Initiation, geometry and mechanics of brittle faulting in exhuming metamorphic rocks: insights from the northern Cycladic islands (Aegean, Greece)

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Abstract. – Initiation, geometry and mechanics of brittle faulting in exhuming metamorphic rocks are discussed on the basis of a synthesis of field observations and tectonic studies carried out over the last decade in the northern Cycladic islands. The investigated rocks have been exhumed in metamorphic domes partly thanks to extensional detachments that can be nicely observed in Andros, Tinos and Mykonos. The ductile to brittle transition of the rocks from the footwall of the detachments during Aegean post-orogenic extension was accompanied by the development of asymmetric sets of meso-scale low-angle normal faults (LANFs) depending on the distance to the detachments and the degree of strain localization, then by conjugate sets of high-angle normal faults. This suggests that rocks became progressively stiffer and isotropic and deformation more and more coaxial during exhumation and localization of regional shearing onto the more brittle detachments. Most low-angle normal faults result from the reactivation of precursory ductile or semi-brittle shear zones; like their precursors, they often initiate between or at the tips of boudins of metabasites or marbles embedded within weaker metapelites, emphasizing the role of boudinage as an efficient localizing factor. Some LANFs are however newly formed, which questions the underlying mechanics, and more generally rupture mechanisms in anisotropic rocks. The kinematics and the mechanics of the brittle detachments are also discussed in the light of recent field and modeling studies, with reference to the significance of paleostress reconstructions in anisotropic metamorphic rocks.

Initiation, géométrie et mécanique des failles dans des roches métamorphiques en cours d'exhumation (Cyclades septentrionales, Egée, Grèce)

Mots-clés. – Failles, Transition ductile-cassant, Détachements, Localisation de la déformation, Exhumation, Cyclades, Egée.

Résumé. – L'initiation, la géométrie et la mécanique des failles dans des roches métamorphiques en cours d'exhumation est discutée à la faveur d'une synthèse des données de terrain et d'études tectoniques conduites depuis une dizaine d'années dans les Cyclades septentrionales. Les roches étudiées ont en partie été exhumées dans des dômes métamorphiques via le jeu de détachements extensifs que l'on peut observer dans les îles de Tinos, Andros et Mykonos. Le passage de la transition ductile-cassant des roches dans le compartiment inférieur des détachements s'est accompagné de l'apparition, selon la distance à celui-ci et le degré de localisation de la déformation, de systèmes asymétriques de failles normales à faible pendage puis de systèmes conjugués de failles normales à fort pendage, suggérant un comportement de plus en plus cassant et isotrope des roches et une déformation de plus en plus coaxiale au fur et à mesure de l'exhumation et de la localisation de la déformation cisaillante régionale sur le détachement cassant. La plupart des failles normales à faible pendage proviennent de la réactivation de zones de cisaillement ductiles à semi-cassantes précurseurs, et s'initient souvent comme celles-ci entre les, ou aux extrémités des, boudins de métabasites ou de marbres intercalés dans les métapélites, ce qui atteste d'un rôle important du boudinage comme facteur de localisation. Certaines de ces failles sont néanmoins néoformées, ce qui pose la question de la mécanique à leur origine, et plus généralement de la rupture dans les milieux anisotropes. A plus grande échelle, la cinématique et la mécanique des détachements sont également discutées à la lumière des travaux récents de terrain et de modélisation, tout comme la signification des reconstructions de paléocontraintes dans les roches métamorphiques anisotropes.

INTRODUCTION

Describing the way brittle faults initiate and develop at different scales in the crust is of key importance to understand the mechanical characteristics and the rheological properties of both crustal rocks and faults [e.g., Reches and Lockner, 1994]. On a regional point of view, studying fault initiation and propagation provides insights into regional fault

kinematics and strain localization [e.g, Cowie *et al.*, 1995; Knott *et al.*, 1996].

There are different ways of studying in the field the initiation of brittle faulting in a material previously devoid of faults. One possible way consists in examining fault initiation in previously unfaulted sedimentary or volcanic rocks in the upper crust. Crider and Peacock [2004] provided a synthesis of observations of meso-scale brittle faults and emphasized the dominant styles of brittle fault initiation in rocks deformed at or near the Earth's surface. Focusing on faults with small amounts of slip because they presumably illustrate faults in their early stages, they studied the termination zones in order to determine the styles of fault initiation. They used space as a proxy for time since structures at and around the fault tips are presumed to represent the earliest stages of fault development, and structures behind the tips, toward the centre of the fault, are presumed to represent later stages. They recognize three styles of fault initiation: initiation from pre-existing structures (formed during an earlier event; e.g., joints), initiation with precursory structures (formed earlier in the same deformation event; e.g., joints, veins, solution seams, shear zones), or initiation as continuous shear zones. A common scenario involves fault initiation by shear along pre-existing or precursory structures, which become linked by differently orientated structures, as stresses are perturbed within the developing fault zone; a through-going fault finally develops.

An alternative way of studying brittle fault initiation consists of examining how brittle tectonic structures initiate and further develop during exhumation of rocks which previously suffered ductile deformation and metamorphism [e.g., Tricart *et al.*, 2004; Mehl *et al.*, 2005]. Rocks passing through the ductile-brittle transition during their way back to the surface record, and therefore potentially document, the initial localization of brittle deformation in a previously ductile material that exhibits foliation and ductile shear zones but devoid of true brittle pre-existing discontinuities. This allows the description of the succession of events that ultimately lead to localization and development of brittle faults. Provided that the kinematics of the system does not change during rock exhumation, one can thus take advantage of the fact that (micro)structures evolve in type while the regional structure enters the brittle domain, for instance during syn-exhumation cooling.

Geodynamic settings where post-orogenic extension has taken place are particularly suitable environments for studying such an initiation of brittle faulting which is progressively superimposed onto ductile and semi-brittle shearing during exhumation and continuous extension. Our observations come from the Aegean extensional metamorphic domes of Tinos, Andros and Mykonos islands. Tinos, Andros and Mykonos are being situated in the northern part of Cyclades (fig. 1A), in the back-arc region of the Hellenic subduction, where crustal thinning has been active since the Oligocene-Early Miocene [Le Pichon and Angelier, 1981; Jolivet and Faccenna, 2000] leading to the formation of the Aegean sea. From Crete to the northern Cyclades, extension was achieved in the Neogene by shallow north-dipping faults and shear zones with a consistent top-to-the-north (or northeast) sense of shear [e.g., Faure *et al.*, 1991; Lee and Lister, 1992; Fassoulas *et al.*, 1994; Jolivet *et al.*, 1994, 1996]. In contrast, recent structural studies in the West Cycladic Islands have documented a rather S or SW

directed kinematics [Serifos: Grasemann and Petrakakis, 2007; Iglseder *et al.*, 2009; Tschegg and Grasemann, 2009; Brichau *et al.*, 2010; Kea: Iglseder *et al.*, 2011; Kythnos: Grasemann *et al.*, 2012].

In Tinos, Andros and Mykonos, a detailed description of the succession of small-scale structures from extensional ductile shear zones to normal faults in the hangingwall and footwall of extensional detachments has been carried out and a clear continuum of strain from ductile to brittle of footwall rocks has been documented [Lee and Lister, 1992; Jolivet and Patriat, 1999; Mehl *et al.*, 2005, 2007; Lecomte *et al.*, 2010]. These studies further allowed constraining both the kinematics of the detachment and the rheological behavior of the extending Aegean continental crust.

In this paper, we aim first at providing a synthesis of field observations on the geometry of meso/macro-scale brittle normal faults and the way they initiated and propagated within the metamorphic rocks exposed on the northern Cycladic islands of Tinos, Andros and Mykonos, relying on previous studies by Mehl *et al.* [2005, 2007] and Lecomte *et al.* [2010] for more complete regional descriptions. We focus on extensional tectonics without considering strike-slip faults (nor reverse faults) that were also recognized in places [Boronkay and Doutsos, 1994; Angelier, 1979b; Doutsos and Kokkalas, 2001; Menant *et al.*, 2012]. Second, we summarize the recent advances in understanding the kinematics and mechanics on low-angle normal faulting (e.g., slip along LANFs in an Andersonian extensional stress field; initiation, reactivation and slip mechanisms of LANFs; fault zone weakening), as well as the still open questions concerning rupture in metamorphic rocks. Finally, we aim at briefly discussing the reliability and significance of paleo-stress reconstructions that were increasingly carried out in anisotropic metamorphic material during the recent years in order to decipher brittle faulting events during exhumation of orogenic hinterlands [e.g., Agard *et al.*, 2003; Tricart *et al.*, 2004; Mehl *et al.*, 2005, 2007].

TECTONIC SETTING OF TINOS, ANDROS AND MYKONOS ISLANDS

The Aegean sea (fig. 1A) is one of the Mediterranean back-arc basins, formed from the late Oligocene to the present above the subduction of the African slab beneath the southern margin of Eurasia [Le Pichon and Angelier, 1981; Jolivet and Faccenna, 2000; see also the syntheses by Ring *et al.*, 2010 and Jolivet and Brun, 2010 and references therein]. Extension started ~ 30 Ma and affected the whole Aegean domain; it is presently localized around the Aegean sea in west Turkey, in the Peloponnese, around the gulf of Corinth, and in Crete [Seyitoglu and Scott, 1996; Armijo *et al.*, 1992; Rigo *et al.*, 1996]. The extended domain was once occupied by the Hellenides-Taurides collision belt that was continuous from Greece to Turkey [e.g., Aubouin and Dercourt, 1965]. Continental Greece and the Peloponnese are dissected into several crustal blocks separated by basins [Papanikolaou *et al.*, 1988] (fig. 1A). Andros, Tinos and Mykonos (fig.1B) belong to the same block as Evia.

Post-orogenic extension has brought to the surface rocks which first underwent HP-LT metamorphism during subduction and late greenschist retrogression at lower to mid-crustal levels during exhumation [Avigad *et al.*, 1997].

The final exhumation results from thinning and tectonic denudation taking place in ductile then in brittle conditions within the metamorphic domes. In the Aegean, these domes are composed of two tectonic units separated by shallow-dipping normal detachments. One of the best exposed extensional shear zones of the Aegean sea crops out on Tinos and Andros islands [Avigad and Garfunkel, 1989; Gautier and Brun, 1994; Jolivet and Patriat, 1999; Mehl et al., 2005, 2007]. The shear zone separates an upper plate devoid of HP-LT metamorphism and made of ophiolitic material, including serpentinites and gabbros, as well as amphibolites [Melidonis, 1980; Katzir et al., 1996] from a HP-LT metamorphic lower plate made of alternating metabasites,

