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Structure, kinematics and metamorphism in the Liguride Complex, southern Apennines, Italy

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Abstract—The Liguride Complex of south Italy is the remains of a Cretaceous–Paleogene accretionary wedge which was emplaced onto the African margin during Early Miocene time. Transport directions of thrust sheets and nappes in the internal zone are roughly towards the present-day northeast. Transport directions match well with the corresponding slip vectors derived from relative motion of the Calabria–Sardinia–Corsica block with respect to Africa (including Adria) during Late Eocene to Oligocene time. Deformation of the internal zone included extensive underthrusting followed by underplating of already lithified sediments, ophiolitic rocks and continental basement on to the base of an accretionary wedge down to depths in places exceeding 30 km. Overthickening by massive underplating caused Late Paleogene gravity-driven extension within the accretionary wedge. The contact between the lower ophiolitic nappe and the upper ophiolitic nappe is an extensional fault zone of this age. Transport directions during the gravity-driven deformation do not match plate derived slip vectors. It is inferred that the gravitational potential of the wedge dominated the kinematics of the Late Palaeogene extensional deformation.

INTRODUCTION

THE structure of the Calabrian arc can be divided into three elements from bottom to top (Ogniben 1969): the Panormide Complex, the Liguride Complex and the Calabride Complex.

The Liguride Complex of southern Italy, the subject of this paper, is a rock-stratigraphic unit comprising a Mesozoic to Cenozoic allochthonous flysch interleaved with blocks of oceanic and continental material. It occupies the middle position within the nappe pile of the Calabrian arc (Ogniben 1969) (Fig. 1). The Liguride Complex crops out in an extensive tract from the Cilento Peninsula to the Catanzaro Trough, and in northeast Sicily (Fig. 1.)

The Calabride Complex consists of a series of massifs of igneous and metamorphic basement nappes and their sedimentary cover extending from the Coastal Chain in north Calabria to the Peloritani of northeast Sicily (Fig. 1). The sedimentary cover varies in composition for various massifs but generally comprises Lias to Cretaceous carbonate rocks overlain unconformably by Tertiary terrigenous clastics.

The Panormide Complex, predominantly comprising thrust sheets of Mesozoic platform carbonates overlain unconformably by Tertiary terrigenous clastics, crops out in the southern Apennines (Fig. 1), in a few windows through metamorphic and igneous basement nappes in Calabria, and in the Palermo mountains of northwest Sicily.

Table 1 summarizes tectonostratigraphic schemes proposed for the Calabrian arc. There are essentially two conflicting schemes which differ in their interpretation of the palaeogeography and kinematics of the mountain belt. Ogniben (1969) placed the original site of the Calabrian basement nappes on the European margin, while Haccard *et al.* (1972) and Amodio-Morelli *et al.* (1976) placed them on the African margin.

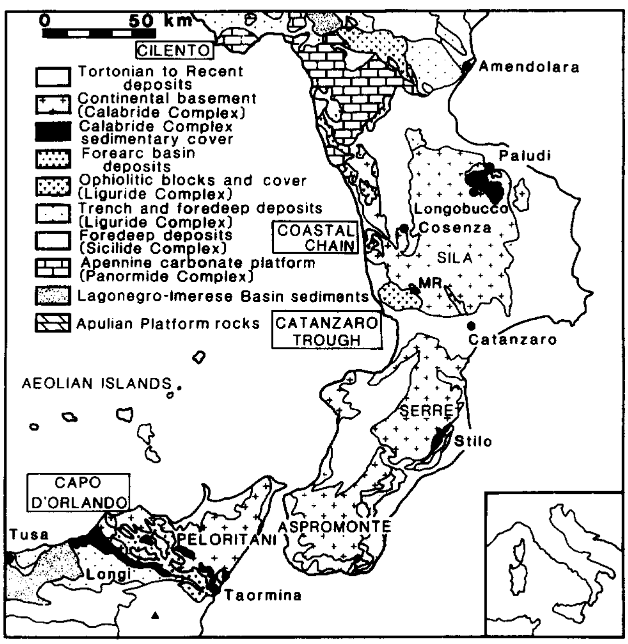

Fig. 1. Geological map of the Calabrian arc. TS, Terranova da Sibari; SA, Spezzano Albanese; MR, Monte Reventino. The lower right panel depicts the Italian mainland where the Apennines are located. The Maghrebides occur on the island of Sicily and on the north African mainland. See text for elaboration. Modified after Amodio-Morelli *et al.* (1976).

Recently there has been a return to the general scheme of Ogniben (1969). Bouillin (1984) and Bouillin *et al.* (1986) noted the similarity between the structure of Calabria and the structure of the internal massifs of

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Table 1. A comparison of tectonostratigraphic schemes for the Calabrian arc

Ogniben (1969)	Amodio-Morelli *et al.* (1976)	Knott (1988)
Calabride Complex	Polia Copanello Unit Castagna Unit	Continental basement nappes (Upper plate) (Paludi Formation, Stilo-Capo D'Orlando Formation)
	?Bagni-Fondacelli Unit Malvito Unit	Upper ophiolitic nappe (Accretionary wedge) (Crete Nere Formation, Pollica Formation, Albidona Formation)
Liguride Complex	Diamante-Terranova Unit Gimigliano Unit Frido Unit	Lower ophiolitic nappe (Accretionary wedge) (Frido Formation)
Panormide Complex	Verbicaro Unit San Donato Unit	Apennine Platform (Lower plate)

Petite Kabylie and Grande Kabylie in North Africa. Bouillin (1984) concluded from this comparison that the Calabrian basement nappes were of European origin.

Faure (1980) documented E–W-trending stretching lineations in the basement nappes of central Calabria, but shear criteria were equivocal indicating both E- and W-directed shear. The work of Knott (1987, 1988) and Dietrich (1988) furnish the first unequivocal kinematic data on transport directions in the Calabrian arc, providing an objective means of discrimination between the conflicting schemes of Ogniben (1969) and Amodio-Morelli *et al.* (1976) (Table 1).

The purpose of this paper is to present new kinematic data in the framework of a plate reconstruction of the western Mediterranean (Dewey *et al.* 1989), to describe the structural history of the southern Italian mountain belt and to demonstrate that a modified version of Ogniben's scheme is the most plausable.

STRUCTURAL ELEMENTS AND SEQUENCE

In the internal zone of the southern Apennines, part of which is shown in Fig. 2, two distinct ophiolitic nappes can be recognized. The lower ophiolitic nappe (LON) comprises polydeformed metasedimentary rocks

