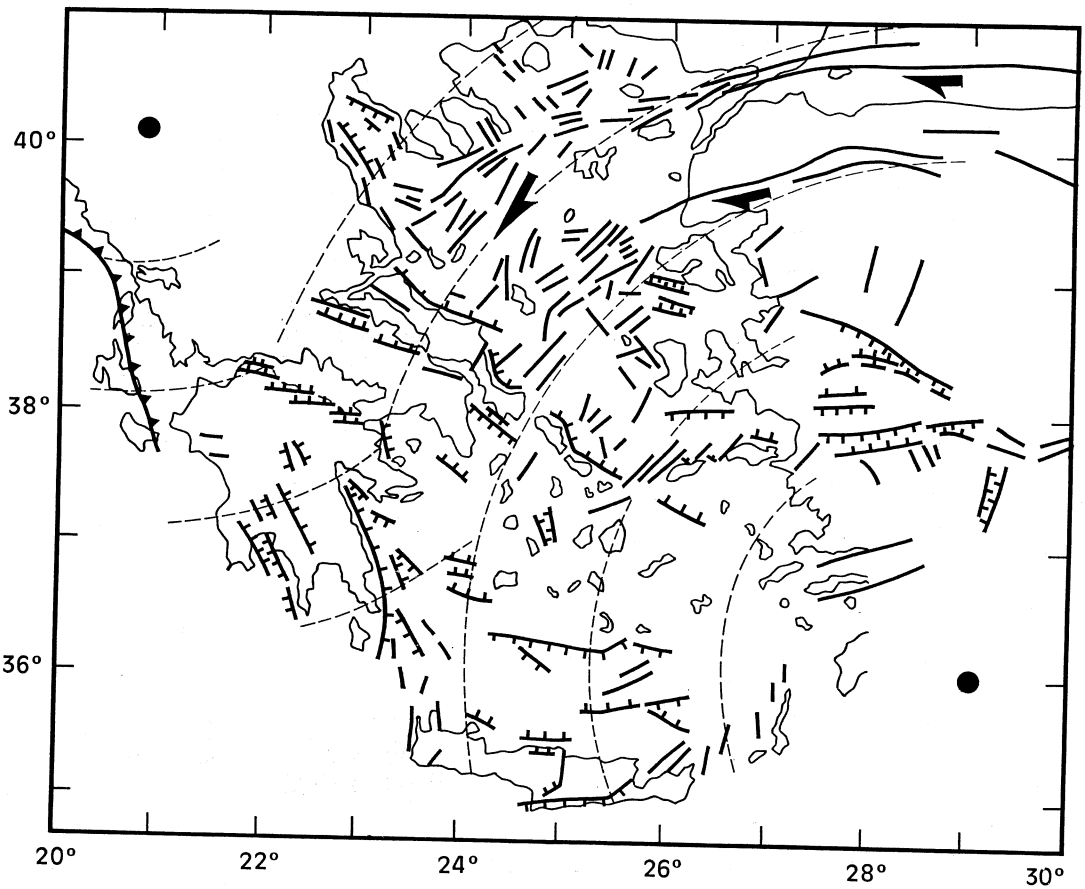

Fig. 3. Brittle structures in the Aegean Sea after Mascle and Martin (1990). Strike-slip faults are distributed in the northern part of the stretched domain and normal faults in the southern part of it. Also shown are small circles about two poles (black dots) which approximately fit the change in strike of strike-slip faults from the E-W trending North Anatolian Fault to central Aegea.

shear strain has been localized along the detachement and a consistent top-to-the-north sense of shear gave rise to a thick shear zone which evolved in time from ductile to brittle. The migmatites yield radiometric ages of 25 Ma (Wijbrans and McDougall, 1988). Later during the shear process a granodiorite intruded the dome at 11 Ma and was affected by the detachement afterwards. Brittle deformation within the upper plate gives shear criteria which are compatible with those of the lower plate, and Miocene sediments were deposited in small tilted basins controlled by north-dipping normal faults. Similar structures were described further north in Mykonos (Faure and Bonneau, 1988).

Everywere the same top-to-the-north or north-east sense of shear is observed associated with a retromorphosis of high-pressure rocks in HT conditions with a gradient along strike from greenschists in Evia to amphibolite and anatexis in Naxos.

Figure 5 shows the direction and sense of shear in several places in the Aegean. One important observation is that the direction changes progressively from South to North. One can easily notice that in the north the direction of shear is parallel to the North Aegean Trough. The small circles shown in this figure are about two poles

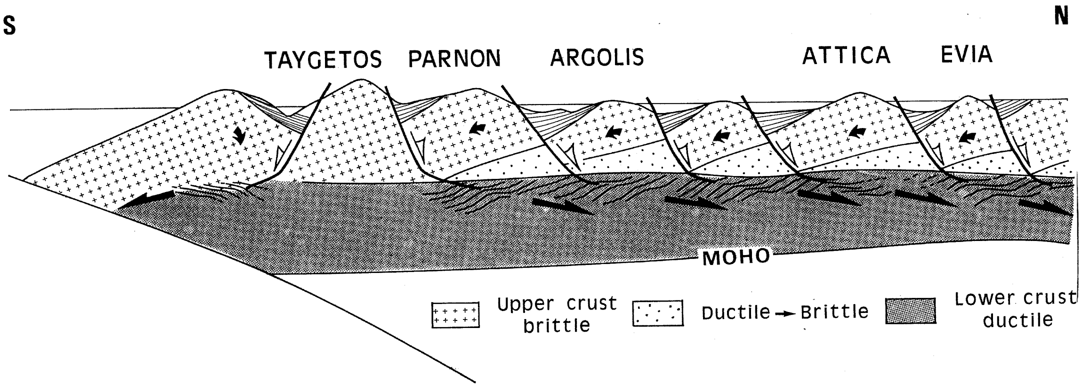

Fig. 4. Interpretation of the brittle-ductile relation along a NE-SW profile trough the Hellenic arc. Tilt of sediment infill is after Papanikolaou *et al.* (1988) and the geometry of normal faults after King *et al.* (1985).

at either extremities of the arc (see also fig. 3). These circles are the same as those shown in fig. 3 which fit the direction of dextral strike-slip faults and which are approximately perpendicular to normal faults.

The same progressive change in strike can be seen also in the dextral strike-slip earthquakes slip vectors. This parallelism between superficial brittle deformation and finite deep ductile one suggests that the deformation of the upper crust 1follows passively a ductile flow in the lower crust.