marbles, and less competent metapelites [Bröcker, 1990; Parra et al., 2002]. In Tinos, radiometric ages suggest a late Cretaceous to Eocene age for the HP event [Bröcker et al., 1993; Bröcker and Franz, 1998]. The contact itself is a shallow-dipping normal fault marked by a thick breccia/cataclasites zone. On the NE sides of Tinos and Andros, the contact is associated with an intense greenschist retrogression of the HP-LT parageneses in the lower plate. The direction of greenschist stretching, deduced from the attitude of stretching lineation, the projection of lineation on shear planes, and the axes of sheath folds is consistently oriented NE-SW (fig.1B). Later brittle extension, marked by joints, veins and normal faults, also shows a consistent NE-SW
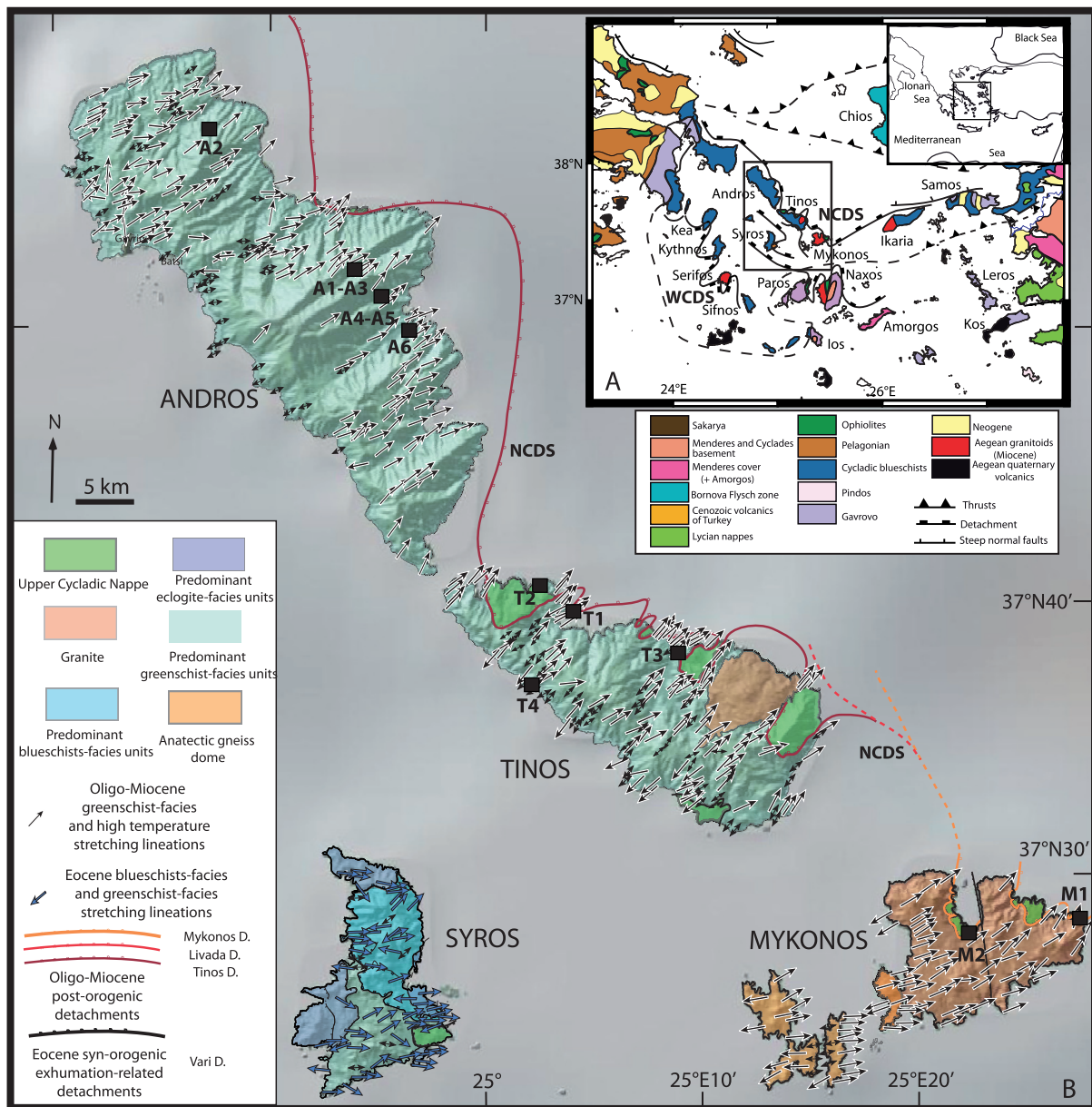
FIG. 1. – Tectonic map of Andros, Tinos and Mykonos islands showing the main lithological units, the detachments and the Oligo-Miocene and Eocene stretching lineations [modified after Jolivet et al., 2010].
Inset: Tectonic map and main geological units of the Aegean region. NCDS: North Cycladic detachment system; WCDS: West Cycladic detachment system. Black squares (A: Andros, T: Tinos, M: Mykonos) correspond to groups of nearby sites from which measurements/observations were reported; for each group average geographical coordinates (lat/lon) are given: A1-A3: 37°51'55"/24°54'36"; A2: 37°57'14"/24°47'49"; A4-A5: 37°50'57"/24°55'30"; A6: 37°49'46"/24°56'38"; T1: Planitis island: 37°39'34"/25°03'54"; T2: 37°40'25"/25°02'20"; T3: Kolimbithra: 37°37'42"/25°08'38"; T4: Ormos Isternia: 37°36'48"/25°02'34"; M1: Cape Evros: 37°28'24"/25°27'34"; M2: Panormos Bay: 37°27'51"/25°22'23"

trend over the whole island [Gautier and Brun, 1994; Jolivet and Patriat, 1999; Mehl *et al.*, 2005, 2007], suggesting that the same direction of extension has persisted throughout the history of extension and exhumation. Kinematic indicators indicate shearing with a sense of shear being almost exclusively top-to-the-northeast in the NE parts of the islands, while the SE part shows a mixture of top-to-the-NE and top-to-the-SW senses of shear, or even evidence of coaxial strain; this supports an increasing non-coaxial (shear) strain toward the main detachment. Retrogression of the HP parageneses increase as well when approaching the detachment.

Mykonos island is mostly made of a monzogranite dated at around 10-13 Ma [e.g., Brichau *et al.*, 2008] (fig. 1B). The granite is a kilometer-scale laccolith intruded into micaschists at the top of migmatitic gneisses belonging to the Cycladic basement. The laccolith constitutes the core of an extensional gneiss dome and shows an intense mylonitisation when approaching the detachment surface [Denèle *et al.*, 2011].

The similarities between the different detachments and the evolution in space and time of the localization of the main movement zone has led Jolivet *et al.* [2010] to propose that all these detachements are part of a single crustal-scale detachment, the North Cycladic detachment system (NCDS, fig. 1A) that reworked the Hellenic accretionary wedge from the Late Oligocene to the Late Miocene. Grasemann *et al.* [2012] similarly proposed that the S-directed kinematics of the low-angle normal faults identified on the islands of Kea, Serifos and Kythnos are mechanically linked and form the West Cycladic detachment system (WCDS, fig. 1A). The relations between the NCDS and the WCDS are not clear, but both systems together accommodated Miocene extension in the Aegean.

During exhumation the domes of Tinos, Mykonos and Andros that formed below the main detachments entered the brittle-ductile transition and low-angle ductile shear zones continued their activity in the brittle field and evolved progressively into low-angle normal faults, while footwall rocks underwent successively ductile and brittle deformation. Mykonos shows a more evolved system with one branch of the detachment having been active quite close to the surface and having controlled deposition of hanging wall syn-rift clastic sedimentary rocks [Lecomte *et al.*, 2010].

Although this paper mainly focuses on extensional structures it is worthwhile to note that strike-slip and reverse faults have been locally described in the Cyclades [Boronkay and Doutsos, 1994; Angelier, 1979b; Doutsos and Kokkalas, 2001]. They accommodate a minor part of the finite deformation. Most strike-slip and reverse faults formed in the Late Miocene, as a possible consequence of the westward motion of the Anatolian block [Menant *et al.*, 2012] and of the overall E-W shortening of the Cyclades [e.g. Avigad *et al.*, 2001].

REVIEW OF FIELD OBSERVATIONS ON MESO-SCALE FAULTS (DISPLACEMENT < 1m)

Because the investigated rocks experienced a continuous change in deformation regime from ductile to brittle, it is worthwhile to recall the nomenclature related to the markers of continuous versus discontinuous displacement along a shear zone [Fusseis *et al.*, 2006]. The brittle to ductile

transition (BDT) (or plastic to viscous transition) is the change from fracturing on one or more discrete surfaces to thermally activated creep within zones of viscous, solid-state flow [e.g., Schmid and Handy, 1991; Fusseis *et al.*, 2006]. A brittle fault is understood hereinafter as a single (but also possibly more complex, composite) surface/zone of deformation across which discontinuous displacement occurred, surrounded by a deformed volume of wall rock (the damage zone). In the field, criteria used to infer a discontinuous slip are slickensides, (calcite) steps and cm to m-scale offsets of lithological markers or foliation. Brittle faults differ from ductile and semi-brittle shear zones, which are regions of localized but continuous displacement and continuous increase of finite strain (from zero strain at the margins of the shear zone to maximum strain in its centre). This continuous character of deformation is indicated by the absence of true slickensides and bending (rather than offset) of foliation on both sides of the shear zone, i.e., no geometrical discontinuity can be seen on the scale of the shear zone. Note however that this distinction between brittle and ductile applies at the scale of the outcrop, with offsets along brittle faults < 1 m; ductile deformation recognized as such at this scale may involve brittle deformation of some minerals at the micro-scale (e.g., thin-section), especially within the BDT.

Occurrence of high-angle and low-angle normal faults

Two kinds of normal faults were observed in the field: steeply dipping normal faults, often arranged into conjugate systems, and shallow-dipping normal faults, i.e., with dips lower than 30°. The latter correspond either to meso-scale structures or to the above-mentioned detachments.

Faults forming in necks between boudins

In Tinos and Andros, many semi-brittle shear zones and brittle normal faults are closely associated with lithological heterogeneities and boudins. Figure 2A shows an asymmetric boudin (i.e., shear band boudinage) of metabasites embedded in a metapelitic matrix. The tips of the boudins are marked by the presence of NE-dipping shear zones. The metapelitic matrix is almost devoid of brittle features. Localization of deformation in the metapelites is weak and is only marked by shear zones, while actual brittle deformation concentrates in the metabasites. Brittle structures occur as steeply-dipping normal faults, either connected to ductile shear zones or displaying conjugate sets (fig. 2B, C).

These observations can be made at different scales; figure 3 clearly illustrates that normal faults nucleated within and at the tips of boudins (fig. 3A, B) and propagated into the matrix in the form of *en echelon* arrays of veins (fig. 3B; see section: Initiation of high-angle normal faults).

Low-angle normal faults developing by 'reactivation' of, or extreme localization along, precursory ductile or semi-brittle shear zones

In Tinos, some shallow-dipping faults lie parallel to shear zones (fig. 4, T1), suggesting that the latter have presumably been reactivated in a brittle manner, or have continuously evolved from ductile to brittle shear zones, the latest increment of deformation being clearly discontinuous. Those features are also well expressed in Andros. Figure 5A show ductile shear zones, the steepest of which

having been reactivated as brittle faults: across these planes, foliation is bent and offset, which supports a late discontinuous shear movement. Calcite steps and slickensides (fig. 5B) are sometimes observed; they are parallel to the stretching lineation, indicating a discontinuous late increment of deformation kinematically consistent with ductile stretching. Such structures emphasize the continuum of kinematics between early ductile shearing and late discontinuous brittle slip.

Newly-formed low-angle normal faults

Synthetic of the main detachment (i.e., NE dipping)

At Tinos, at Kolimbithra (fig. 1, T3), below the main Tinos detachment, NE-dipping low-angle normal faults form in metapelites in the absence of any boudin and cut across the ductile foliation (fig. 6A, C, D and E). These shallow-dipping planes clearly initiated and propagated through the metamorphic pile without superposing onto any pre-existing/ precursory structure. Consequently, they did not form by reactivation, but rather as newly-formed faults. They are contemporaneous with vertical veins (fig. 6A, F). These shallow-dipping normal faults are also associated with high-angle normal faults (fig. 6B, F).

Antithetic of the main detachment (i.e., SW dipping)

A second kind of shallow-dipping faults that initiated and moved exclusively in a brittle manner is illustrated by some SW-dipping faults. Their cross-cutting relation with the late crenulation cleavage and folds shows that they are not controlled by any pre-existing/precursory structure such as ductile shear zones, which confirms that they are newly formed. They are observed just below the detachment (fig. 1, T1 and T2), where no SW dipping shear zones are encountered. Some vertical veins cut across shallow-dipping fault planes, some others are crosscut by the fault planes, which suggests that both types of structures are likely coeval [Mehl et al., 2005] (fig. 4).