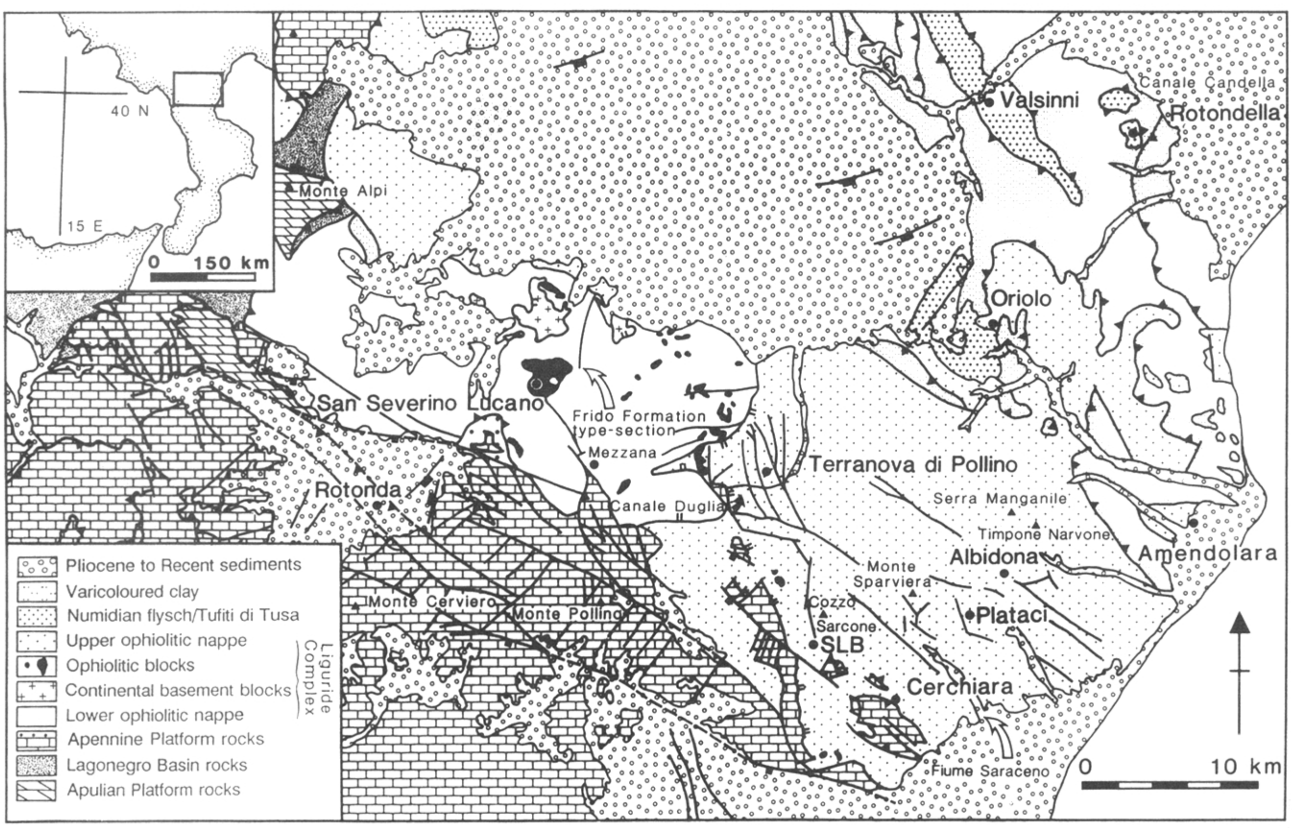


Fig. 2. Geological map of part of the southern Apennines (based on mapping by the author at 1:10,000 scale and reinterpretation of published geological maps from the 1:25,000 series, Carta Geologica della Calabria). Thrusts—barbs on overthrust plate. Normal faults—squares on downthrown side. Low-angle extension faults—double ticks on hangingwall plate. SLB—San Lorenzo Bellizzi.

ranging in age from Late Jurassic to Eocene and metamorphosed to incipient blueschist facies. The upper ophiolitic nappe (UON) comprises a Late Jurassic to Late Cretaceous succession showing signs of metamorphism only along the contact with the underlying lower ophiolitic nappe. Metasedimentary rocks of both nappes are interleaved with kilometre-size blocks of oceanic and continental material. Structural analysis of the internal zone will be concerned with erecting a local deformation sequence that will not necessarily be applicable elsewhere in the Liguride Complex (e.g. Carrara & Zuffa 1976). Significant variations in timing of formation of structures are likely to occur on the scale of the southern Apennines and Calabrian arc.

Deformation phases are labelled chronologically a–e. The structures in the internal zone were generated during deformation events D_a to D_e. D_a represents any deformation prior to subduction, e.g. extensional shear deformation of continental crust during Jurassic rifting. Structures belonging to D_b through D_e formed during progressive deformation within the internal part of an E-directed accretionary wedge from Eocene to Early Miocene time. Orientations are shown in Fig. 3.

D_b—*transposition foliation and thrusts*

The earliest recognizable mesoscopic structures in the deformation sequence (besides D_a noted above) are a slaty cleavage in pelitic rocks (Fig. 4a) associated with a spaced cleavage in psammatic lithologies (Fig. 4b). The deformation (D_b) that generated these structures also caused transposition with isoclinal fold limbs lying parallel to bedding (bedding-parallel foliation) (Fig. 4c). Downward-facing folds related to subsequent generations indicate that there was a major overturning of stratigraphy during D_b.

Vergence of F_b intrafolial folds is difficult to determine, but gouge fabrics along folded D_b fault planes that continue up-section in the direction of transport, indicate thrust transport roughly towards the present-day northeast. Early folds (F_b) are locally visible along mylonite zones, probably associated with emplacement of serpentinite sheets and continental blocks (Fig. 4d). These contacts have been reactivated during later phases (see below). A slaty transposition foliation is widespread in the internal zone. Locally, pelitic rocks develop a schistosity and the local increase in fabric and recrystallization intensity is related to strain increase in the proximity of major shear zones, either within the metasedimentary rocks or next to large blocks of oceanic and continental affinities (Fig. 4d).

NNE-directed thrusting

In the metasedimentary rocks of the internal zone D_b transposition caused development of isoclinal folds, slaty cleavage and a NNE-trending stretching lineation (Fig. 3a). F_b fold axes cluster around a roughly WNW trend (Fig. 3e) after the effects of D_c folding are removed by rotation of the F_b data points around a pole defined by the F_c fold axis. The earliest deformation recorded in the metasedimentary rocks (D_b) is related here to the original thrust emplacement of the oceanic and diorite blocks. However, the majority of kinematic data from the metasedimentary rocks in contact with the blocks is related to later reworking of these contacts during subsequent deformation.

D_b minor thrusts occur within small strain shadow zones adjacent to the serpentinite megaboudins and are not reworked by later deformation. In the shadow zones minor thrusts with gouge fabrics and shear bands indicate thrust transport towards 010, roughly perpendicular to the general trend of F_b fold axes (mean trend 100; Fig. 3e). It appears then, from the few but consistent data, that thrust sheet transport during the first stages of contraction was roughly towards the north-northeast (Fig. 8a).

D_c—*crenulations and thrusts*

The second generation of structures developed in the internal zone include NE-verging folds and thrusts. Folds are locally overturned and are accompanied by an axial planar crenulation cleavage (Fig. 5a). F_c folds fold D_b thrust planes, schistosity and spaced cleavage (S_b) (Fig. 5b) and are commonly downward facing (Fig. 6). Locally F_b isoclines are refolded around F_c asymmetric folds (Fig. 7a). D_c shear zones are associated with extensive calcite and aragonite veining (Spadea 1976). Calcite–aragonite veins produced during D_b are locally deformed into asymmetric shapes by D_c shear movement (Fig. 5c).

Crenulation cleavage (S_c) is localized and generally weakly developed. S_c crenulation cleavage is best developed within major F_c fold hinges and also adjacent to the large blocks of oceanic and continental material (Fig. 5a). Aragonite is commonly found in the microlithons of the crenulation cleavage and was recrystallized to coarser aragonite crystals during this event (Spadea 1982). Longitudinal strain estimates from fold profiles indicate a minimum finite ductile shortening of 36% during D_c (Figs. 7b & c).

NE-directed thrusting

Stretching lineations (L_{ca}) belonging to this deformation (D_c) trend roughly northeastwards (Fig. 3b) with a mean azimuth of 050. Crenulation lineations and S_b–S_c intersection lineations (L_{cb}) (Fig. 3c), cluster about a mean azimuth of 320. F_c fold hinges trend roughly northwestward (Fig. 3f) and mineral fibre lineations on D_c thrust surfaces, gouge fabrics and shear bands associated with D_c thrusts indicate transport directed roughly towards 025 with a less common group directed towards 075 (Fig. 8b).