This flow is in turn a consequence of the roll back of the Hellenic trench which drags the Aegean lithosphere southward between the two poles shown in fig. 3. Such a ductile flow leads to a distributed polar flow in regions of reduced crustal thickness. In this hypothesis 1) block rotations shown by paleomagnetic data (Laj *et al.*, 1982; Kissel *et al.*, 1986) would be restricted to regions of thick crust, and 2) block rotations would only affect the upper part of the crust.

A simplified interpretation of the 3D geometry is shown in fig. 6. The metamorphic core complex of Naxos and Paros is seen as through a window opened in the upper crust by a large amount of extension.

Further to the northeast brittle extension is observed with normal faults dipping to the northeast, toward the Aegean Sea. Papanikolaou *et al.* (1988) showed that the major normal faults in this region are all north-dipping and that the sediments deposited in the basins between the blocks are tilted toward the south thus giving a sense of shear compatible with the one seen in rocks deformed at depth (fig. 4). This suggests that the same deformation has been active for at least 25 Ma.

3. Alpine Corsica and the Tyrrhenian Sea

3.1. *Tectonic context*

The Tyrrhenian Sea opened in the back-arc region of the Apennines convergence zone and the Calabrian Arc (fig. 7).

Most of it is floored with thinned continental crust and with oceanic crust in two small basins in its southeastern corner, north of Sicily. Oceanic spreading began 5 Ma ago in the western basin and shifted eastward 2 Ma ago in the eastern basin where it is still active.

Extension and crustal thinning also followed the same trend of eastward migration. There is a younging toward the east starting with the eastern margin of Sardinia which was rifted between 15 and 10 Ma (Kastens and Mascle, 1990; Kastens *et al.*, 1988; Mascle and Réhault, 1990; Sartori, 1990). In the northern part of the Tyrrhenian Sea, rifting started in Alpine Corsica where it reworked the Alpine nappe stack in the late Oligocene and early Miocene (Burdigalian) (Jolivet *et al.*, 1990). More recent extensional features are observed further east in the Alpi Apuane (Carmignani and Kligfield, 1990), and in Monte Cristo and Elba islands where detachements were

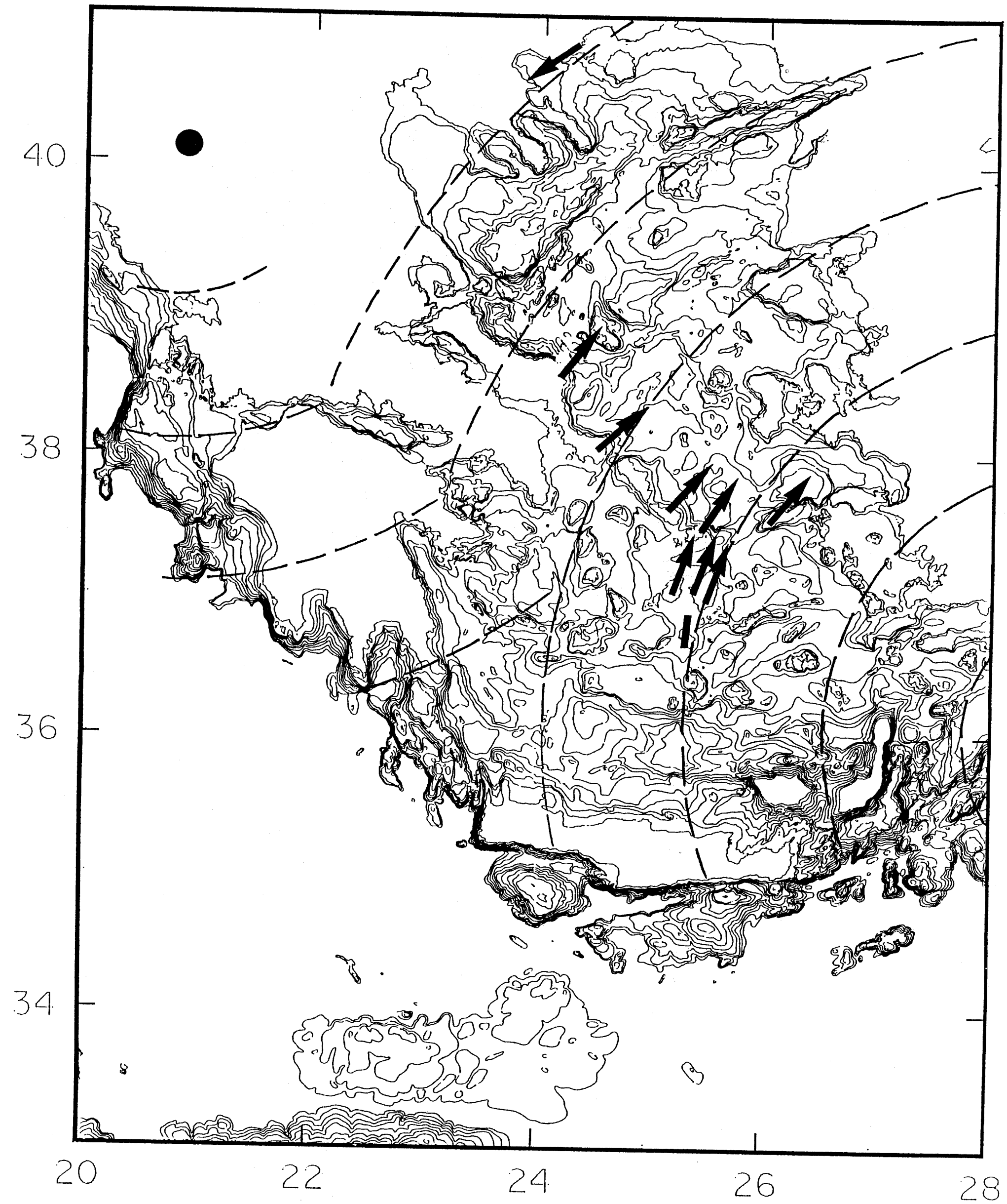

Fig. 5. Bathymetry of the Aegean sea (100, 200, 400 ... 2000 m curves are shown) and ductile shear directions and senses. Paros and Naxos are on a bathymetric high between two tilted blocks (eastward extensions of Evia and Attica). Also shown are the small circles defined in fig.2 which also approximately fit the ductile shear direction. Ductile and brittle deformation follow the same finite geometry.