Although low-angle normal faults are much scarcer in Andros, figure 7 illustrates a similar occurrence. A NE-dipping brittle fault developed along the most steeply dipping pre-existing shear zone. A seemingly conjugate SW dipping low angle normal fault formed, cutting across the foliation and without any obvious relationship to the background pattern of ductile/semi-brittle shear zones. This strongly suggests that this normal fault initiated with a low dip as a newly formed fault, making a high angle to the vertical direction of the regional maximum principal stress.
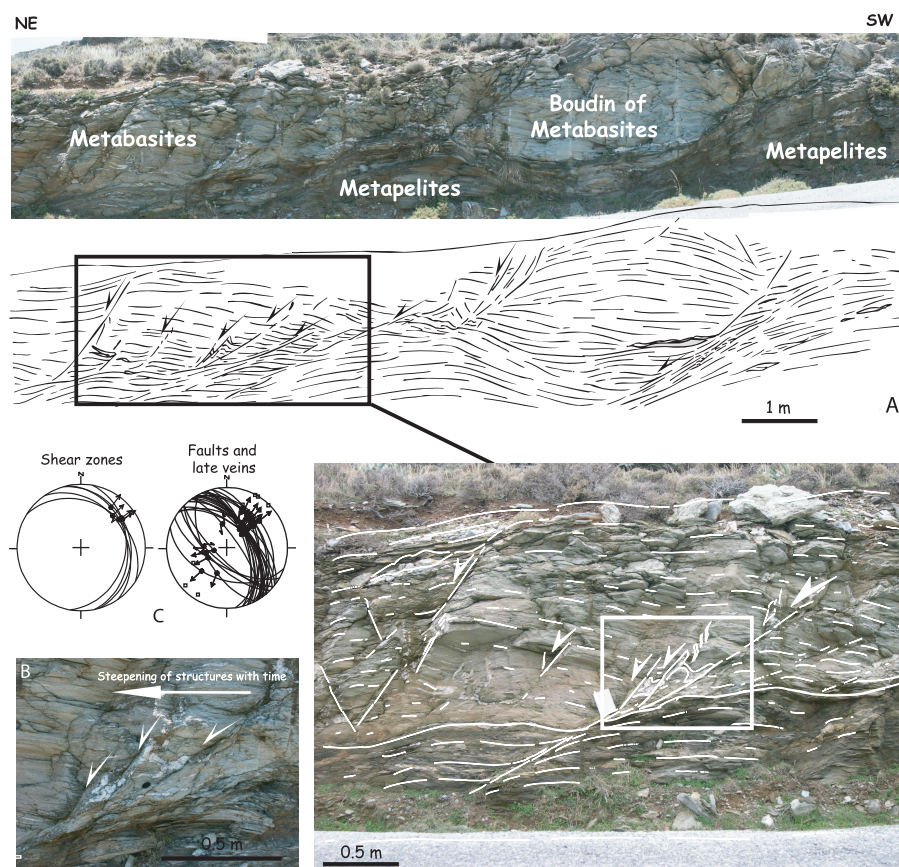
FIG. 2. – A: Evolution of deformation from ductile to brittle around an asymmetric boudin of metabasites embedded in a metapelitic matrix (Andros, A4-A5 in fig.1). The metapelitic matrix is almost devoid of brittle features. Localization of deformation in the metapelites is weak and is marked only by shear zones, whereas actual brittle deformation concentrates in the metabasites between the boudins. B: Example of progressive steepening of structures with increasing brittle behavior. C: Examples of patterns of ductile shear zones and brittle faults (lower hemisphere, equal area projection). Thin curves represent fault/shear planes and dots with arrows indicate striations/projection of the stretching lineation. Small white squares represent poles to veins. Note that the dip angles are higher for late brittle faults than for earlier ductile shear zones, and that deformation becomes more coaxial while evolving from ductile to brittle.

Progressive steepening from ductile shear zones to brittle faults

Numerous outcrops of Tinos and Andros show a succession of progressively steepening shear planes. The early shallow planes are almost parallel to the underlying ductile shear zones (fig. 8). They progressively steepen with increasing brittle behavior, with first a slight bending of the foliation plane on either sides of the semi-brittle shear zone and then a clear offset in a sense compatible with the ductile shear (figs. 2C and 8A, B). The dip angles of the late brittle faults are therefore higher than those of early shear zones (fig. 8D, E).

Initiation of high-angle normal faults

A more evolved pattern consists of conjugate steeply-dipping normal faults, which cut across the entire metamorphic pile (fig. 9).

In Andros, sub-vertical joints and veins are often associated in *en echelon* arrays, first reported by Papanikolaou [1978]. They occur in the more competent layers, i.e. in the boudins of metabasites (fig. 3B) and in quartzitic layers of the metapelitic outcrops (fig. 10). These *en echelon* arrays of veins and joints define rough planes whose orientation, dip and kinematics are comparable and consistent with conjugate sets of normal faults; they reflect the on-going localization of normal fault zones. Sometimes, these *en echelon* structures evolve toward actual through-going normal faults (fig. 2B). This evolution is statistically more common for NE dipping planes.

Newly-formed high-angle normal faults often cut across, hence postdate, the low-angle normal faults (fig. 8C).

REVIEW OF FIELD OBSERVATIONS ON MACRO-SCALE LOW-ANGLE NORMAL FAULTS (LANFS)

Field observations allowed documenting major low-angle brittle fault planes with extensional kinematics.

In Tinos (Planitis, fig. 1, T1), the sharply defined NE shallow-dipping brittle Tinos detachment separates a zone of talc-rich breccias and cataclasites at the base of the metabasites of the Upper Cycladic Unit from the mylonites of the lower unit (fig. 11A).

In Mykonos (Cape Evros, fig. 1, M1), the contact between the granite and the metabasites corresponds to a ductile shallow dipping shear zone, the Livada detachment (fig. 11D). It consists of a thin, folded ultra-mylonite parallel to the granite mylonitic foliation. All kinematic indicators within the mylonites and the ultramylonites show a top-to-the-NE shear sense. Locally, the granite-metabasite contact is reworked by brittle low-angle normal faults either localized on the top of the ultra-mylonitic shear zone (fig. 11D), or cutting through it, indicating a late brittle increment of extensional deformation with the same sense of motion. The top of the metabasites near Cape Evros is cut by a brittle cataclastic detachment, the Mykonos detachment. Within 2 m beneath the detachment, the Upper Cycladic nappe is brecciated, forming cataclastic rocks. The sharp Mykonos detachment separates the metabasites
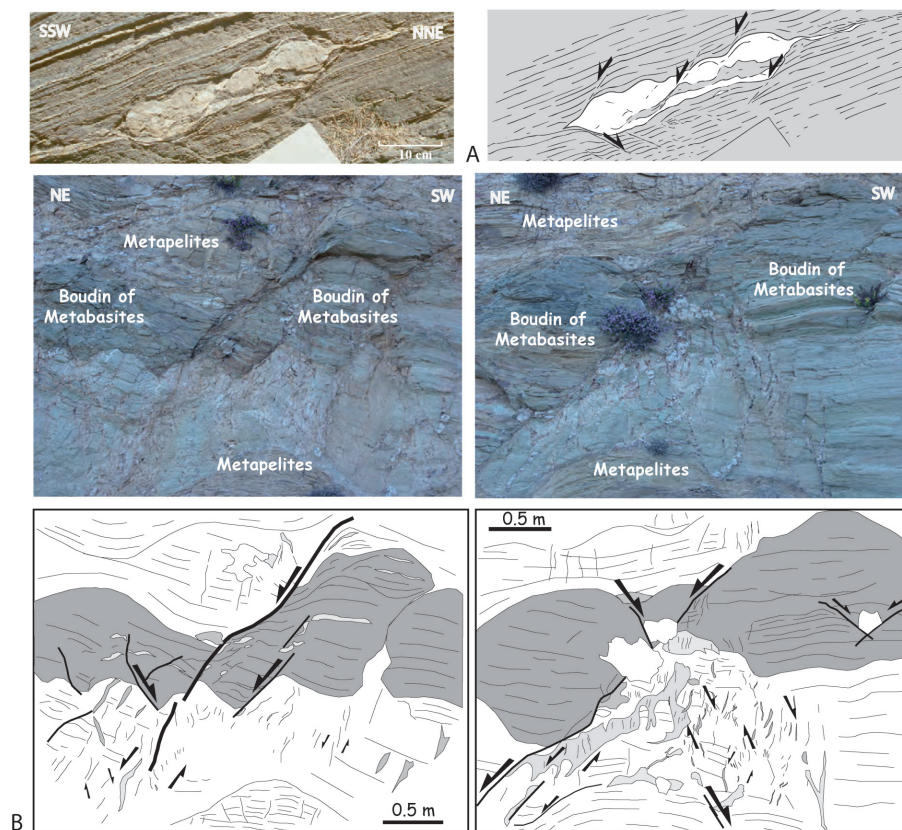

FIG. 3. – Boudinage and initiation of brittle faulting. A: Initiation of brittle faulting within or at the tips of a symmetric boudin (Tinos, Ormos Isternia, T4 in fig.1). B: Initiation of brittle faults between boudins of metabasites and propagation as *en echelon* vein patterns in metapelites (Andros, A1-A3 on fig.1).

from a late Miocene continental syn-rift sedimentary unit (fig. 11C). It is associated with 30 cm thick gouge and high-angle normal faults sole into the detachment. At Panormos Bay (fig. 1, M2), the metabasites are not preserved and the brittle detachment juxtaposes the sedimentary unit directly over the cataclastic Mykonos granite (fig. 11B).

As a result, the major extensional detachments on Tinos, Andros and Mykonos consist of cataclasites or fault gouges that either overprinted or reworked 0.5 to 5 km thick mylonitic ductile shear zones, or that were localized along the margin of these shear zones. A planar fault surface,

formed at shallow crustal levels, usually caps (fig. 11B, C) or is capped by (fig. 11A) the cataclastic rocks. Although the detachments could have worked coevally in relay at different crustal levels [Jolivet *et al.*, 2010; Lecomte *et al.*, 2010], in response to the migration of the brittle-ductile transition, field observations support an overall evolution of mylonitic shear zones toward cataclastic then brittle detachments during exhumation, with increasing localization of shear movement along a progressively thinner fault zone (fig. 12A). The discrete fault surfaces represent the ultimate localization of shear; all show dips lower than 15° (fig. 11).
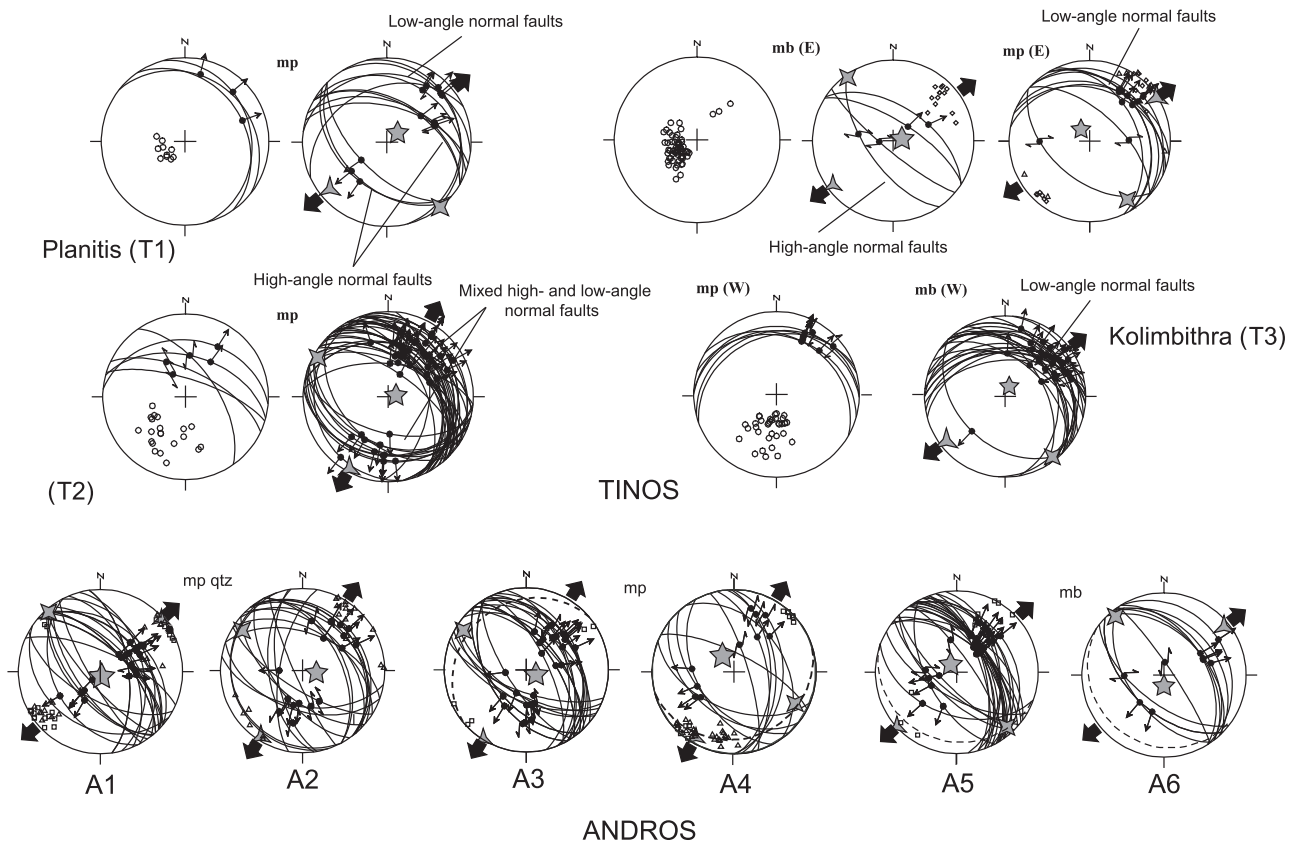
FIG. 4. – Examples of microstructural data illustrating the orientation of ductile shear zones and normal faults. Diagrams: Lower hemisphere equal area projection. Thin curves represent fault/shear planes and dots with arrows indicate striations/projection of the stretching lineation. Foliation planes shown as dashed lines. Small white squares/triangles represent poles to veins/joints. Small white circles represent poles to ductile shear zones. Stars indicate stress axes with five branches (σ1), four branches (σ2) and three branches (σ3) computed using Angelier's [1984, 1990] inversion methods. Divergent large black arrows: direction of horizontal extension.
mp: metapelites; mb: metabasites: mp qtz: quartzitic metapelites. W/E: western/eastern Kolimbithra.
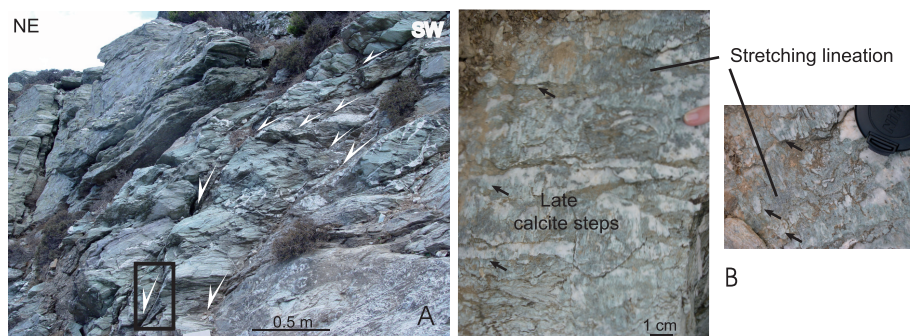
FIG. 5. – A: Example of pattern of NE-dipping ductile shear zones (small white arrows: sense of shear) in Andros (A1-A3 on fig. 1). The steepest ones have been reactivated as normal faults (large white arrows). B: Zoom on a shear plane showing stretching lineation parallel to late striated calcite steps (arrows), indicating a late increment of true brittle slip along precursory shear zones.