D_d—*extension*

This deformation produced flattening, semi-ductile extension faults, mesoscopic and macroscopic foliation

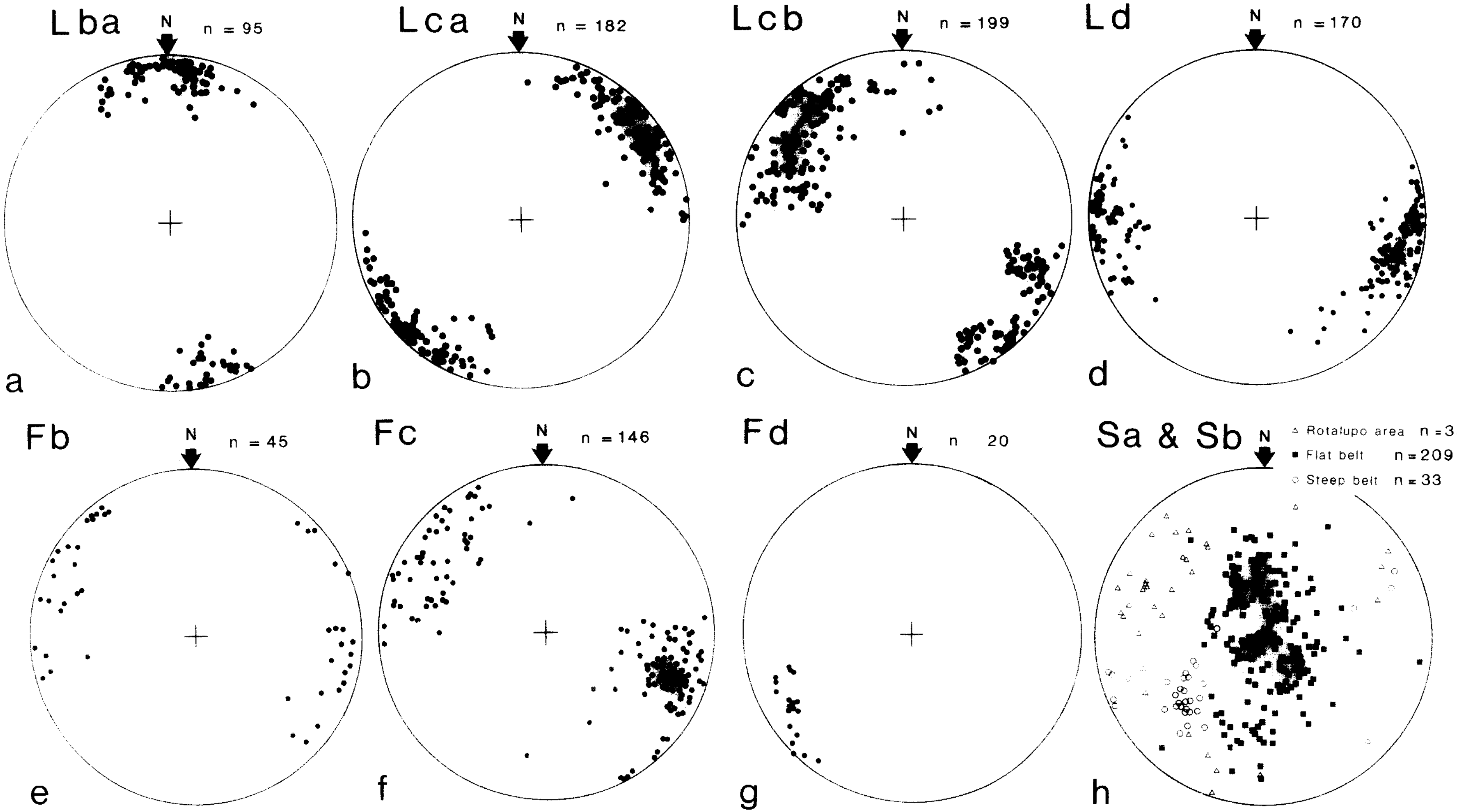


Fig. 3. Stereogram plots (equal-area, lower-hemisphere projection) of structural elements from the internal zone of the southern Apennines around San Severino Lucano. (a) NNE-trending stretching lineations associated with F_b folds, and fibre lineations on D_b thrust surfaces (L_{ba}). (b) NE-trending stretching lineations associated with F_c folds and fibre lineations on D_c thrust surfaces (L_{ca}). (c) NW-trending bedding–cleavage intersection lineations (L_{cb}) associated with F_c folds. (d) E-trending stretching lineations and fibre lineations on D_d semi-ductile extension fault surfaces (L_d). (e) Roughly WNW-trending F_b fold axes. (f) Roughly NW-trending F_c fold axes. (g) SW-trending F_d minor crenulation fold axes. (h) Poles to bedding (S_a) and transposition foliation (S_b) (undifferentiated). Division into flat belt and steep belt is the result of later (F_{c-e}) folding with regional fold axes (F_c) trending roughly northwestward.