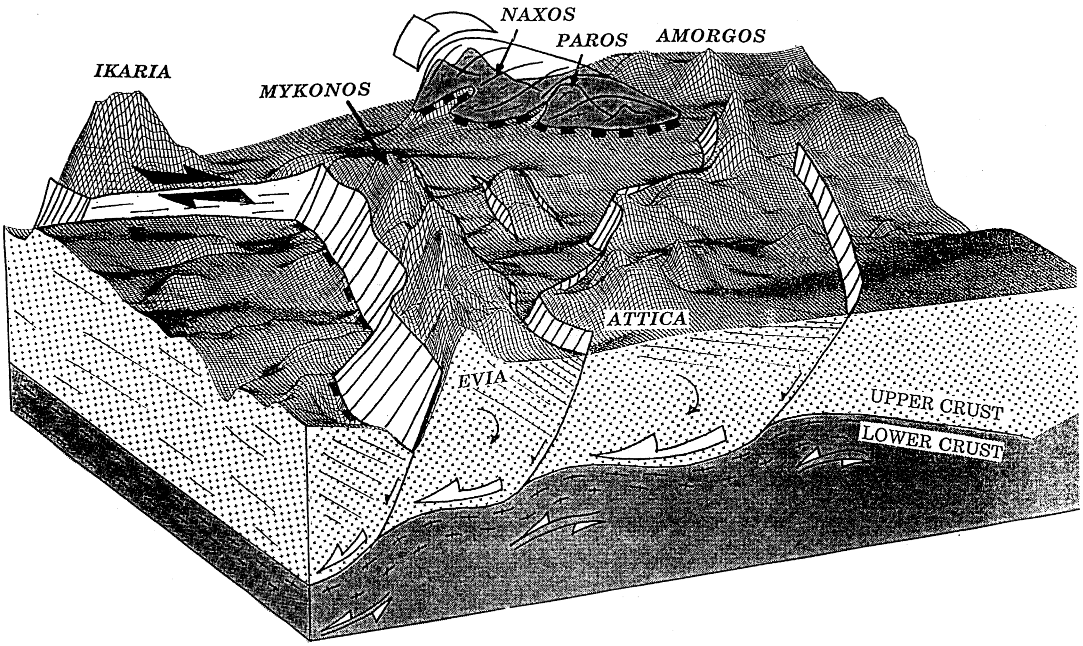

Fig. 6. Schematic 3D interpretation of the brittle-ductile relation in the Aegean Sea. Bird eye view of the International Bathymetric Chart of the Mediterranean Region, seen from the northwest. Deep crustal rocks in Paros and Naxos correspond to uplifted lower and middle crust which is still actively deforming below Evia and Attica.

formed during the intrusion of granites between 9 and 5 Ma (Bouillin, 1983; Daniel *et al.*, in preparation).

Alpine Corsica provides an example where shallow-dipping normal shear zones of crustal scale are actually observed in the transitional domain between brittle and ductile crust (Jolivet *et al.*, 1990; 1991; Fournier *et al.*, 1991).

Extension reworked a nappe stack formed during the early stages of Alpine orogeny from the late Cretaceous to the late Eocene (fig. 8 and 9). Crustal thickness is 25 km under Alpine Corsica (Hirn and Sapin, 1976; Morelli *et al.*, 1976). High-pressure low-temperature metamorphic rocks were reworked during their way back to the surface with decreasing P-T conditions through the greenschist facies.

Two major non-coaxial flat-lying shear zones localized the extensional strain: one was the contact between the Schistes Lustrés nappe and the continental basement of western Corsica, the other was the base of the non-metamorphosed Balagne ophiolitic nappe. The deformation evolved in time from ductile to brittle and previously ductile rocks were progressively incorporated in the brittle crust. Toward the end of this process the crust was covered with shallow marine sediments which were deposited in half grabens controlled by the two major extensional shear zones (fig. 9).

The sense of tilt of the upper crustal blocks and the sense of ductile shear are both top-to-the-east. The distribution of the ductile strain also prefigurates the geometry of the upper crustal blocks. This example shows how preexisting thrusts are reworked in the middle crust and localize the extensional ductile strain, consequently controlling the geometry of upper crustal blocks.

3.2. *HP-LT structures, compression*

Alpine tectonic history in Corsica started with crustal thickening in the Late Cretaceous and

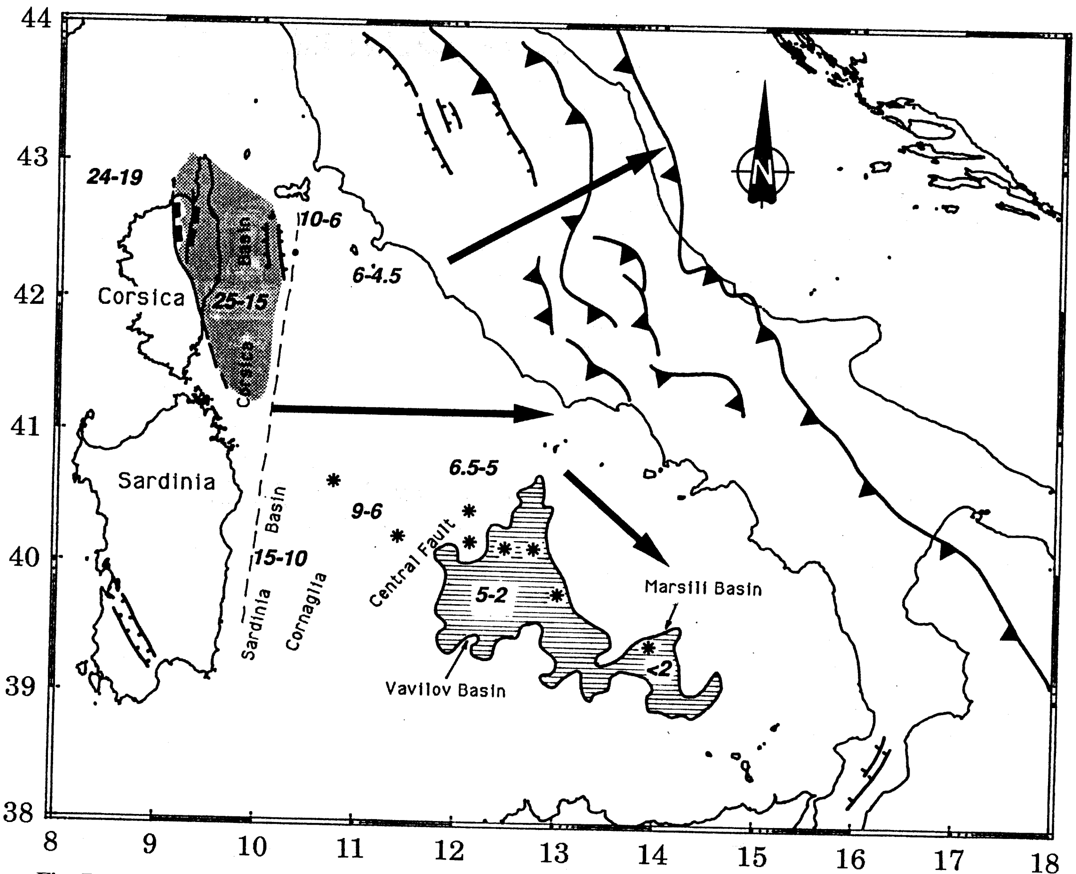

Fig. 7. Map of the Tyrrhenian Sea showing the ages of extension. Ages are given in millions of years before present. Arrows show the trend of younging toward the west for both extension in the basin and crustal thickening in the mountain range. The oceanic crust of Vavilov and Marsili basins is horizontally ruled. Shaded area represents the region of Alpine Corsica and Corsica basin where the oldest extensional events are recognized.