FACTORS CONTROLLING THE INITIATION AND GEOMETRY OF MESO-SCALE BRITTLE FAULTS

Lithological control on the initiation and geometry of brittle faults

Shallow-dipping faults are clearly more numerous in poorly competent metapelites than in metabasites, i.e. in weak lithologies (fig. 4). In contrast, faults often steepen in more competent formations such as marbles and metabasites.

A possible explanation for the higher density of low-angle normal faults in metapelites is first related to the development of many of them from precursory shear zones. Precursory ductile and semi-brittle shear zones are more numerous in poorly competent material because in such material the strain rate is higher, the deformation is more penetrative and therefore the spacing of the precursory shear zones, which may be re-used as normal faults, is smaller than in more competent rocks. Furthermore, the feasibility of subsequent brittle reactivation of these precursory shear zones is made possible by the high mica and chlorite content of metapelitic rocks that accounts for a low friction angle that favoured brittle slip at shallow dip.

Concerning those low-angle normal faults that are likely newly-formed, several hypotheses can tentatively be invoked: (1) brittle faults initiated at dip higher than 45º (like in more competent rocks) and then were subsequently tilted by simple shear at the scale of the metapelite body (this hypothesis can be ruled out in most places, see section *Evidence for slip at shallow dip*); (2) although not clearly observed at the outcrop scale, viscous relaxation along chlorite and mica-rich foliation planes, likely efficient close to the BDT, may have induced strain partitioning at the microscale, causing refraction of the segments of the developing fault planes that seemingly display a bulk shallow dip. In that hypothesis, the steepening with time of extensional microstructures in the weak metapelitic lithologies can probably be explained by an increase of viscous relaxation time as rocks were exhumed and cooled: as viscous relaxation diminished, strain partitioning (refraction) became less efficient and the bulk rheology evolved toward that of an isotropic brittle/stiff rock (see above).
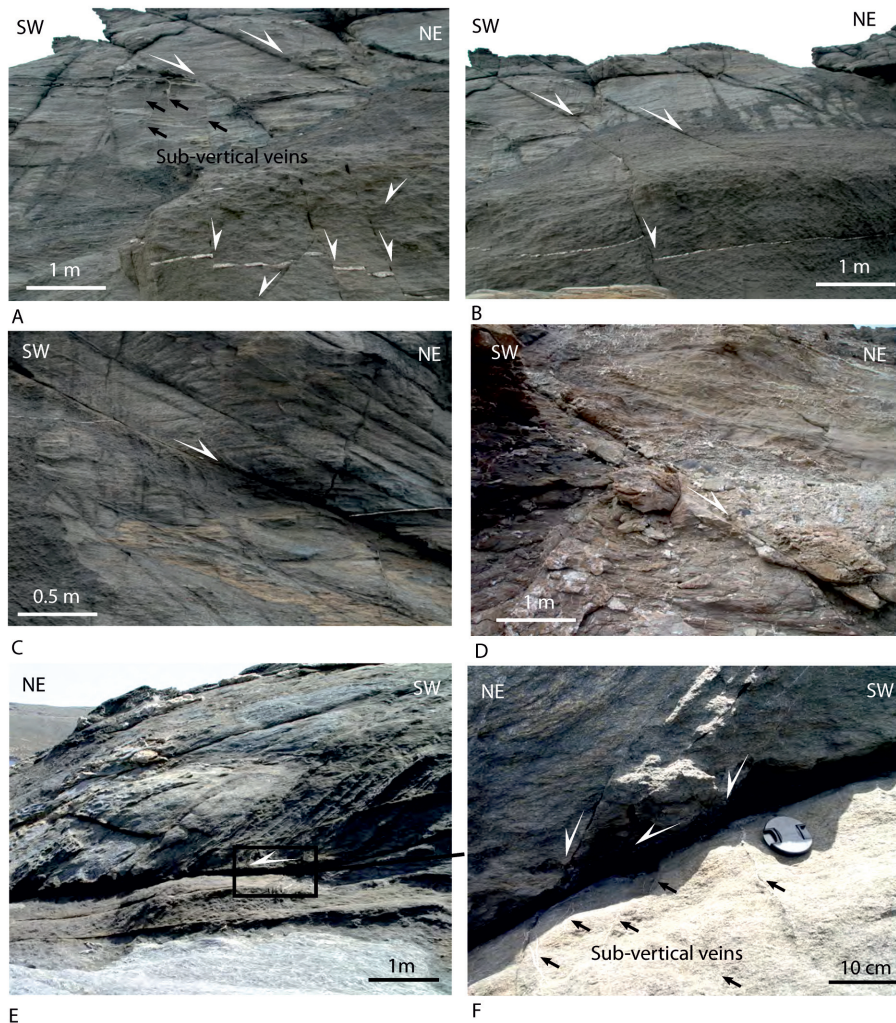
Fig. 6. – Examples of NE-dipping low-angle normal faults (synthetic of the Tinos detachment) at Kolimbithra (Tinos, T3 on fig. 1). A, B, C, E, F: within metapelites, T3, east of the bay; D: within cataclasites, T3, west of the bay. The low-angle normal faults are associated with high-angle normal faults (A, B, F) and with sub-vertical veins (A, F).

Brittle faulting in anisotropic rocks

The structure of the metamorphic material, despite variations of lithologies (marbles, quartzites, metabasites and metapelites mainly) is basically characterized by a set of clearly defined foliation planes, which could have played a dominant role in the mechanical behavior and possibly controlled fault/fracture occurrence and geometry.

Rupture in layered and anisotropic rocks have poorly been addressed in the geological literature. Peacock and

Sanderson [1992] addressed the effects of layering and anisotropy on fault geometry. They showed that in layered sedimentary rocks, the geometry of faults varies with the orientation of layering with respect to the stress field. Where σ1 is nearly normal to layering or anisotropy, conjugate faults develop symmetrically about σ1 like in an isotropic material. Where rocks have interbedded layers with different mechanical behaviors, faults tend to initiate as extension fractures orthogonal to the more brittle layers but oblique to the less brittle layers. Where σ1 is oblique at
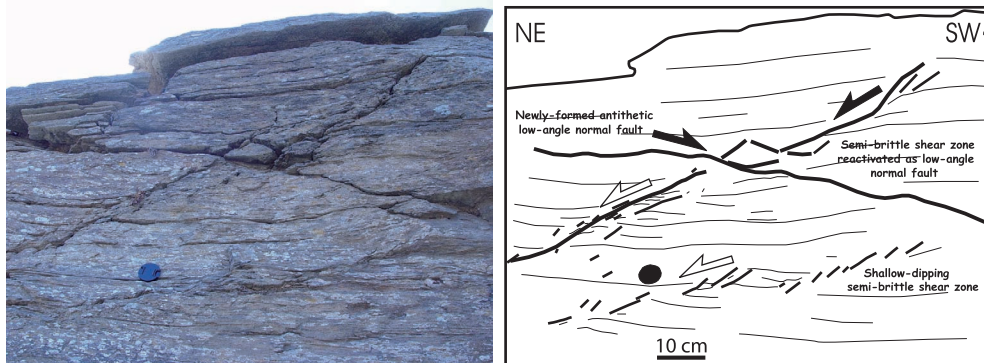
Fig. 7. – Example of pseudo-conjugate low-angle normal fault system in Andros (A4-A5 on fig.1). A NE-dipping low-angle normal fault developed along the steepest precursory shear zone. A "conjugate" SW dipping low-angle normal fault formed, cutting across the foliation and without any obvious relationship to the background pattern of ductile/semi-brittle shear zones. This suggests that this normal fault initiated with a low dip as a newly formed fault.
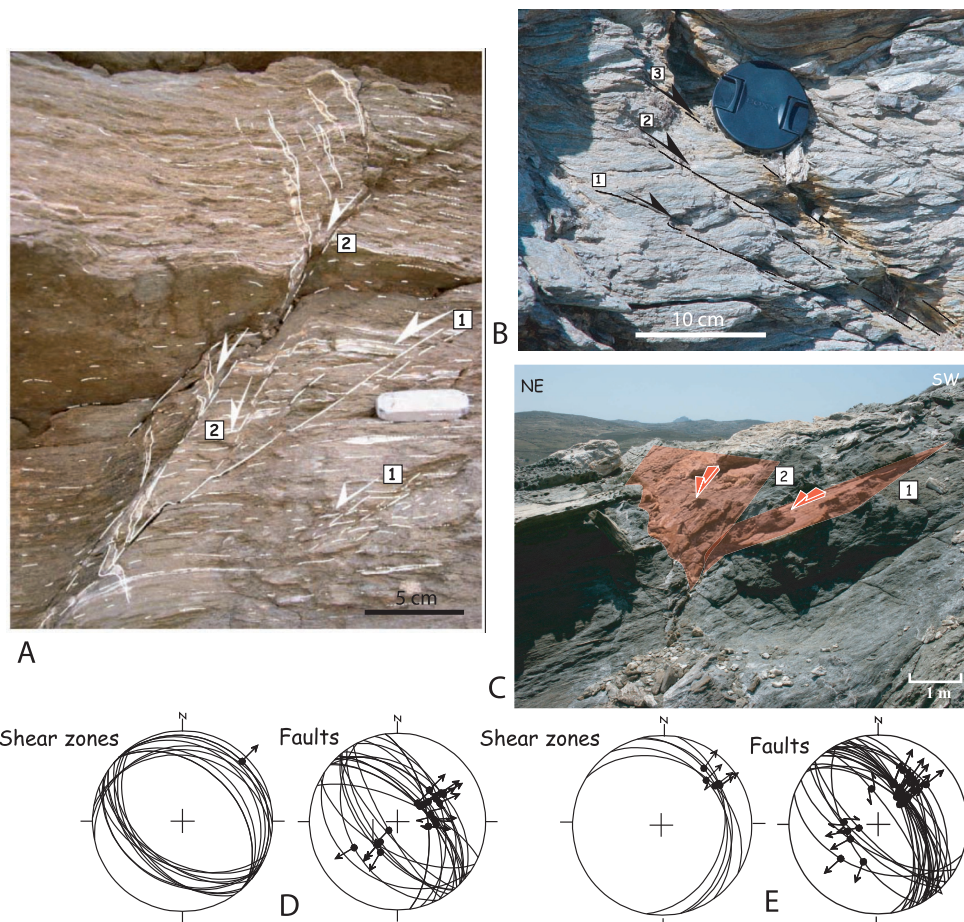
Fig. 8. – A, B: progressive steepening of shear planes from ductile shear zones (1) to brittle faults (2, 3) in Andros (A: A4-A5 on fig.1; B: A1-A3 on fig.1). C: Chronology of brittle faulting: a late high-angle normal fault clearly cuts across a low-angle normal fault (Tinos, Kolimbithra, T3 on fig.1). D, E: microstructural data from Andros illustrating that normal faults have dips steeper than those of ductile shear zones, either more or less symmetric (D, A1 on fig. 1) or asymmetric (E, A5 on fig. 1).