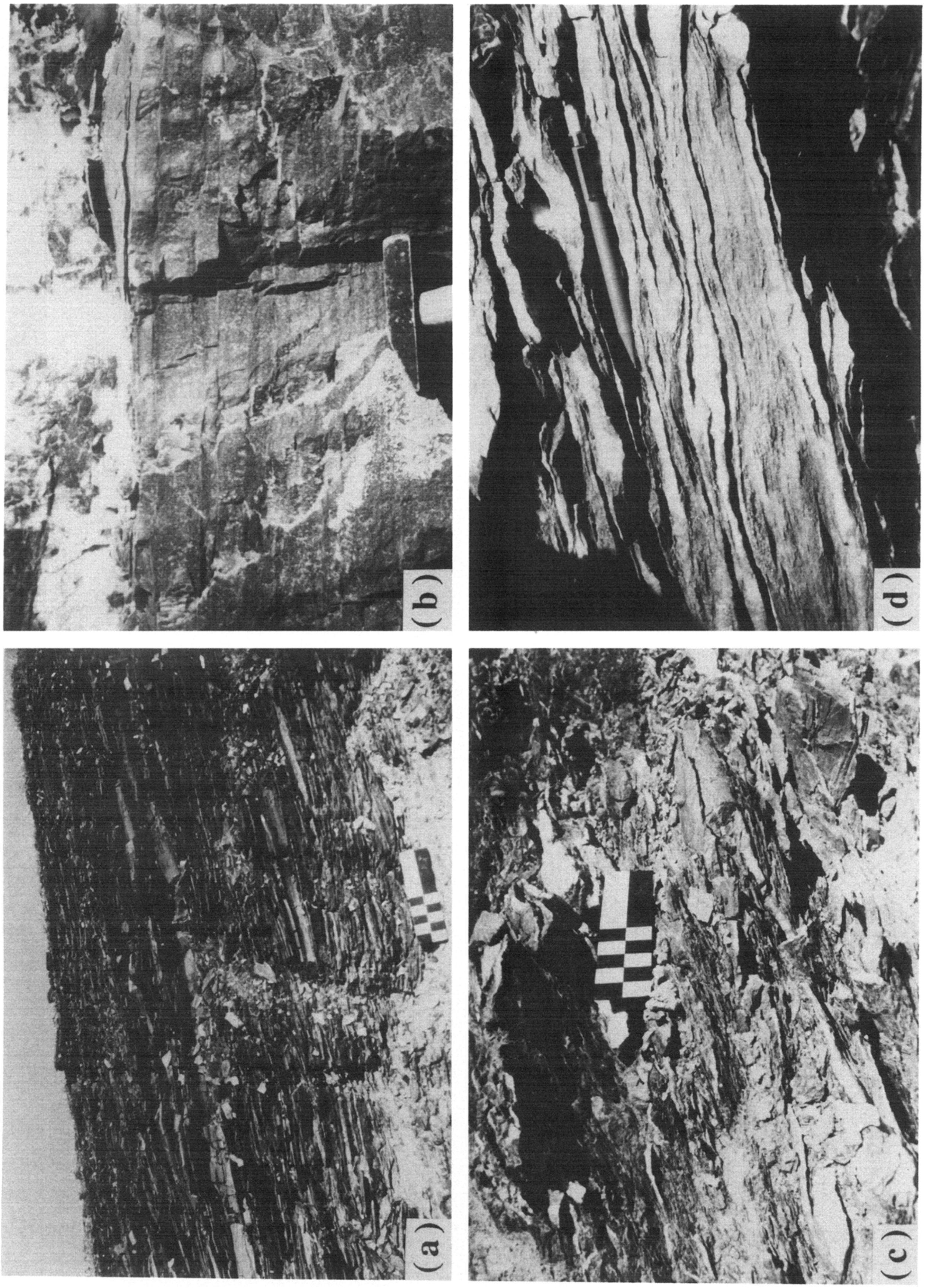
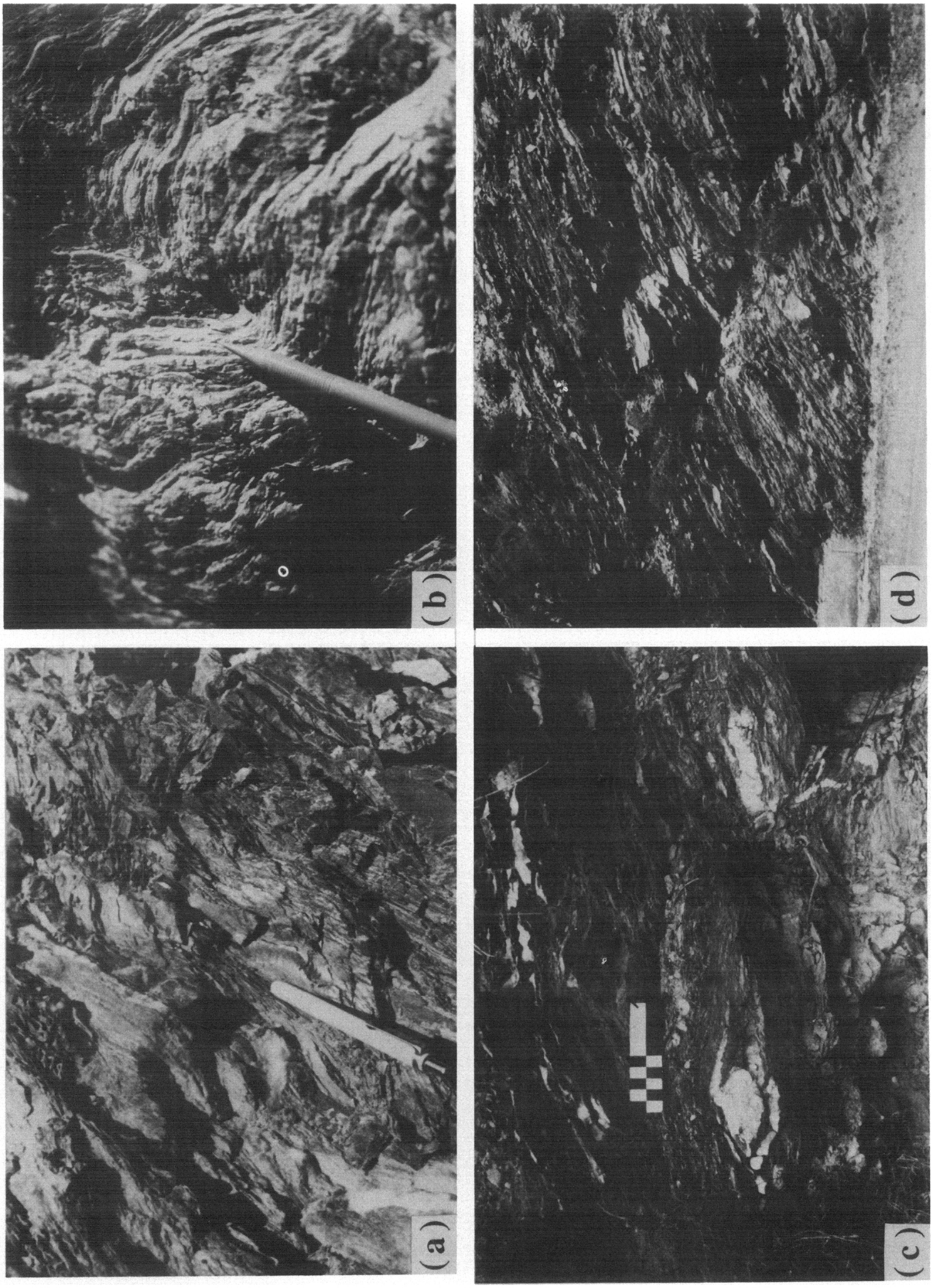


Fig. 4.


(a)
(b)
(c)
(d)

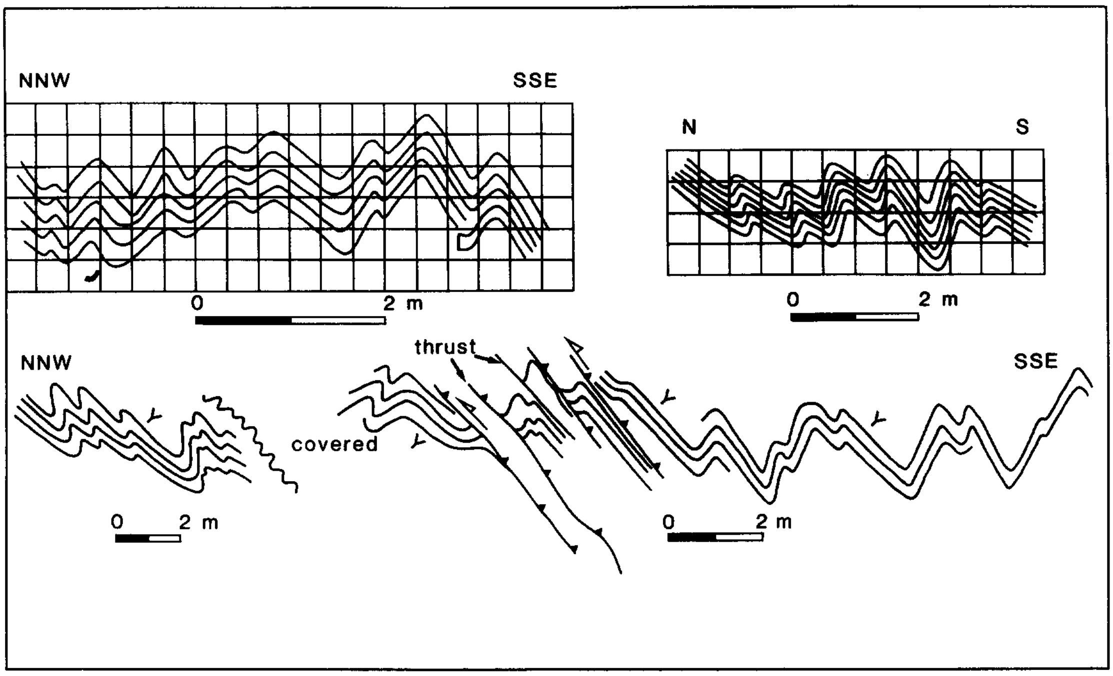


Fig. 6. Field sketches of D_c folds and thrusts. Folds within the grids are the outcrop view of the folds shown projected normal to their axes in Figs. 7(b) & (c). Overturning of stratigraphy (Y represents the younging direction—downward in this case) prior to D_c thrusting is attributed to the formation of major recumbent folds during D_b. The lower section is oblique to the transport direction of D_c thrusts which is northeastwards.

boudinage, and macroscopic boudinage of serpentinite and diorite blocks. Development of extensional crenulation cleavage is localized along contacts between metasediment and the serpentinite or diorite blocks. These contacts, originally related to thrust emplacement of the blocks, were reactivated during this phase. In places semi-ductile faults belonging to this generation cause omission of part of the stratigraphy. These faults also cut down stratigraphic section in the direction of transport (Figs. 5d and 7d). Major changes in metamorphic grade across the largest of the extensional fault zones (i.e. UON–LON contact) indicate that they are post-metamorphic.