Eocene. West-vergent large overthrust brought a set of oceanic units onto the continental basement of Europe (Caron, 1977; Mattauer *et al.*, 1981, Amaudric du Chaffaut, 1982, Warburton, 1986; Waters, 1990). The continental margin was deformed during the overthrusting episode, and involved in the nappe stack. High-pressure and low-temperature metamorphism occurred during this episode and the P-T conditions of 14 kbar-450 °C were reached in some units, continental or oceanic (Caron *et al.*, 1981, Caron and Péquignot, 1986; Lahondère, 1988). These eclogites were then uplifted during continuing overthrusting while the thrust front progressed westward onto Western Corsica. During the uplift they retrogressed into blueschists. Clear westward shear sense criteria are associated with these early episodes, and simple shear predominated over flattening (Mattauer *et al.*, 1981).

3.3. *Greenschists structures, extension*

Where greenschist paragenes, which overgrow the earlier HP-LT ones, are observed, they correspond to a different geometry of deformation. They are usually associated with eastward sense of shear, and sometimes with coexisting

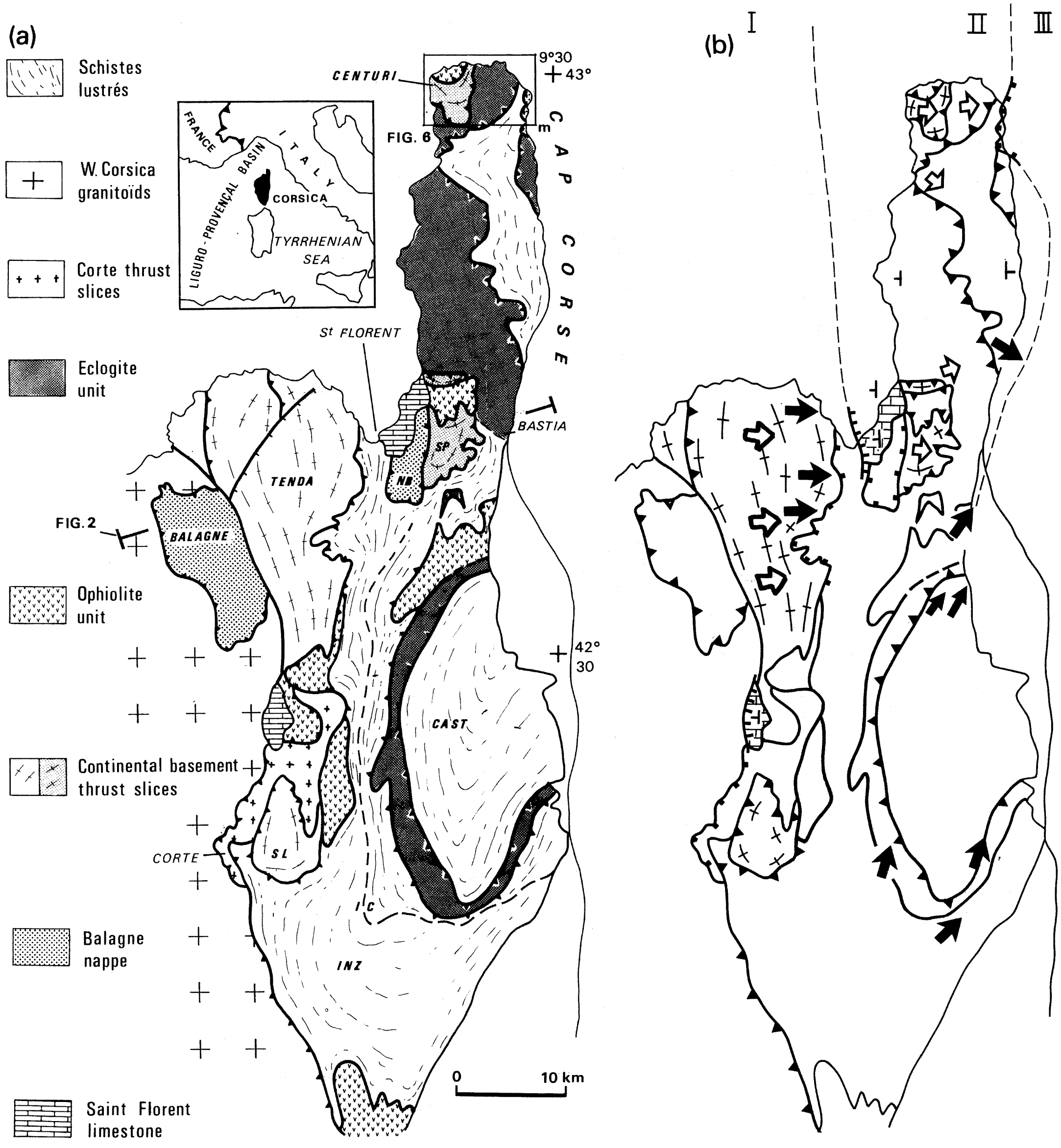

Fig. 8. Geological map of Alpine Corsica (a)) and sense of ductile shear during the greenschist retrogression (b)) (after Jolivet *et al.*, 1990, modified). Black arrows represent predominant eastward simple shear, and open arrows represent distributed flattening and eastward shear.

westward and eastward ones (Fournier *et al.*, 1991). The greenschist deformation is partitioned between zones of high simple shear strain, with top-to-the-east sense of shear, and zones of predominant flattening with both senses observed. This partition is seen at all scales (Jolivet *et al.*, 1991; Daniel *et al.*, in preparation) (fig. 10).