25-75º to anisotropy, one set of faults developed at a high angle to layering with another at a low angle; in that case, large dihedral angles, up to 90º may be observed and σ1 does not bisect this angle.

Rupture in anisotropic rocks has received much more attention for engineering purposes. However, works have mainly focused on the failure criteria of continuously anisotropic materials such as slates rather than material with layering [e.g., Jaeger, 1960; Kwasniewski, 1993; Ramamurthy, 1993; Nasseri *et al.*, 1997; Tien and Kuo, 2001; Tien *et al.*, 2006; Goshtasbi *et al.*, 2006; Saroglou and Tsiambaos, 2008]. In experimentally deformed schist samples for instance, material failure is usually due to sliding along schistosity planes for a large range of loading orientations (e.g., 20 to 70º with respect to the schistosity plane). The sudden increase of strength observed during experiments when the loading orientations evolve to either parallel or perpendicular to the schistosity planes reflects the transition from sliding to strain localization and shearing of the rock matrix, with a clear influence of the confining pressure [e.g., Duveau *et al.*, 1998]. Sliding along weakness planes and matrix shearing are both required to reliably model the brittle strength of anisotropic rocks, at least at the scale of the rock sample.

Considering the extensional kinematics with the maximal principal stress axis σ1 always oblique at large angle to the foliation, whatever the cause of this obliquity (see section *Reliability and significance of paleostress reconstructions in anisotropic metamorphic rocks*), the matrix shearing mode of failure and the occurrence of conjugate normal fault patterns was expectedly favoured in the field, at least at the outcrop scale: accordingly, we did not find any field evidence for reactivation of foliation planes, except in precursory shear zones where foliation is sub-parallel to shear planes,

accounting for the calcite steps parallel to the former stretching lineation observed on some high-angle fault planes reactivating high-angle shear zones (fig. 5B). Even the few strike-slip faults documented in Tinos or Mykonos [Mehl *et al.*, 2005; Lecomte *et al.*, 2010], that formed with horizontal σ1 and σ3 principal stress axes, show typical conjugate fault geometries without any evidence for sliding along foliation planes. At the microscale, however, the influence of the foliation anisotropy cannot be ruled out (viscous relaxation close to the BDT, above section *Lithological control on the initiation and geometry of brittle faults*).

To summarize, the question of the possible control of foliation anisotropy on faulting must be addressed at different scales. At the outcrop scale, it seems at a first glance that normal faulting geometry was more or less independent on the pre-existing foliation mechanical anisotropy, being mainly controlled by the occurrence of precursory shear zones, lithological contrasts and the overall evolution of rock properties (e.g., stiffness) during exhumation. That means that within the metamorphic pile, the main cause of anisotropy was related, at the meso-scale, to either lateral/ vertical contrasts of rheology linked to lithology or to the precursory ductile shear zones (and at the macro-scale to preexisting nappe thrusts), and less to the mechanical anisotropy itself. However, it cannot be excluded that for instance the shallow-dipping attitude of some newly-formed normal faults in metapelites that, at the outcrop (meso-) scale seem to cut across the foliation pattern, in fact results from shearing along such planes of anisotropy at the microscale.

It is worthwhile noting that in contrast to faults caused by shear failure and that develop oblique to the foliation anisotropy, the geometry of late veins formed by (effective) tension failure, that are always oblique at high angle to the foliation, seems to be independent of such anisotropy
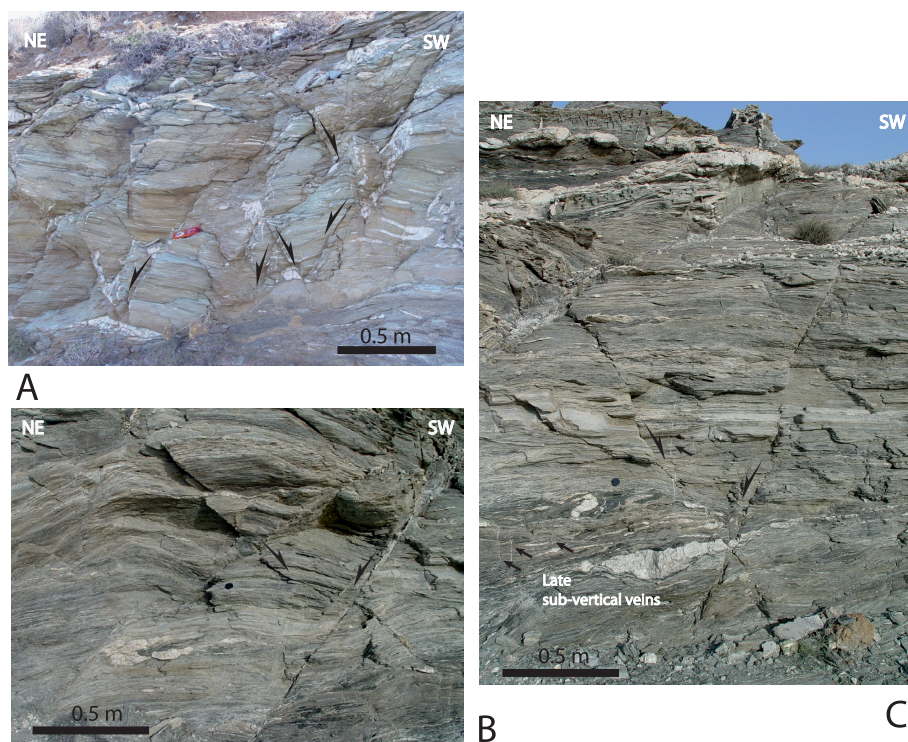

FIG. 9. – Examples of late high-angle normal faults in Andros (A)(A4-A5 on fig. 1) and Tinos (B, C; T2 on fig. 1), associated with sub-vertical veins (C).

(except their development under a σ1 axis possibly reoriented perpendicular to the foliation, see section *Reliability and significance of paleostress reconstructions in anisotropic metamorphic rocks*).

Asymetric vs symetric patterns of normal faults

In Tinos, brittle structures clearly evolve from rather asymmetric with a majority of NE early shallow-dipping planes (fig. 4, T3), toward the more symmetric pattern of the late steeply dipping faults (fig. 4, T1). Only the symmetric patterns of high-angle normal faults are encountered in Andros (fig. 4, A1-A6), which illustrates a less evolved stage of localization of shear strain than Tinos [Mehl *et al.*, 2007]. These fault patterns reflect an evolution of deformation in the footwall of the detachment from non-coaxial stretching in the early stages of deformation to a more coaxial extension during the later stages of exhumation of the footwall; note that a roughly similar evolution is observed with decreasing distance to the detachment.

Reactivation of, or localization along, preferentially NE verging precursor ductile shear zones, explain the larger number of NE-dipping low-angle normal faults, especially when approaching the detachment. The evolution with time and exhumation towards a more symmetric pattern of steeply dipping faults supports that the faulting regime became more coaxial throughout the whole island, including within the previous km-thick mylonitic shear zone and the thinner cataclastic shear zone, while simple shear progressively localized on a single fault plane, the brittle detachment itself (fig. 12A).

As a result, asymmetric patterns of low-angle normal faults are preferentially (but not exclusively) observed in weak lithologies and in the vicinity of the Tinos detachment. Symmetric high-angle normal fault patterns developed (i) possibly in an early stage of the strain history, either in more competent lithologies or in any lithologies but away from the main detachment shear zone, and (ii) in a late stage of the strain history within nearly all lithologies including the shear zone/fault rocks themselves (mylonites, cataclasites) which became stiffer (and less anisotropic) during exhumation while shearing ultimately localized along the brittle detachment.

SCENARIO OF LOCALIZATION PROCESS AND INITIATION OF BRITTLE FAULTS IN EXHUMING CYCLADIC METAMORPHIC ROCKS

Metasediments enclosing competent boudins of marbles or metabasites allows observation of how boudinage predated normal faulting in contrasted lithologies. Other localities are demonstrative of normal faulting in a several ten/hundred meters thick mass of metasediments with more homogeneous mechanical behavior. In addition, high angle normal faults cut across the entire rock pile and generally postdate the low-angle normal shear zones and faults.

A first-order scenario of evolution of deformation from ductile to brittle, under a continuous kinematic evolution is proposed in figure 12. Primary localization of ductile deformation is closely linked to boudinage. Extensional shear zones often localize in the less competent matrix at the tips or in the necks between boudins of early veins or
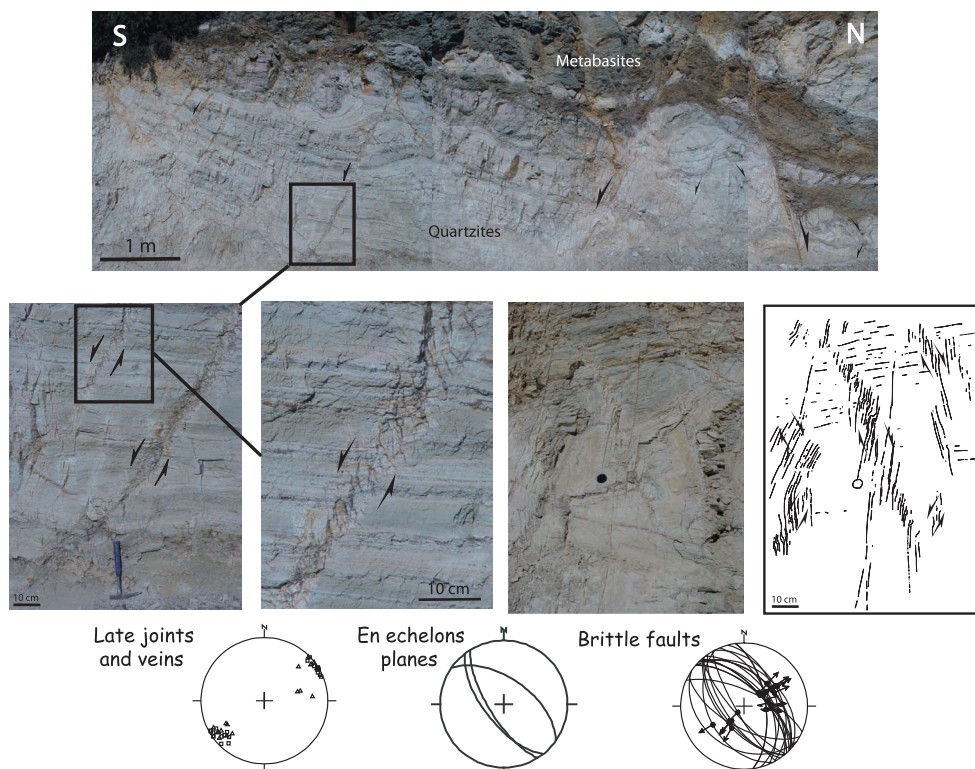
Fig. 10. – Examples of high-angle normal faults and en echelon sub-vertical vein patterns reflecting ongoing localization and incipient normal faulting in quartzitic metapelites (Andros, A1-A3 on fig. 1). Microstructural data: same key as in figure 4.

competent lithologies (metabasites, marbles), that is, in zones of stress concentration. The initiation of shear zones therefore postdated boudinage, in good agreement with the increasing degree of localization from boudins to shear zones.