Subhorizontal mylonitic shear zones adjacent to the main blocks of oceanic and continental material (Fig. 4d) rework and truncate earlier structures and are characterized by recrystallization of calcite and aragonite in metasedimentary rocks, and crystal-plastic deformation of serpentinite.

Mesoscopic boudinage in the internal zone of pre-existing (D_b and D_c) calcite–aragonite veins (Fig. 7d) is associated with recrystallization and development of new veins (D_d) cross-cutting D_c structures. D_d veins are subvertical and the long axes of new mineral fibres are sub-horizontal indicating a subvertical orientation of the local axis of maximum finite shortening. The maximum finite extension direction during this deformation (D_d) was sub-horizontal with a roughly E–W trend.

E-directed extension faulting

Kinematic indicators from extensional deformation (D_d) are commonly found near and along the contacts between the metasedimentary rocks and the large blocks of oceanic and continental material. Semi-ductile extension faults at the base of the San Severino Lucano serpentinite block (Fig. 2) consistently indicate a top-to-the-east sense of shear (Figs. 5d and 7d). Calcite mineral fibres, associated lineations along with gouge fabrics also indicate top-to-the-east shear. Lineations (L_d) along the base of the San Severino Lucano serpentinite block trend roughly towards 100 (Fig. 3d).

Mesoscopic semi-ductile extension faults offset the serpentinite–metasediment contact and indicate roughly top-to-the-east sense of shear during the extensional deformation. However, mineral fibre growth and lineations along this contact indicate shear movement of the serpentinite block towards 032 (Fig. 10). These lineations are likely to be products of the last movement along the contact and may not be related to the exten-

Fig. 4. (a) Slaty cleavage in pelite from the lower ophiolitic nappe of the internal zone (small scale bars = 2 cm, right of centre). Bedding is isoclinally folded with bedding-parallel cleavage developed. (b) Spaced cleavage in quartzarenite from the lower ophiolitic nappe of the internal zone (hammer head is 10 cm long). (c) F_b isoclinal fold within S_b transposition foliation (small scale bars = 2 cm). (d) Mylonite in calcarenite at the base of the San Severino Lucano serpentinite sheet. Note refolded F_c (?) isoclinal folds, lower ophiolitic nappe (pencil is 14 cm long).

Fig. 5. (a) S_c crenulations, Cozzo del Lupo near San Severino Lucano, lower ophiolitic nappe. Pencil orientated parallel to crenulation hinge (pencil is 14 cm long). (b) S_b spaced cleavage folded by F_c crenulations (pencil is 14 cm long). (c) Asymmetric calcite–aragonite vein rotated during D_c non-coaxial shear (small scale bars = 2 cm). (d) Semi-ductile extension fault within metasedimentary rocks beneath the San Severino Lucano serpentinite sheet (small scale bars = 2 cm).

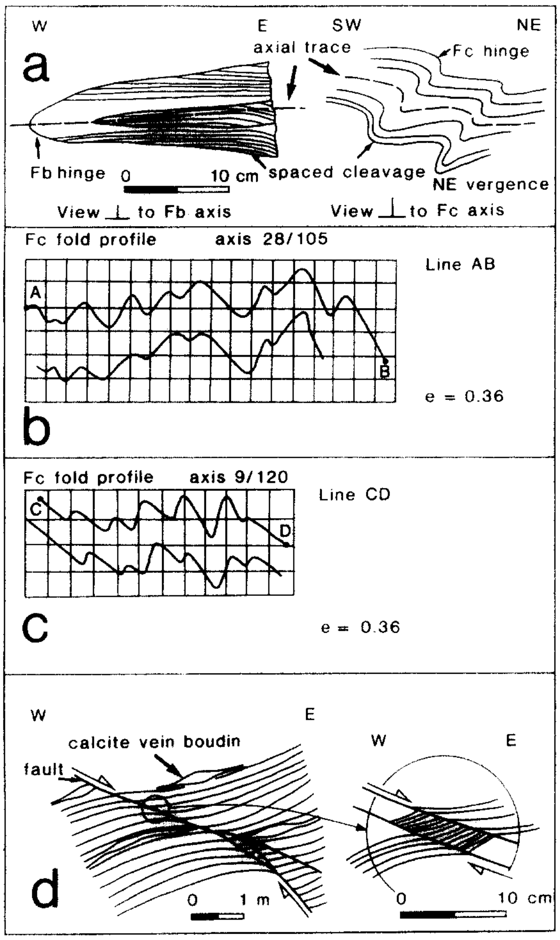


Fig. 7. Field sketches of mesoscopic structural elements from the internal zone. (a) Left—view perpendicular to axis of F_b fold with associated axial planar spaced cleavage. Right—view of same outcrop seen perpendicular to F_c fold axis showing folded axial trace of F_b fold and folded spaced cleavage. F_c folds verge northeastward. The scale is the same for both views. (b) Profile of F_c fold with fold axis orientated 28/105. Measurement of finite longitudinal strain indicates approximately 36% minimum D_c ductile shortening. (c) Profile of F_c fold with fold axis orientated 9/120. Measurement of finite longitudinal strain indicates approximately 36% minimum D_c ductile shortening. (d) Field sketch of D_d semi-ductile extension fault at the base of the San Severino Lucano serpentinite sheet (also shown in Fig. 5d). Sense of shear is given by the gouge fabric along the fault shown in the circular inset. Boudins of calcite veins related to layer parallel extension are truncated by the semi-ductile extension faults. The boudins may be related to the earlier stages of ductile extension during D_d or, alternatively D_b or D_c.

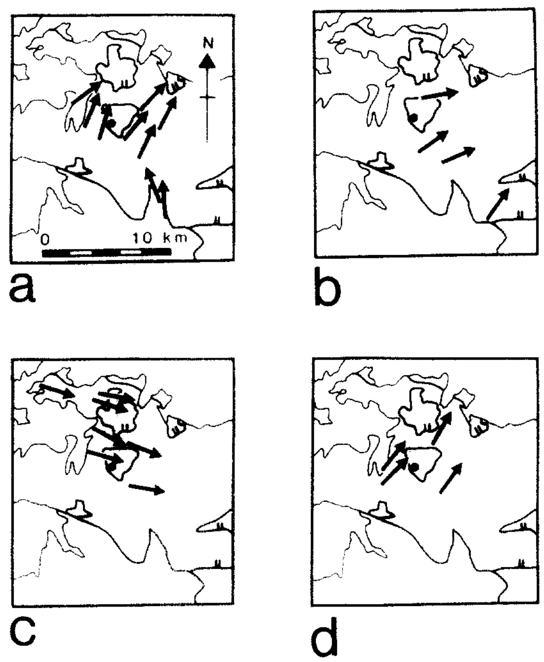


Fig. 8. Kinematic data from part of the internal zone of the southern Apennines around San Severino Lucano (dot). (a) Mineral fibre lineations (L_{ba}) are from early NNE-directed thrusts which deform Late Eocene metasedimentary rocks but are themselves deformed by later thrusts (e.g. D_c thrusts). (b) Mineral fibre lineations (L_{ca}) from NE-directed thrusts which deform Late Eocene metasedimentary rocks but are themselves truncated by later semi-ductile extension faults. (c) Mineral fibre lineations (L_d) from E-directed semi-ductile extension faults which deform Late Eocene metasedimentary rocks but are themselves folded by later (F_e) folds. (d) Mineral fibre lineations (L_e) from NE-directed thrusts which deform Late Eocene metasedimentary rocks and truncate all earlier structures. Each arrow represents the mean direction of at least 30 readings at one site and the direction is based on various kinematic indicators associated with each fault plane (e.g. shear bands). See Fig. 2 for location.

sion faulting: a distinct stretching lineation (L_d) is localized within the metasedimentary rock up to 50 m from the contact with a trend of 280 (Figs. 3d and 8c). The trend of the minor crenulation hinges is roughly towards 230 (Fig. 3g), in contrast to the roughly 320 trend of crenulations associated with D_c (L_{cb}; Fig. 3c).