Domains with predominant flattening show a

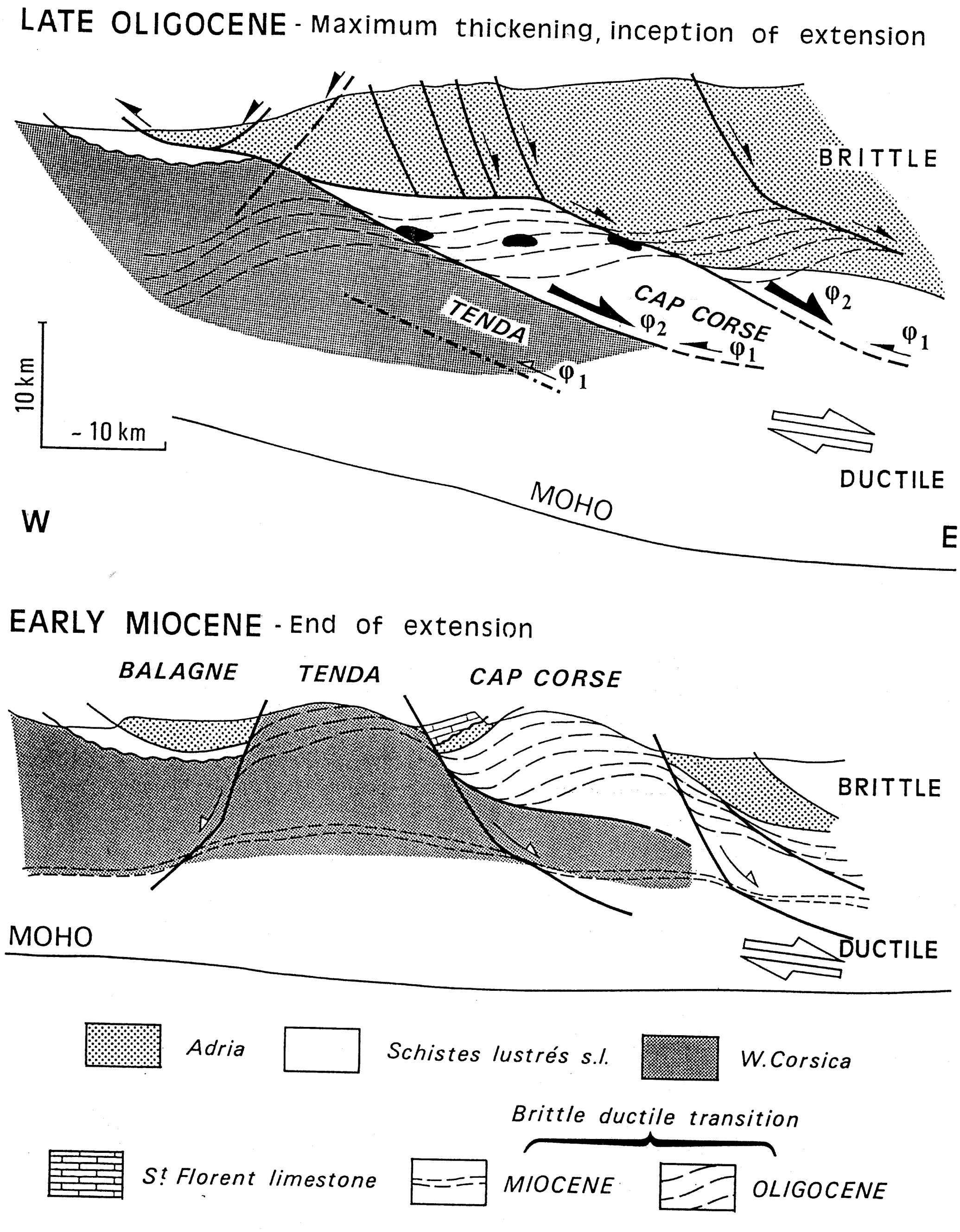

Fig. 9. Geometry of Alpine Corsica in the late Oligocene when extension began, and at present. The position of the Oligocene brittle-ductile transition is shown in both stages. The distribution of strain is shown in the brittle-ductile transition.

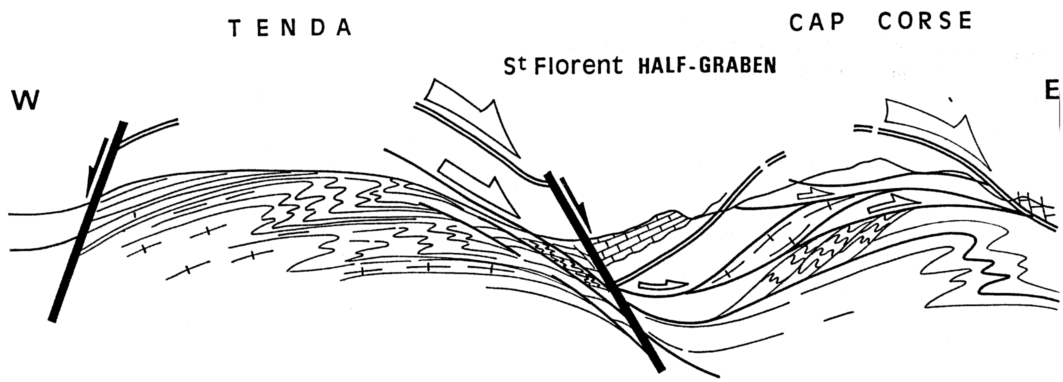

Fig. 10. Cross-section of Alpine Corsica showing the late structural features related to extension. The early HP-LT foliation is reworked by tight folds with west-dipping axial planes. They imply a strong flattening component in the Tenda massif and Cap Corse. In between is a zone of high eastward simple shear strain west of the Saint Florent half graben.

crenulation cleavage, deeping to the west, and associated with tight folds which rework the HP-LT foliation. Shear bands dipping either eastward or westward were formed simultaneously.

Eastward shear was localized along east-dipping shear zones which occur at all scales from centimetric to kilometric. Earlier large overthrusts localized the eastward shear as they were reactivated as ductile normal faults. There is no doubt that this deformation is related to crustal extension for the following reasons:

1) The recent deformation was partly achieved during a top-to-the-east ductile shear close to the brittle-ductile transition and was later superimposed by brittle shear indicating a transition in time from ductile to brittle regime. 2) Extensional brittle structures in the Early Miocene St Florent limestone and sense of tilt are compatible with the eastward sense of shear observed in the ductile rocks. 3) The movement along major «thrust» contacts is associated with retrograde metamorphism which overprinted the early high-P low-T paragenesis at less severe P-T conditions. They also bring tectonic units with contrasted metamorphic evolutions into close contacts. 4) There is a regional correlation between retromorphosis and recent deformation since the high-P low-T paragenesis is better preserved in southern of Alpine Corsica where the recent deformation is less pervasive. 5) Highly non-coaxial deformation is localized along east-dipping shear zones close to brittle normal faults which bound tilted Miocene basins; in between the geometry is more symmetric and the finite strain therefore more coaxial. The large kilometric-scale east-dipping shear zones clearly produced extension at crustal scale. The fact that the same geometry is seen at all scales implies that small-scale structures, which are penetrative inside the entire nappe stack, are also related to extension. 6) Late extensional brittle structures are observed at many sites in the metamorphic rocks.