Rheological heterogeneities and boudinage have to be considered as an efficient factor to initiate localization [Jolivet *et al.*, 2004; Mehl *et al.*, 2005]. These authors propose a scenario of evolution of early-localization of deformation: first, boudinage localizes deformation at intervals depending on the contrast of viscosity between strong and weak layers and of the thickness of the competent layers. Once initiated, this process is facilitated because the resistant layers are thinner and thus easier to deform at the tips or in the neck between boudins. There, the local increase of strain rate and/or stress concentrations allow development of extensional shear zones. This kind of observation has also been reported by Tricart *et al.* [2004] in the Alpine Queyras Schistes Lustrés.

The evolution toward brittle behavior is marked by the reactivation of the extensional shear zones as low-angle normal faults, by the progressive straightening of extensional structures and the development of *en echelon* arrays of

veins or joints (mode I opening) (fig. 12C). The ultimate step of localization consists in sliding across the *en echelon* patterns and the formation of steeply-dipping normal faults generally displaying conjugate patterns (fig. 12C).

As mentioned above, the lithological control is also very important during the last brittle increments of deformation: brittle behavior is preferentially observed (and presumably appeared earlier) in more competent layers (metabasites and quartzitic layers). Although the first-order scenario we propose is in good agreement with the sequential evolution of structures from ductile to brittle, the rheological behavior of materials appears as a key point in the localization process: rheological heterogeneities probably had a dominant effect on the depth where the structures initially localized during their way back to the surface.

This work documents that ductile shear zones localize brittle deformations. Interestingly, a number of recent papers have in turn documented that ductile shear zones could have been localized by brittle precursor fractures [e.g., Guermani and Pennacchioni, 1998; Pennacchioni, 2005].
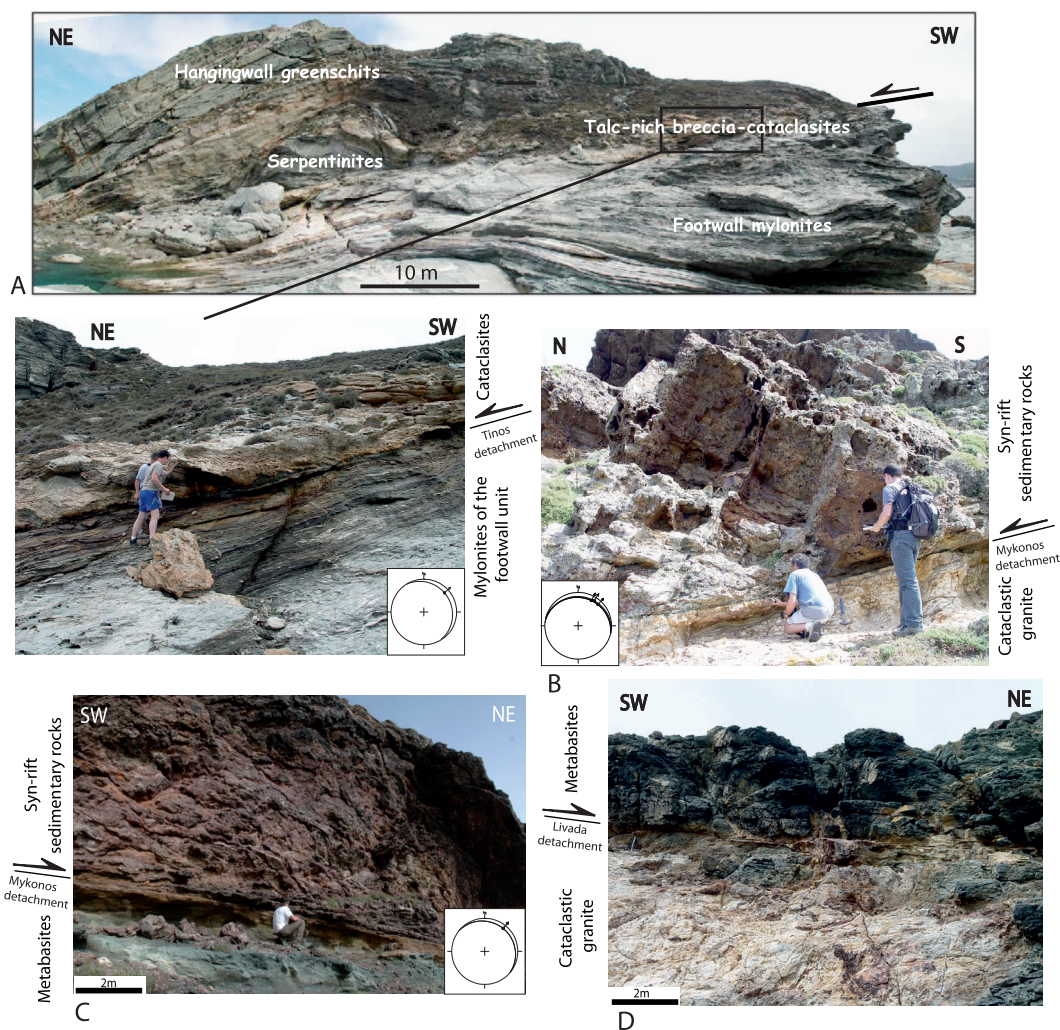
FIG. 11. – Examples of low-angle detachments in Tinos and Mykonos.
A: view of the Planitis island (T1 on fig. 1) with zoom on the Tinos detachment. B: Mykonos detachment putting syn-rift sedimentary rocks on top of the cataclastic granite (B; Mykonos, Panormos bay, M2 on fig. 1) or on top of metabasites of the Upper Cycladic unit (C: Mykonos, Cape Evros, M1 on fig. 1). D: Livada ductile detachment putting metabasites on top of cataclastic granite, reworked by a late low-angle normal fault (Mykonos, Cape Evros, M1 on fig. 1).
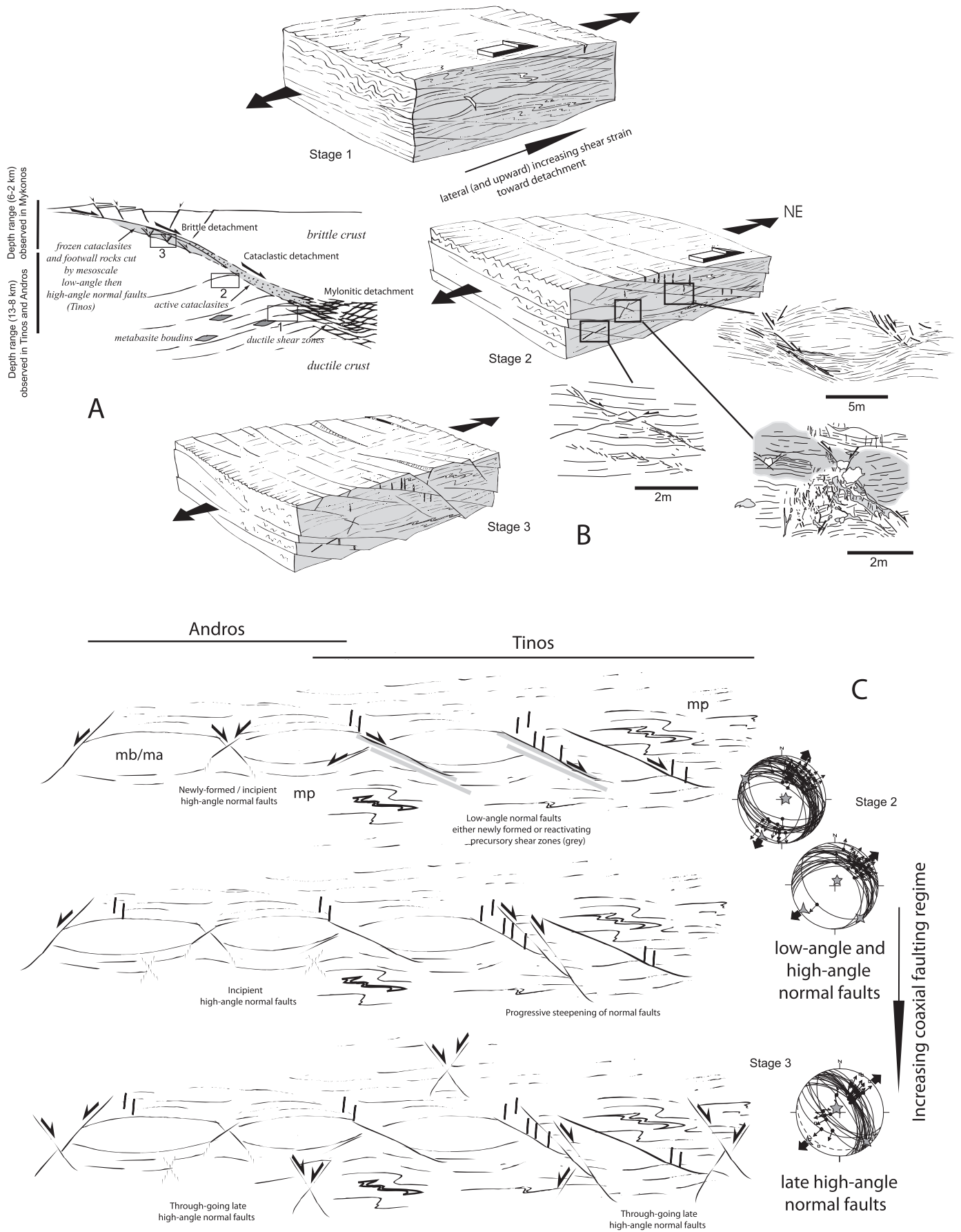
FIG. 12. – Scenario of progressive localization of structures from ductile to brittle. A: Evolutionary shear zone at the scale of an extending crust, (B) Schematic evolution in the footwall of the Tinos detachment (B) [modified after Mehl *et al.*, 2005]. C: Sequence of brittle faulting corresponding to the stages 2 and 3 of (B).

FIELD CONSTRAINTS ON THE MECHANICS OF LOW-ANGLE NORMAL FAULTS IN METAMORPHIC DOMES

The mechanics of low-angle normal faults (LANFs) is a controversial topic [e.g., Scott and Lister, 1992; Axen, 1992; Abers, 2009; Collettini, 2011]. According to the Anderson-Byerlee frictional fault mechanics, in an extending crust submitted to an unperturbed Andersonian stress regime (with a sub-vertical σ1 axis) and displaying faults with friction coefficients of 0.6 to 0.85, brittle normal faults are expected to initiate at 60° dip and to rotate down to 30° while active [e.g. Sibson, 1985, 1990]. The existence of active low-angle normal faults is much debated because the theory of fault mechanics implies that faults are locked when the dip is less than 30°. Active normal faults may then be further reoriented passively as inactive structures to low angles either due to rotation during later episodes of normal faulting along new, steeply dipping structures or due to isostatic adjustments [e.g., Wernicke and Axen, 1988; Buck, 1988].

However, a number of field observations from the Cycladic islands suggest that slip may occur along low-angle normal faults at very shallow dip in the brittle field, as documented at other places from structural and seismicity studies [e.g., Woodlark basin: Taylor and Huchon, 2002; Corinth rift: Rigo *et al.*, 1996; Appenines: Collettini and Barchi, 2002].

Evidence for slip at shallow dip

Sedimentary evidence

In Mykonos, sedimentary deposits are observed at Cape Evros (fig. 1, M1). They are made of sandstones and are bounded by steep normal faults soling within the cataclastic Mykonos detachment. These deposits display a fan-shaped geometry, the dip of strata evolving from 30°SW at the base to sub-horizontal on the top of the fans; a thin sub-horizontal sedimentary layer overlies the fan-shaped deposits (fig. 13A). This attitude of these hanging wall rift basin deposits, which are in many places shallowly dipping, and have locally steep dip domains in opposite directions therefore demonstrate that slip on the brittle-cataclastic Mykonos detachment unambiguously occurred while it was at very low dip. Paleomagnetically constrained rotation about an horizontal axis of the footwall Mykonos granite [Morris and Anderson, 1996; Avigad *et al.*, 1998] therefore does not require (hence does not imply) a steeper dip for the detachment [Lecomte *et al.*, 2010; Denèle *et al.*, 2011].

Microstructural evidence

Many additional microstructural observations support that meso-scale LANFs as well as brittle detachments have slipped at shallow dip in their present attitude (or very close to it) without having undergone significant post-slip tilt. Evidence come from the close association of these faults with sub-vertical veins (in Planitis, fig. 13B and C; in Kolimbithra, fig. 13D and E) consistent with a nearly vertical shortening direction [see also Mehl *et al.*, 2005]. In Tinos, this a priori vertical position of the compressive stress axis is further confirmed by the presence of dacitic dikes on the island, that are assumed to have intruded as vertical sheets and are still vertical [Avigad *et al.*, 1998].