D_e—*crenulations and brittle thrusts*

A second crenulation cleavage with reverse asymmetry is commonly associated with asymmetric folds which fold D_d extension faults and the contact bounding diorite and serpentinite blocks. These folds (F_e) verge roughly northeastwards and are associated with NE-directed brittle thrusts. F_e folds are generally downward facing. Folds of this style, orientation and generation are the only structures common to both the lower ophiolitic nappe and the upper ophiolitic nappe which suggests that the two nappes were juxtaposed prior to this deformation.

Late NE-directed thrusting

F_e folds (Fig. 3h) are roughly coaxial with F_c folds (Fig. 3f). Mineral fibre lineations and shear bands formed during D_e along the contact at the base of the Timpa Rotalupo diorite block indicate transport roughly towards 032 (Fig. 8d).

The lower ophiolitic nappe (internal zone) shares the late NE-directed thrusts with the upper ophiolitic nappe (external zone). Prior to D_d the upper and lower ophiolitic nappes experienced different deformation and metamorphic histories (see below). The two nappes were juxtaposed during the extensional phase (D_d). Structures related to the earliest deformation event recorded in the external zone are considered roughly time equivalent to D_e structures from the internal zone.

METAMORPHISM

Basaltic rocks occur as pillow lavas, pillow breccias, hyaloclastites and dykes in serpentinite, gabbro and

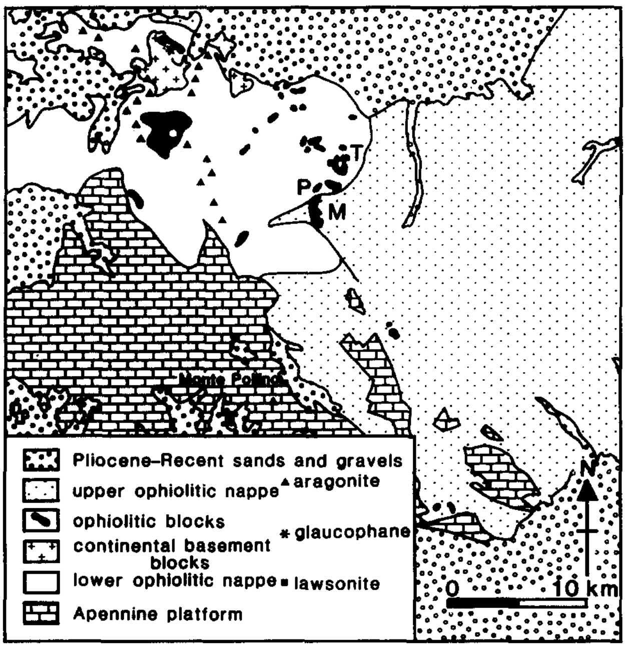


Fig. 9. Geological map of the internal zone (see Fig. 2 for location) showing distribution of high P/T-ratio mineral phases within the lower ophiolitic nappe. The upper ophiolitic nappe contains no high P/T-ratio metamorphism. M, Timpa delle Murge; P, Timpa Pietrasasso; T, Monte Tumbarino. Distribution of mineral phases after Spadea (1976, 1982).

amphibolitic diorite. Geochemically the basalts, gabbros and ultramafics in all ophiolitic nappes of the Liguride Complex have affinities with oceanic material and are geochemically similar to other Alpine and Mediterranean ophiolites (Beccaluva *et al.* 1982). Gabbro and associated basaltic intrusions show a multiphase retrograde metamorphism (amphibolite to greenschist facies; Lanzafame *et al.* 1979), whereas pillow lavas and associated dykes were only affected by spilitization (Beccaluva *et al.* 1982).

Rodingitization (Coleman 1967) within serpentinite blocks occurred in two separate events (Spadea 1982). The first event occurred prior to emplacement of the serpentinite bodies within the metasedimentary rocks. During this event, which probably took place in an oceanic environment, diopside and garnet were crystallized. The second event took place under higher pressures with pumpellyite as a major phase. This event was probably related to deformation within a subduction complex (Spadea 1982) and is interpreted here as a syn-accretion metamorphic event.

High P/T-ratio metamorphic assemblages are well preserved without extensive overprinting. This is particularly evident in the southern Apennines where aragonite is widespread in metasedimentary rocks of the lower ophiolitic nappe (Spadea 1976, Lanzafame *et al.* 1979, Spadea 1982) (Fig. 9). Two groups of high pressure rocks have been identified (Beccaluva *et al.* 1982). The first group includes metabasalt containing glaucophane, lawsonite, Na-pyroxenes, crossite, phengite, pumpellyite, albite and sphene. This group belongs to the blueschist facies of Turner (1981). The second group of high P/T-ratio rocks comprises metabasalt and metasedimentary cover with albite, lawsonite, chlorite, pumpellyite, epidote, sphene and rare blue amphibole and sodic pyroxenes characterizing the metabasalts. This group belongs to the lawsonite–albite–chlorite facies (Turner 1981).

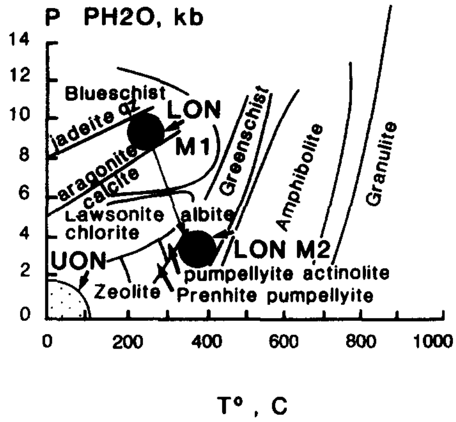


Fig. 10. Diagram showing the approximate position, in P–T space, of the early high P/T-ratio metamorphism (M_1, lawsonite–albite–chlorite facies to blueschist facies) in the lower ophiolitic nappe, and the later (M_2 greenschist facies) metamorphic overprint. P–T diagram after Turner (1981).

The mineral assemblages and textures of both these groups of rocks show a polyphase high P/T-ratio metamorphism overprinted by a lower pressure, greenschist facies metamorphism (Beccaluva *et al.* 1982) (Fig. 10). The later greenschist facies metamorphism is related here to uplift during a syn-extension metamorphic event. During high P/T-ratio metamorphism for the lawsonite–albite–chlorite facies, fluid pressures were between 3 and 4 kb and temperatures between 250 and 350°C. For the blueschist facies, fluid pressures ranged from 6 to 8 kb and temperatures were 350 ± 50°C (Beccaluva *et al.* 1982).