4. Discussion

These two examples are symptomatic of more general features seen in many metamorphic core complexes, concerning both the uplift history of HP-LT rocks and the geometry of extension in the crust.

1) Only the last part of the retrograde P-T path of HP-LT metamorphic rocks corresponds to litospheric extension. The earlier part of it, from eclogites to blueschist is contemporaneous with crustal thickening during convergence. A cool gradient is maintained by addition of cool units under the nappes stack as the wedge front moves forward (Davy and Gillet, 1986). Extension is probably active at the same time in the upper part of the belt to allow fast unroofing of the deep crustal units (Platt, 1986). After crustal

thickening has stopped, and lithospheric extension has started, the uplift path then follows a higher-temperature gradient, and extensional structures are related to high-temperature and low-pressure mineral associations.

The detachment seen in Corsica or in the Cyclades below the unmetamorphosed ophiolitic units thus has a two-staged history. The early stage corresponds to surficial extension during continuing crustal thickening, and the more recent one is related to true lithospheric extension which leads to crustal collapse and ultimately formation of a rift with a thin crust.

2) The two examples of Corsica and Cyclades also show how ductile and brittle deformation are related to one another (fig. 11). The upper brittle crust deforms by normal faults which bound tilted blocks. The sense of tilt is consistent over large regions of the collapse domain as shown by the Aegean example. The major normal faults root in flat-lying shear zones in the domain of brittle to ductile transition. We see such shear zones in Corsica where they rework the ancient thrust contacts as ductile normal faults. Their relation to superficial structures is attested by the presence of tilted Miocene sediments which were deposited above the detachement. There is a partition of strain inside the brittle-ductile transition zone, between these zones of high simple shear strain, and zones of a large flattening component between them.

It can be argued that this partition corresponds to a succession in time of two events: a deeper event associated with flattening and a subsequent localization of strain along flat-lying shear zones, while the rocks are coming up to the surface. There is no definite answer to this discussion in our field observations.

Under the brittle-ductile transition, the behaviour of rocks is entirely ductile and strain is less localized. The Moho below a collapsing orogen is generally flat though there is strong localization of strain along discrete shear zones in the upper and the middle crusts. This leads to the conclusion that extension is distributed by flow through the entire thickness of the lower crust,

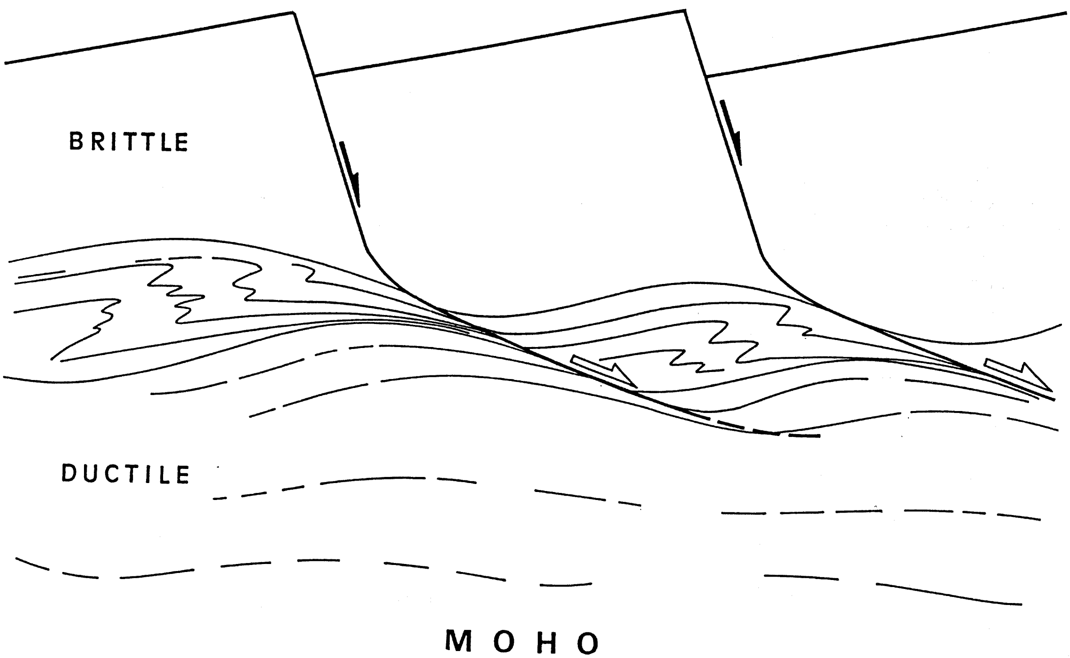

Fig. 11. Synthetic cross-section of a thickened continental crust subjected to extension. See explanation in text.

and consequently that there are no large shear zones cutting through it.

This geometry probably applies for many collapsing orogens and some of its features might be applicable to the formation of passive margins from normal continental crust. But not all of them: there are two fundamental differences between a localized rift set in a normal crust, and a basin- and range-like structure set on a overthickened crust. 1) The thermal structures are different: the lower part of the thick crust is warmer and the thickness of its ductile layer is greater, thus reducing the overall resistance of the crust. This leads to a distribution of extension over large areas (the entire area previously occupied by the mountain belt) and to a deformation mechanism dominated by the ductile behaviour. 2) Ancient mountain belts are characterized by large planar structures, the early thrust planes, which will localize strain during subsequent extension. The example of Corsica and Aegea shows that the asymmetry of extension is the result of the early anisotropy of the reworked nappe stack. Because of these differences the use of collapsed mountain belts to infer conclusions on the geometry of extension in other contexts must then be done with caution. They however remain the only places where deep extensional structures can be directly observed on the field.

REFERENCES

AMAUDRIC DU CHAFFAUT, S. (1982): Les unités alpines à la marge orientale du massif cristallin corse, *Travaux du laboratoire de géologie*, **15** (Presses de l'École Normale Supérieure, Paris).