Axen and Selverstone [1994] already argued for a vertical maximum stress axis around low-angle normal faults in metamorphic core complexes.

Note that because brittle slip along the Tinos and Mykonos detachments occurred at shallow dip, the sub-vertical attitude of the maximum principal stress as derived from minor joints, veins and normal faults as well from sub-vertical barite dikes (in Mykonos) argue in favor of the mechanical weakness of the branches of the NCDS.

Evidence for reactivation of precursory shear zones

In many cases, even if slickensides are not always observable, the shallow dip of the fault planes synthetic of the main detachment and their geometrical association with boudins lead us to conclude that they correspond to reactivated ductile shear zones. When brittle slip occurs along previous ductile shear planes in a direction parallel to the stretching lineation, this superimposition can be viewed as a kind of reactivation of a precursory ductile structure.

Some shallow-dipping faults can thus be considered as having developed with precursory structures such as ductile shear zones. However, shear zones do not consist of surfaces of displacement discontinuity; they are not themselves, strictly speaking, faults. So the term reactivation, commonly understood as sliding along a pre-existing discontinuity, should be used with care. Reactivation corresponds here to a continuum of shear from ductile to brittle with an increasing localization within a precursory shear zone, that locally modified either the mechanical properties of the rocks, the local strain rate or the local stress field to enhance shallow-dipping brittle faulting during decrease of P and T. Noticeably, only the more steeply dipping shear zones show reactivation as brittle faults.

At a much larger scale, the NCDS has been proposed to partly reactivate the Vardar ocean suture zone including the contact zone between the Pelagonian domain and the Cycladic blueschists, and mechanically weaker lithologies [Jolivet *et al.*, 2010]. The distribution of extensional deformation thus seems to be largely controlled by the presence of a weak rheological level, i.e. the inherited thrust contact at the base of the Pelagonian.

New insights into the mechanics of LANFs

There are two major questions that we address in this section: (1) the mechanical feasibility of slip along shallow-dipping pre-existing or precursory shear zones (i.e., the extensional reactivation of these shear zones as brittle LANFs); (2) the initiation of LANFs. Although of major interest also, it is out of the scope of this paper to discuss the amounts of regional extension accommodated by these LANFs.

The observations in the north Cycladic islands unambiguously demonstrate that models involving either regional rotated stress axes (i.e., non Andersonian regional stress regime outside the fault zone) or rotation of the fault planes do not apply here. Both LANFs and detachments are oriented at high angle to, and have slipped (at least during the late increments of displacement) under, a subvertical σ1, and can therefore be classified as 'weak' faults. The physical cause of fault weakness is still a matter of debate. Slip is made theoretically possible by elevated pore fluid pressure

with low tensile strength [Axen, 1992; Collettini and Barchi, 2002], by weakening of fault rocks through reaction softening [Gueydan *et al.*, 2003; Grasemann and Tschegg, 2012], by stress rotations either in the fault core [Axen, 1992] or at the base of the seismogenic zone [Westaway, 1999; 2005] or by the presence of a preexisting shallow dipping nappe with a competence contrast with the crust [Le Pourhiet *et al.*, 2004; Huet *et al.*, 2011a, b].

Most of the low-angle normal faults observed in the northern Cyclades consist of cataclasites or fault gouges that either overprinted or reworked 0.5 to 5 km thick mylonitic ductile shear zones, or were localized along the margin of these shear zones. As mentioned earlier, a planar

fault surface, formed at shallow crustal levels, usually caps, or is capped by, the cataclastic fault rocks (fig. 11). This type differs from discrete fault cores (1-20 m thick) separating hangingwall and footwall blocks affected by brittle processes: in the Zuccale fault (Elba island), the damage zone is characterized by fractures, small displacement faults and veins. The fault rocks within the fault core formed by diffusion mass transfer processes and/or cataclasis with grain-size reduction, rotation and translation of grains [Collettini, 2011]. Note however the occurrence of ductile deformation in the footwall of this fault (Calamiti schists), with E-W stretching and top-to-the-east shearing [Daniel and Jolivet, 1995].
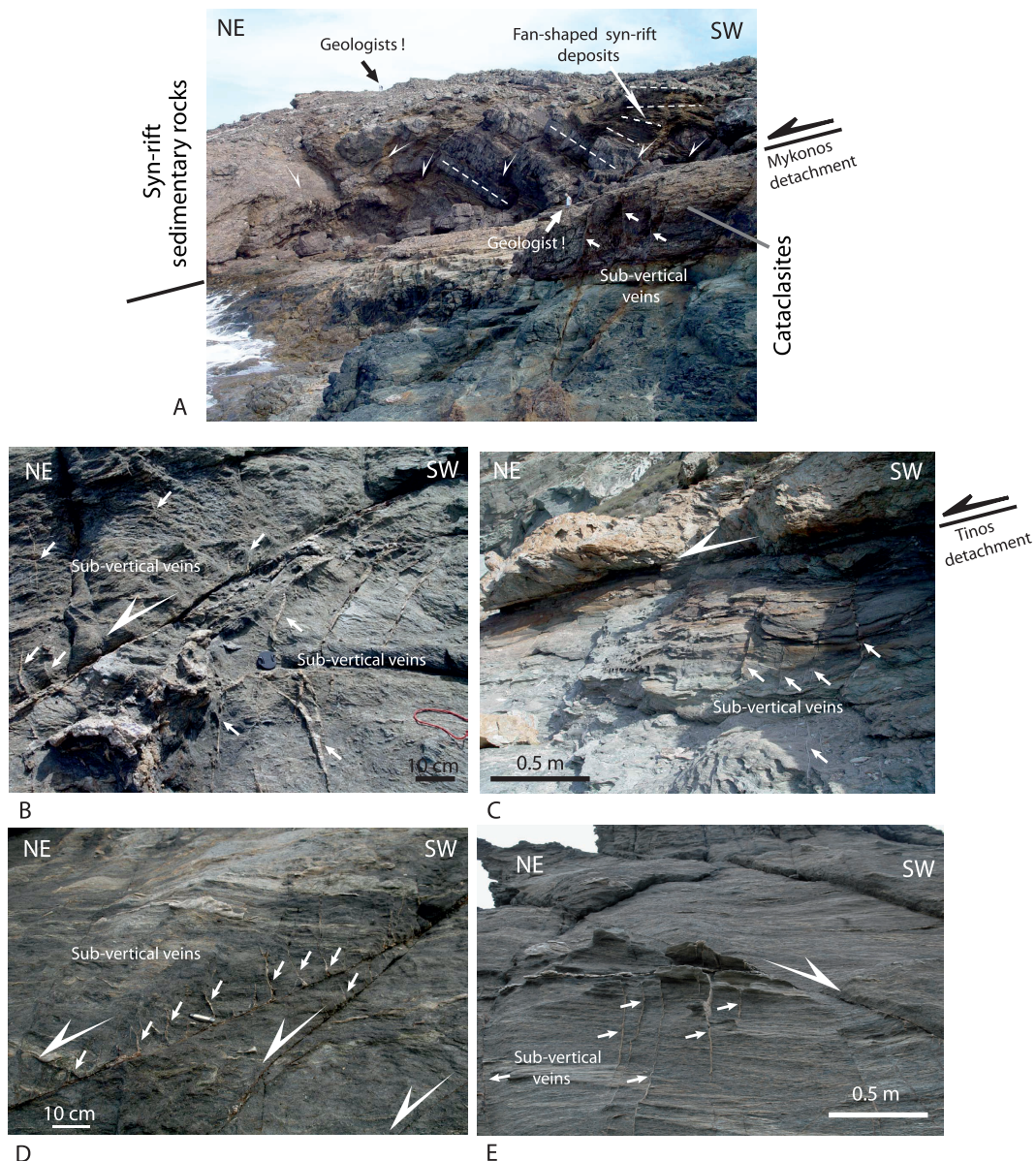
FIG. 13. – Sedimentary and microstructural evidence for slip at shallow dip of normal faults and detachments in Tinos and Mykonos.
A: Sedimentary deposits (Mykonos, Cape Evros, M1 in fig. 1) bounded by steep normal faults soling within the Mykonos detachment and displaying a fan-shaped geometry, the dip of strata evolving from 30°SW at the base to sub-horizontal on the top of the fans A thin sub-horizontal sedimentary layer overlies the fan-shaped deposits. This attitude of the hanging wall rift basin deposits precludes any post-slip tilt and demonstrates that slip on the Mykonos detachment unambiguously occurred while it was at very low dip.
B, C, D, E: Close association of low-angle normal faults with sub-vertical vein sets in Tinos (Planitis, T1 in fig. 1, B, C) and Kolimbithra (T3 in fig. 1, east of the Bay) (D, E), indicating that brittle slip along the Tinos detachments and low-angle normal faults occurred at shallow dip and with a sub-vertical attitude of the maximum principal stress.

A possible mechanism of slip at shallow dip involves weakening of the fault rocks as for the Zuccale fault [Collettini and Holdsworth, 2004]. Field-based and micro-structural studies suggest an evolution from an initial brittle cataclasite to a narrow foliated fault core formed as the result of syntectonic fluid-rock interactions. Fluids reacted with the fine-grained cataclasite to produce fine-grained aggregates of weak phyllosilicate-rich fault rocks leading to reaction softening. In addition, the fine grain sizes trigger the widespread onset of stress-induced dissolution and precipitation processes (grain-size sensitive flow). Experimental analogues suggest a switch in rheology from a cataclastic deformation to a pressure solution-accommodated frictional slip, the switch being associated with a decrease in friction to 0.2 or less. With a friction coefficient of 0.2, LANFs could move easily in a stress field with vertical $\sigma 1$. Furthermore, such weak faults would be incapable of generating big earthquakes because at low-sliding velocity the pressure-solution accommodated deformation is a velocity strengthening process favoring aseismic slip [Collettini and Holdsworth, 2004].

In Tinos, the primary high mica and chlorite content of rocks (especially metapelites of the Cycladic blueschists) may account for the intrinsic weakness of the rock material, without requiring any additional metamorphic reactions to produce weak mineralogical phases. In contrast, taking into account the widespread occurrence of fluid-assisted vein development in the cataclasites associated with the Tinos detachment, fluids likely played a major role in strain localization, causing softening either by enhancing ductility or increasing pore pressure. The first fluid input from the surface down to the brittle-ductile transition was likely made possible by the early shear zones formed between the boudins [Jolivet *et al.*, 2004; Famin *et al.*, 2005] that were invaded by fluids issued from the surface, which had in turn favored further slip across the same shear zones and development of brittle faults.

The feasibility of brittle slip accommodating large displacements along shallow-dipping pre-existing (ancient thrusts) or precursory (extensional ductile shear zone) structures has alternatively recently been investigated by means of analytical and numerical modelling [Lecomte *et al.*, 2011; 2012]. Lecomte *et al.* [2011] proposed a new model for fault reactivation by introducing an elasto-plastic frictional fault gouge as an alternative to the common dislocation models with frictional properties. Contrary to the classical model, which implies that the dilation angle equals the friction angle, the model permits an incompressible or a compacting (thinning) fault gouge as deduced from laboratory and field observations. The predicted locking angles (dip angles below which the faults are inactive) differ in most cases by less than 10° from the classical model, and in addition, a significant amount of strain (in the elasto-plastic regime) is predicted to occur on badly oriented faults prior to locking in a strain-hardening regime. This has led to conclude that plastic strain on badly oriented faults is favored by compaction of the fault gouge [Lecomte *et al.*, 2011]. This strain results in a rotation of principal stresses within the fault and therefore modifies the effective friction of the fault. Note that such a local stress rotation within the fault zone has alternatively been explained by a decrease of the elastic compressibility towards the fault [Faulkner *et al.*, 2006].

The model of Lecomte *et al.* [2011] predicts different modes of reactivation of a precursory shear zone, including a complete reactivation mode (steady state slip) and a partial reactivation mode for which stress magnitudes in the embedded medium, that increase since slip along a badly oriented normal fault zone poorly releases stresses, reach the failure criterion; in the partial reactivation mode, the LANF slips transiently and new high-angle faults may form in the surrounding medium. This case of mixed low-angle and high-angle normal faulting and tension failure (fig. 14) is illustrated in Kolimbithra (fig.1, T3) just below the Tinos detachment (e.g., fig. 6A). The interest of this model is to possibly account for repeated events of slip on a LANF alternating with formation of high-angle faults in the surrounding medium, either with development of tensile failure or not.