The age of the high P/T-ratio metamorphism is considered to be older than Late Albian time (Haccard *et al.*

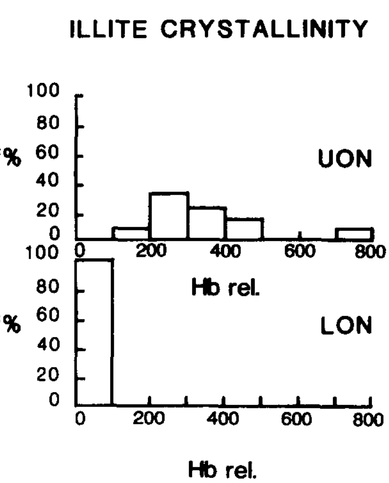


Fig. 11. Illite crystallinity data from the upper ophiolitic nappe (UON) and the lower ophiolitic nappe (LON). Hb rel. values from the upper ophiolitic nappe are widespread and significantly higher than Hb rel. values from the lower ophiolitic nappe indicating a major increase in metamorphic grade downward across the contact between the two nappes.

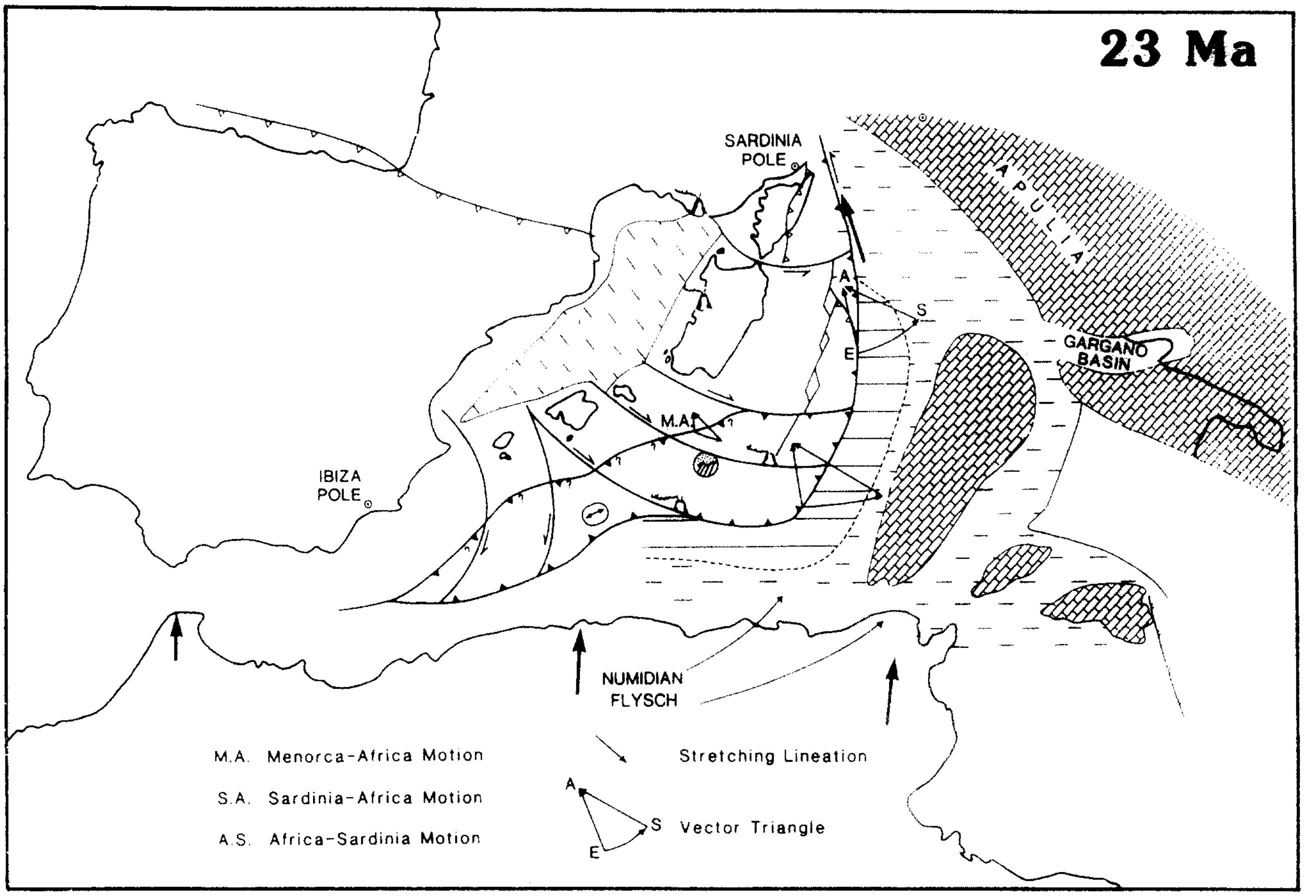


Fig. 12. Kinematic reconstruction depicting the evolution of part of the Western Mediterranean region. The frame of reconstruction shows the events occurring and the relations of tectonic elements at 23 Ma, earliest Miocene (Aquitanian). Diagonal bricks—Mesozoic carbonate platform. Solid horizontal lines—oceanic crust. Dashed diagonal lines—newly extended continental crust. Dashed horizontal lines—basin on continental crust. Line with filled barbs—active thrust (barb on overthrust side). Line with unfilled barbs—old thrust. Bold arrows—direction of Africa's motion relative to Europe. Double arrows within circles—extension. Volcanoes—volcanism. Long thin arrows—direction of Numidian flysch sedimentation. Parallelogram—pull-apart basin. Circle with dot in centre—pole of rotation (unlabelled if site of future pole). Dots above diagonal lines within circles—unconformity. Short straight arrow—stretching lineation. Triangles—vector triangles showing relative motion of various blocks with respect to Europe and Africa (after Dewey *et al.* 1989).

1972, Spadea 1982) based on the age of metasedimentary rocks surrounding the high-pressure blocks ('olistoliths', Spadea 1982). However, identification of Eocene fauna in the same high-pressure metasedimentary rocks from the lower ophiolitic nappe (Bouillin 1984), and evidence that the aforementioned blocks are tectonic slices and not olistoliths (Knott 1988) implies that at least part of the high P/T-ratio metamorphism is of post-Early Eocene age. A post-Early Eocene age for the high P/T-ratio metamorphism in the southern Apennines is favoured here and corroborates with other evidence for subduction at this time (Dewey *et al.* 1989) (Fig. 12).

STRUCTURAL HISTORY

Although the Liguride Complex does have some of the characteristics of an accretionary complex such as landward-steepening of thrusts, internally younging thrust sheets and externally younging tracts, these are not obvious at first glance. This is because the original geometry of the accretion-related structures has been severely modified by deformation related to extension during gravity-spreading, emplacement onto the African margin, collision of the Calabria–Sardinia–Corsica block with Adria and extension in the Tyrrhenian basin. Figure 13 shows a synthesis of the structural history of the Calabrian convergent margin from Eocene to Miocene time assimilating data presented here and in Dewey *et al.* (1989).

In Eocene time (Fig. 13a) NW-directed subduction of Neotethys occurred beneath Calabria and underplating of Neotethys oceanic–transitional crust. The future upper ophiolitic nappe (UON) (e.g. Timpa delle Murge ophiolitic block, Fig. 9) formed the ophiolitic forearc region of the upper plate and the lower ophiolitic nappe (LON) comprised a series of major, underplated duplexes derived from the lower plate.