ANDERSEN, T.B. and B. JAMVEIT (1990): Uplift of deep crust during orogenic extensional collapse: a model based on field studies in the Sogn-Sunnfjord region of Western Norway, *Tectonics*, **9**, 1097-1111.

ANGELIER, J., N. LYBERIS, X. LE PICHON, E. BARRIER and P. HUCHON (1982): The tectonic development of the Hellenic arc and the sea of Crete, a synthesis, *Tectonophysics*, **86**, 159-196.

AVIGAD, D. and Z. GARFUNKEL (1991): Uplift and exhumation of high-pressure metamorphic terranes: the example of Cycladic blueschists belt (Aegean Sea), *Tectonophysics*, **188**, 357-372.

BLAKE, M.C., M. BONNEAU, J. GEYSSANT, J.R. KIENAST, C. LEPVRIER, H. MALUSKI and D. PAPANIKOLAOU (1981): A geogical reconnaissance of the Cyclacic blueschist belt, Greece, *Bull.Geol.Soc.Am.*, **92**, 247-254.

BONNEAU, M. (1984): Correlation of the Hellenide nappes in the southeast Aegean and their tectonic reconstruction, in *The Geologic Evolution of the Eastern Mediterranean*, edited by J.E. DIXON and A.H.F. ROBERTSO (Geol. Soc. Spec. Publ., Blackwell, London), Vol. **17**, pp. 517-527.

BONNEAU, M. and J.R. KIENAST (1982): Subduction, collision et schistes bleus: exemple de l'Egée, Grèce, *Bull.Soc.Geol.Fr.*, **7**, 785-791.

BOILLOT, B., M. RECQ, E.L. WINTERER, A.W. MEYER, J. APPLEGATE, M. BALTUCK, J.A. BERGEN, M.C. COMAS, T.A. DAVIES, K. DUNHAM, C.A. EVANS, J. GIRARDEAU, G. GOLDBERG, J. HAGGERTY, L.F. JANSA, J.A. JOHNSON, J. KASAHARA, J.P. LOREAU, E. LUNA-SIERRA, M. MOULLADE, J. OGG, M. SARTI, J. THUROW and M. WILLIAMSON (1987): Tectonic denudation of the upper mantle along passive margins: a model based on drilling results (ODP leg 103, western Galicia margin, Spain), *Tectonophysics*, **132**, 335-342.

BOUILLIN, J.P. (1983): Examples de déformations locales liées à la mise en place de granitoides alpins dans des conditions distensives: l'ile d'Elbe (Italie) et le Cap Bougaroun (Algérie), Rev. Géol. Dyn. Gèogr. Phys., 24, 101-116.

BREWER, J.A. and D.K. SMYTHE (1984): MOIST and the continuity of crustal reflector geometry along the Caledonian-Appalachian orogen, *J. Geol. Soc., London*, **141**, 105-120.

BUICK, I.S. (1991): Mylonite fabric development on Naxos, Greece, *J. Struct. Geol.*, **13**, 643-655.

CARMIGNANI, L. and R. KLIGFIELD (1990): Crustal extension in the northern Apennines: the transition from compression to extension in the Alpi Apuane core complex, *Tectonics*, **9**, 1275-1305.

CARON, J.M. (1977): Lithostratigraphie et tectonique des schistes lustrés dans les Alpes cottiennes septentrionales et en Corse orientale, Thése de doctorat d'état, Sc. Geol. Mém. n° **48**, Strasbourg.

CARON, J.M., J.R. KIENAST and C. TRIBOULET (1981): High pressure-low temperature metamorphism and polyphase Alpine deformation at Sant'Andrea di Cotone (Eastern Corsica, France), *Tectonophysics*, **78**, 419-451.

CARON, J.M. and G. PEQUIGNOT (1986): The transition between blueschist and lawsonite-bearing eclogites based on observations from Corsica metabasalts, *Lithos*, **19**, 205-218.

CHEADLE, M.J., S. MCGEARY, S. WARNER and M.R. MATTHEWS (1987): Extensional structures on the western UK continental shelf: a review of evidence from deep seismic profiling, *Geol. Soc. London, Spec. Pub.*, **28**, 445- 466.

DAVY, PH. and PH. GILLET (1986): The stacking of thrust slices in collision zones and its thermal consequences, *Tectonics*, **5**, 913-930.

FAURE, M. and M. BONNEAU (1988): Données nouvelles sur l'extension néogène de l'Egée: la déformation ductile du granite miocéne de Mykonos (Cyclades, Grèce), *C.R. Acad. Sci. Paris*, **307**, 1553-1559.

FAURE, M., M. BONNEAU and J. PONS (1991): Ductile deformation and syntectonic granite emplacement during the late Miocene extension of the Aegea (Greece), *Bull. Soc. Geol. Fr.*, **162**, 3-12.

FOURNIER, M., L. JOLIVET, B. GOFFÈ and R. DUBOIS (1991): The Alpine Corsica metamorphic core complex, *Tectonics*, **10**, 1173-1186.

GAUTIER, P., M. BALLÈVRE, J.P. BRUN and L. JOLIVET (1990):

Extension ductile et bassins sédimentaires Mio-Pliocène dans les Cyclades (îles de Naxos et de Paros), *C.R. Acad. Sci. Paris*, **310**, 147-153.

HIRN, A. and M. SAPIN (1976): La croute terrestre sous la Corse: données sismiques, *Bull. Soc. Geol. Fr.*, **18**, 1195-1199.

JACKSON, J.A., G. KING and C. VITA-FINZI (1982): The neotectonics of the Aegean: an alternative view, *Earth Planet. Sci. Lett.*, **61**, 303-318.

JACKSON, J.A. and N.J. WHITE (1989): Normal faulting in the upper continental crust: observations from regions of active extension, *J. Struct. Geol.*, **11**, 15-36.

JOLIVET, L., R. DUBOIS, M. FOURNIER, B. GOFFÉ, A. MICHARD and C. JOURDAN (1990): Ductile extension in Alpine Corsica, *Geology*, **18**, 1007-1010.

JOLIVET, L., J.M. DANIEL and M. FOURNIER (1991): Geometry and kinematics of ductile extension in Alpine Corsica, *Earth. Planet. Sci. Lett.*, **104**, 278-291.