The component of compaction of the fault zone involved in the model of Lecomte *et al.* [2011] therefore leads to a significant drop of the effective friction of LANFs which allows faults with internal friction of 0.3 to slip at dip as low as 20°. In this regime, the thick fault model predicts that deviatoric stress rises with accumulated plastic strain on LANFs, favoring a stable slip regime, in agreement with the observations that those faults are generally aseismic (absence of earthquakes with magnitudes greater than 5.5) [e.g., Collettini *et al.*, 2006; Rigo *et al.*, 1996; Bernard *et al.*, 1997]. However, within the rotated state of stress of the fault zone, it is also possible to newly form well-oriented secondary faults. These smaller faults form in a slip-weakening regime and are to that respect dynamically unstable. Their orientations depend on the dilation angle of the fault zone but in general, they are confined to the width of the fault zone and therefore their size is limited. Therefore, seismic activity on these secondary shears is

FIG. 14. – Mode of partial extensional reactivation of a preexisting low-angle shear zone with tensile failure in the surrounding medium (according to the thick fault model of Lecomte *et al.* [2011]), illustrated by the association of low-angle normal faults with high-angle normal faults and sub-vertical veins in Kolimbithra (fig. 6A). Yield criteria and Mohr circles plotted in red/black define the stress state in the embedded medium/inside the shear zone. Both media are characterized by five mechanical parameters. The Poisson ratio and the shear modulus characterize their elastic properties. The friction angle ϕ, the dilation angle ψ and the cohesion Co characterize their plastic properties. The superscripts *in* and *out* refer to parameters within and outside the shear zone, respectively. Because of stress continuity condition, the two Mohr circles must intersect on the fault plane defined by its dip δ. β is the angle between the plastic flow and the shear zone; it tends toward ψ with increasing strain [for more details, refer to Lecomte *et al.*, 2011]. ϕ^{eff} : effective friction of the fault zone.

necessarily of limited magnitude as it is often observed on active LANFs and other weak faults [Lecomte *et al.*, 2012].

Whereas slip along shallow-dipping fault zones, hence the ability of LANFs to accommodate significant amounts of extensional displacement is still little understood, the nucleation of LANFs within intact rocks is even more problematic. As emphasized by Collettini [2011], to explain the initiation of these structures it is not possible to invoke a stress rotation near the brittle-ductile transition [e.g. Melosh, 1990; Westaway, 1999; Yin, 1989] since they are within the brittle crust. At the same time stress rotation induced by: a) a weak fault core sandwiched in a strong crust [e.g. Axen, 1992], or b) a fractured damage zone affected by changes in elastic properties [Faulkner *et al.*, 2006], are unlikely since at fault initiation no fault core nor damage zone are present. This is the case for the small-scale shallow-dipping normal faults described above (section *Newly-formed low-angle normal faults*). A promising way consists of taking into account rock anisotropy, like the elastic anisotropy of fault core rocks [Healy, 2009]. Natural fault zones are generally characterized by one narrow core zone flanked by wider damage zones. Some fault cores show foliated rocks with intrinsic anisotropy related to the strong preferred alignment of phyllosilicate minerals and extrinsic anisotropy from arrays of grain boundary pores and micro-cracks. Healy [2009] proposed that a stress rotation occurs in such fault core and that this applies to the nucleation of LANFs when layers with distinctly weaker material properties, whether anisotropic or isotropic, are inclined with respect to the imposed maximum compression. Such a rotation of σ_1 is enhanced by increasing pore fluid pressures and additional extrinsic anisotropy. Unfortunately, measurements of elastic stiffness from natural shallow crustal fault rocks are often lacking to reliably constrain this elastic anisotropy.

In the Cyclades however, some LANFs lack foliated fault core rocks. The cores of these fault zones are dominated by gouges and breccias, and even those with significant clay content show little anisotropy. Stress rotations are however possible in granular quasi-isotropic fault rocks: cataclastic deformation can produce an increase in Poisson's ratio, and this will rotate σ_1 towards the fault core [Healy, 2009].

At a smaller scale, the structures from Andros shown on figure 7 strongly suggest that the low-angle normal fault system (reactivated and newly-formed) have slipped with an uncommon high angle to the vertical direction presumably parallel to the σ_1 axis. The newly-formed brittle SW dipping antithetic LANF did not initiate with the theoretical (and expected) angle with respect to σ_1, but may have rather been influenced/controlled in some way by the NE dipping fault reactivating a precursor shallow-dipping shear zone. Again, although such observations (Andros, Tinos) have to be confirmed, there is no way to invoke a rotation of the stress field or any other mechanisms of fault rotation.

In theory, newly-formed low-angle normal faulting is predicted either by the CamClay model [Roscoe, 1970] within a material that can compact plastically or by a non associated Mohr-Coulomb plasticity model [Lecomte *et al.*, 2011; Le Pourhiet, 2013]. The only compacting materials where newly formed low-angle normal faults have been experimentally observed are clays and sandstones [Besuelle *et al.*, 2000]. Compacting bands are also well recognized in reservoirs [e.g., Saillet and Wibberley, 2010]. However, this compacting behavior, whatever its cause, seems very unlikely in the metamorphic material investigated in our study. As a result, despite the wealth of new models, our mechanics still fails at satisfactorily explaining the initiation of low-angle normal faults in metamorphic rocks.

RELIABILITY AND SIGNIFICANCE OF PALEOSTRESS RECONSTRUCTIONS IN ANISOTROPIC METAMORPHIC ROCKS

In Tinos, inversion of fault-slip data for stress has been carried out using Angelier's [1984, 1990] methods [Mehl *et al.*, 2005]. The results confirm the sub-vertical orientation for the maximum principal stress axis, in spite of the variations between lithologies (fig. 4). This sub-vertical orientation is consistent with the patterns of sub-vertical late veins often associated with brittle normal faults. The compression direction is located in the acute angle between late conjugate fault systems, but located in the obtuse angle of the LANF systems, σ_1 making an angle greater than 45° with each low angle fault plane. The extension axis remains sub-horizontal or gently dipping with a clearly defined NE-SW direction.

Andersonian's mechanics suitably explains the formation of the observed steeply-dipping conjugate planes. During their way back to the surface, footwall rocks undergo a decrease in temperature and pressure and evolve towards a more competent rheology and isotropic behavior, so that their angle of internal friction increases. The larger the friction angle, the smaller the angle between σ_1 and the fault plane.

In contrast, numerous outcrops investigated in Andros reveal paleostress tensors with stress axes neither vertical nor horizontal [Mehl *et al.*, 2007]. There is no unambiguous marker of the paleo-horizontal plane, but interestingly the computed stress axes show particular relationships with the tilted foliation: σ_1 axis is roughly perpendicular to the foliation while σ_2 and σ_3 roughly lie within the foliation plane (fig. 4, A3 to A6). It is comparable to Tinos where the reconstructed σ_1 axes are found generally perpendicular to the flat-lying foliation [fig. 4; Mehl *et al.*, 2005]. This is in agreement with the common attitude of late veins with respect to foliation. Assuming an unperturbed Andersonian state of stress, Mehl *et al.* [2007] rotated the foliation back to horizontal and interpreted all the brittle structures as having formed under a vertical maximum stress axis σ_1 in both islands, but having subsequently been locally tilted in a late stage of deformation in Andros. A similar reasoning has led Tricart *et al.* [2004] to propose that in the Queyras, the whole Ligure-Piemont Schistes Lustrés unit has been tilted as a monocline along the extensionally reactivated Penninic front during a late stage of Alpine deformation.

Late tilting of the metamorphic pile in Andros could be attributed to large-scale open folds described by Papanikolaou [1978] and Avigad *et al.* [2001], or to late high-angle normal faulting [Philippon *et al.*, 2012], or both. In all cases, the possible flat attitude of the foliation at the time brittle structures developed deserves consideration, since the foliation is not a priori a marker of the paleo-horizontal. Three hypotheses can be made: (1) the foliation was flat at the time of brittle deformation, and was subsequently tilted

during late doming by high-angle normal faults; but this large-scale tilting does not fully account for field evidence of local tilting with subvertical foliation [e.g., Mehl *et al.*, 2007]; (2) doming began to develop in ductile conditions but the curvature remained gentle and ductile folding remained limited before brittle deformation occurred, so the foliation remained nearly flat at this stage on most of the islands. Doming and folding in Andros were thus mostly achieved after the onset of brittle deformation, although still within greenschist conditions [Avigad *et al.*, 2001]; (3) despite a first-order continuous evolution from ductile to brittle, local rheological contrasts and/or strain rate variations could have led to alternating ductile and brittle behavior across the transition, leading, for instance, to brittle deformation within stiff metabasites while the weaker pelitic matrix was still deforming more or less ductilely by folding. Doming and large-scale folding could have remained limited at the time of occurrence of the first increment of brittle deformation, and have later tilted brittle structures developed mainly in competent material. This may suggest that folding, which is related to NW-SE shortening perpendicular to extension, certainly initiated in ductile conditions but ended in the brittle field.

Although a late component of tilting by folding or by high-angle normal faulting remains likely, an alternative, although provocative view must however be considered, in which foliation was not horizontal but already domed before the onset of brittle faulting, and in which the regional vertical $\sigma 1$ axis has been locally reoriented toward the normal to the foliation plane, mimicking a pre-tilting Andersonian state of stress. Among other hypotheses, fault slip inversion methods assume that the analyzed body of rock is physically homogeneous and isotropic and if pre-fractured, it is also mechanically isotropic, i.e., the orientation of fault planes on which slip accumulates is random. In practice these methods were extensively and successfully applied to sedimentary rocks that are somewhat anisotropic because of bedding and fractures [see discussion in Lacombe, 2012]. As reported in this paper, these methods were also applied to brittlely deformed anisotropic foliated metamorphic rocks, and yielded regionally significant results in terms of direction of extension and of continuous kinematics from ductile to brittle during exhumation. However, the possible influence of the pre-existing foliation anisotropy on brittle faulting in terms of local re-orientation of the maximum principal stress $\sigma 1$ toward perpendicular to the foliation has been poorly investigated hence little documented to date.

To test such possible reorientation, that would suggest that the assumption of an Andersonian state of stress could be no longer valid if the material is strongly anisotropic, more information about rock properties as well as numerical modeling are required. If such an effect of the foliation anisotropy is documented in the future, it will draw attention on the need for more caution when concluding on the late tilting of the whole metamorphic pile and the timing of acquisition of the dome shape in Tinos and Andros. It will more generally also question the validity of the Andersonian stress hypothesis when inverting fault-slip data for stress in strongly anisotropic rocks: using fault-slip data inversion under the Andersonian stress hypothesis to infer the paleo-horizontal and paleo-vertical as stated by some [Hippolyte *et al.*, 2012], although possible in sedimentary environments, should in this case be considered with care.

CONCLUSIONS

This synthesis of field observations in the northern Cycladic islands brings some new insights into the way brittle faults initiate in metamorphic rocks, which are being exhumed up to the upper crust. Our approach is complementary to that of Crider and Peacock [2004] who reviewed the styles of initiation of faults in the upper crust within previously unfaulted (sedimentary) rocks. The various styles they recognized are partially encountered in the present study, such as initiation as mode I fractures or as precursory shear zones. We document the influence of preexisting rheological and structural anisotropy and the likely control of strain rate variations, local change of rock properties and or local stress perturbations by or within ductile or semi-brittle precursory shear zones on the initiation and the geometry of brittle faults.

Although the proposed sequence of initiation of mesoscale brittle faults during exhumation is rather well constrained, some structures are not satisfactorily accounted for by existing rock mechanics models. In addition, despite many recent advances, there is not yet any convincing unified mechanical model to describe initiation of, and even slip along large-scale LANFs and detachments.

Finally, with respect to inversion of fault-slip data for stress in sedimentary rocks (in which field Jacques Angelier (1947-2010) has undoubtedly been a pioneer, e.g., Angelier, [1975, 1979a, 1984, 1990]), applying the technique to anisotropic metamorphic rocks, although likely providing geologically significant results, may require in the future careful and thoughtful consideration and further investigation of the validity of the Andersonian stress hypothesis.

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