KASTENS, K., J. MASCLE, C. AUROX, E. BONATTI, C. BROGLIA, J. CHANNELL, P. CURZI, K.C. EMEIS, G. GLAÇON, S. HASEGAWA, W. HIEKE, G. MASCLE, F. MCCOY, J. MCKENZIE, J. MENDELSON, C. MÜLLER, J.P. RÉHAULT, A. ROBERTSON, R. SARTORI, R. SPROVIERI and M. TORII (1988): ODP Leg 107 in the Tyrrhenian Sea: insight into passive margin and back-arc basin evolution, *Geol. Soc. Am. Bull.*, **100**, 1140-1156.

KASTENS, K. and J. MASCLE (1990): The geological evolution of the Tyrrhenian Sea: an introduction to the scientific results of ODP Leg 107, in KASTENS, K.A., J. MASCLE *et al., Proc. ODP, Scientific Results*, **107**, 3- 26.

KING, G., Z. OUYANG, P. PAPADIMITRIOU, A. DESCHAMPS, A. GAGNEPAIN, G. HOUSEMAN, J. JACKSON, C. SOUFLERIS and J. VIRIEUX (1985): The evolution of the Gulf of Corinth (Greece) an aftershock study of the 1981 earthquake, *Geophys. J. R. Astron. Soc.*, **80**, 677-693.

KISSEL, C., C. LAY, A. POISSON, Y. SAVASCIN, K. SIMEAKIS and J.L. MERCIER (1986): Paleomagnetic evidence for neogene rotational deformations in the Aegean domain, *Tectonics*, **5**, 783-795.

LAHONDÈRE, D. (1988): Le métamorphisme éclogitique dans les orthogneiss et les métabasites ophiolitique de la région de Farinole (Corse), *Bull. Soc. Geol. Fr.*, **4**, 579-586.

LAY, C., M. JAMET, D. SOREL and J.P. VALENTE (1982): First paleomagnetic results from Mio-Pliocene series of the Hellenic sedimentary arc, *Tectonophysics*, **86**, 45-67.

LE PICHON, X. (1981): Land-locked oceanic basins and continental collision, the eastern Mediterranean as a case example?, *Symposium on Mountain Building, Zurich, July 1981*, pp. 201-211.

LE PICHON, X. and J. ANGELIER (1979): The Hellenic arc and trench system: a key to the neotectonic evolution of the eastern Mediterranean area, *Tectonophysics*, **60**, 1-42.

LISTER, G.S., G. BANGA and A. FEENSTRA (1984): Metamorphic core complexes of cordilleran type in the Cyclades, Aegean Sea, Greece, *Geology*, **12**, 221-225.

LISTER, G.S. and G.A. DAVIS (1989): The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region U.S.A., *J. Struct. Geol.*, **11**, 65-94.

MAKRIS, J. (1978): The crust and upper mantle of the Aegean region from deep seismic sounding, *Tectonophysics*, **46**, 269-284.

MASCLE, J. and L. MARTIN (1990): Shallow structure and recent evolution of the Aegean Sea: a synthesis based on continuous reflection profiles, Marine Geology, **94**, 271-299.

MASCLE, J. and J.P. RÉHAULT (1990): A revised seismic stratigraphy of the Tyrrhenian Sea: implications for the Basin evolution, in KASTENS, K.A., J. MASCLE *et al., Proc. ODP, Scientific Results*, **107**, 617-636.

MATTAUER, M., M. FAURE and J. MALAVIELLE (1981): Transverse lineation and large scale structures related to Alpine obduction in Corsica, *J. Struct. Geol.*, **3**, 401-409.

MCKENZIE, D. (1972): Active tectonics in the Mediterranean region, *Geophys. J.R. Astron. Soc.*, **30**, 109-185.

MERCIER, J.L., D. SOREL and K. SIMEAKIS (1987): Changes in the state of stress in the overriding plate of a subduction zone: the Aegean Arc from the Pliocene to the Present, *Ann. Tect.*, **1**, 20-39.

MORELLI, C., P. GIESE, A. HIRN, B. COLOMBI, C. EVA, I. GUERRA, H. LETZ, R. NICOLIC, C. REICHERT, S. SCARASCIA and P. WIGGER (1976): Seismic investigations of crustal and upper mantle structure of the northern Apennines and Corsica, in BIJU-DUVAL, B. and L. MONTADERT, *Structural History of the Mediteranean Basins* (Technip, Paris), pp. 281-286.

PAPANIKOLAOU, M., V. LYKOUSIS, G. CRONIS and P. PAVLAKIS (1988): A comparative study of neotectonic basins across the Hellenic arc: the Messiniakos, Argolikos, Saronikos and Southern Evoikos gulfs, *Basin Research,* **1**, 167-176.

PLATT, J.P. (1986): Dynamics of orogenic wedges and the uplift of high pressure metamorphic rocks, *Geol. Soc. Am. Bull.*, **87**, 1037-1053.

ROBERT, E. (1982): Contribution à l'étude géologique des Cyclades (Grèce): l'île de Paros, Thesis, Université Paris Sud.

SARTORI, R. (1990): The main results of ODP Leg 107 in the frame of neogene to Recent geology of periTyrrhenian areas, in KASTENS, K.A., J. MASCLE *et al., Proc. ODP Scientific Results*, **107**, 715-730.

SERANNE, M. and M. SÉGURET (1987): The Devonian basins of Western Norway: tectonics and kinematics of an extending crust, *Geol. Soc. London, Spec. Publ.*, **28**, 537-550.

TAYMAZ, T., J. JACKSON and D. MCKENZIE (1991): Active tectonics of the north and central Aegean Sea, *Geophys. J. Int.*, **106**, 433-490.

URAI, J. L., J.B.H. JANSEN and R.D. SHUILING (1989): Deformation in a middle crustal extensional shear zone: An example for Naxos (Greece), *Terra Abstr., EUG V, Strasbourg*, **1**, 380.

WARBURTON, J. (1986): The ophiolite-bearing Schistes Lustrés nappe in Alpina Corsica: a model for the emplacement of ophiolites that have suffered HP/LT metamorphism, *Geol. Soc. Am. Mem.*, **164**, 313-331.

WATERS, C.N. (1990): The cenozoic tectonic evolution of Alpine Corsica, *J. Geol. Soc., London*, **147**, 811-824.

WERNIKE, B. (1981): Low-angle normal faults in the Basin and Range province: nappe tectonics in an extending orogen, *Nature*, **291**, 645-648.

WERNIKE, B. (1985): Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci.*, **22**, 108-125.

WIJBRANS, J.R. and I. MCDOUGALL (1988): Metamorphic evolution of the Attic-Cycladic metamorphic belt on Naxos (Cyclades, Greece) utilizing $^{40}Ar/^{39}Ar$ age spectrum measurements, *J. Metamorphic Geol.*, **6**, 571-594.