Following and during underplating in Aquitanian time (Fig. 13b) gravity-driven crustal thinning occurred by normal faulting in the upper crust and ductile extensional flow and low-angle extension faulting at lower structural levels. Progressive thinning at the back of the wedge caused subsidence and deposition in the forearc

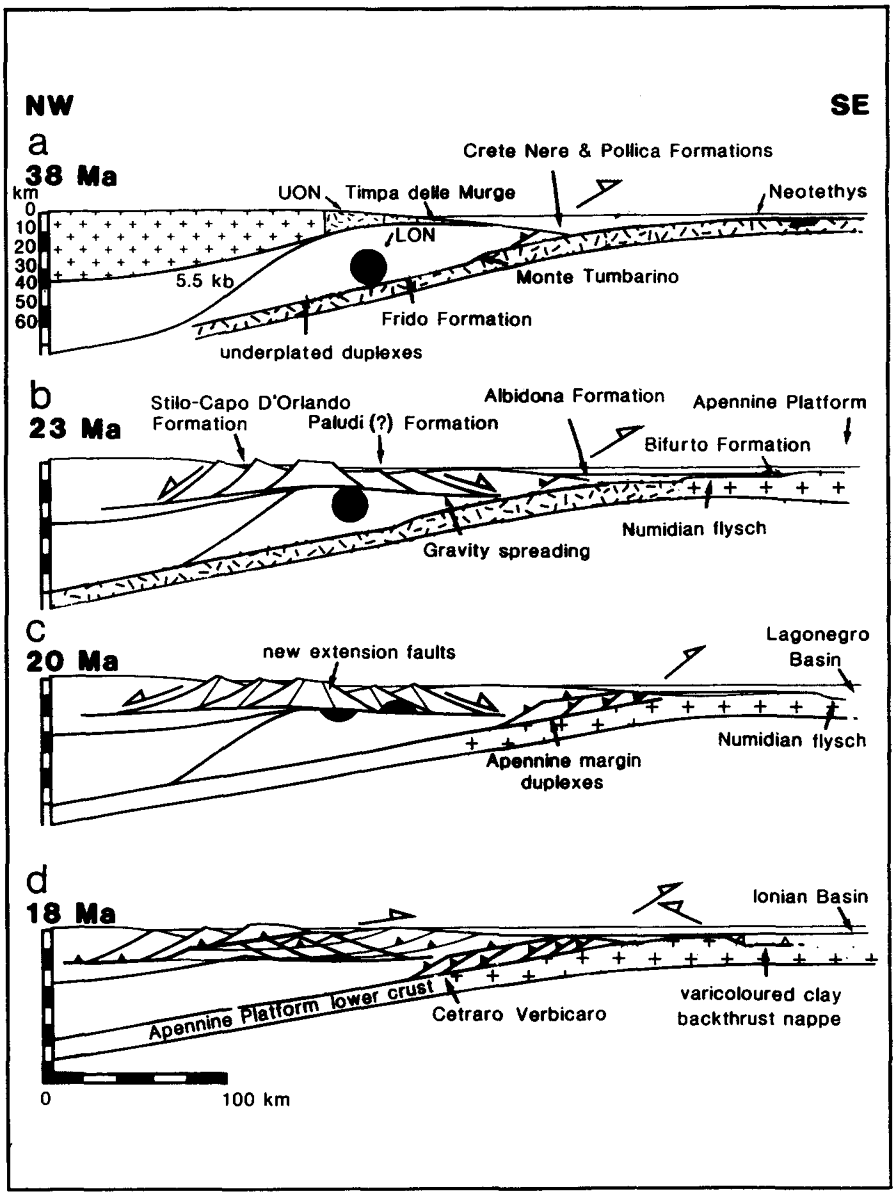


Fig. 13. Structural history of the Cenozoic convergent margin near Calabria from Eocene to Miocene time. Calabria lies on the southeast margin of the European plate. (a) Eocene time. (b) Aquitanian time. (c) Early Burdigalian time. (d) Late Burdigalian time. The cross-sections are area balanced and lengths are constrained by a plate reconstruction of the Western Mediterranean (Dewey *et al.* 1989). See text for elaboration.

region (Stilo-Capo d'Orlando Formation and Paludi (?) Formation). The lower ophiolitic nappe was progressively uplifted by continued underplating and lay in close proximity to the upper ophiolitic nappe.

In early Burdigalian time (Fig. 13c) a system of new extension faults developed and extension and thinning continued in the internal zone. The Apennine Platform margin (Cetraro and Verbicaro thrust sheets) was progressively incorporated into the wedge.

During collision (*sensu lato*) in late Burdigalian time (Fig. 13d) a wave of out-of-sequence thrusts passed through the wedge placing basement thrust slices onto the fore-arc basin sediments and reimbricating the earlier accreted rocks of the Liguride Complex. The position at this time of the varicoloured clay backthrust nappe was within the Lagonegro-Ionian Basin. During later thrusting a thin flake from the wedge was emplaced onto the African margin. The present-day Calabride Complex and Liguride Complex in the Calabrian arc represent the most internal and deepest part of the wedge exposed. The Liguride Complex in the southern Apennines represents a more shallow and external part of the wedge. The two regions have been displaced left-laterally by several tens (?or even hundreds) of kilometres along the southern Apennines lateral shear zone (Knott & Turco 1991). Below, some of the events are discussed in more detail in the sequence in which they occurred.

Stratal disruption (D_a)

In the internal zone of the Liguride Complex sedimentary stratal disruption has not been recognized. Discrete zones of tectonic stratal disruption do occur, however, and these are usually localized along the contacts between the blocks of oceanic and continental affinities and the surrounding metasedimentary rocks. These zones generally truncate structures formed during the first deformations (D_b and D_c). Stratal disruption along these zones occurred during two distinct events, the first under high-grade, ductile conditions with the development of augen-gneiss envelopes around large (100 m) granitic inclusions and pervasive recrystallization of quartz. These structures are overprinted by a second deformation involving brittle disruption with localized cataclasis of the blocks and formation of slickensides and a thick, powdery serpentinite gouge under high P/T-ratio metamorphic conditions.

The ductile deformation may be related to: (1) ductile extension during Mesozoic rifting; (2) thrusting during subduction accretion; or (3) extension within the wedge following major underplating. The first interpretation is favoured as the early, high-grade (amphibolite) ductile deformation is only developed in the material of continental affinity and therefore pre-dates incorporation of these blocks within the metasediment. Conditions attendant during this high-grade metamorphism were fluid pressures of roughly 6–8 kb and temperatures of roughly 550–650°C that occurred at mid- to lower crustal levels (Spadea 1982). The overprinting deformation may be related to cataclasis along semi-ductile extension faults (D_d) or along thrust planes active during subduction (D_b and D_c) or collision (D_e and later deformations).

Underthrusting (D_b)

Widespread lack of stratigraphic coherence in metasedimentary rocks of the internal zone, even at outcrop-scale, is attributed here to pervasive transposition during the first deformation affecting the metasedimentary rocks (D_b). Transposition generally occurs in regions affected by strong non-coaxial deformation and is accompanied by elements such as pervasive cleavage and intrafolial folds, both of which are present in the Liguride Complex (Figs. 4a & c). Flat-lying transposition foliation and the sub-horizontal attitude of F_b fold axes implies that transposition occurred in a shallowly dipping major shear zone, which is to be expected between underriding and overriding plates in a subduction zone (Fig. 13). Syntransposition veins bearing aragonite (Spadea 1976) suggest that shearing took place at depths exceeding 10 km. Northeast-directed thrusting occurred in the early stages of underthrusting (Fig. 13), probably during Late Eocene to Early Oligocene time. This matches well with kinematic data derived from plate motions at this time (Dewey *et al.* 1989: slip vectors have been restored from their present-day orientations in Fig. 12).

Underplating (D_c)

Underplating in accretionary wedges is considered to be accomplished by duplex formation and accretion of oceanic crust to the base of the wedge at depths often exceeding 10 km (e.g. Silver *et al.* 1985, Sample & Fisher 1986). Thrust faults and folds belonging to the second generation of structures in the internal zone of the Liguride Complex (D_c) have duplex geometries (Fig. 6) (Boyer & Elliott 1982, Butler 1983). Slaty cleavage development in pelite and spaced cleavage in psammite accompanied thrusting.

Spaced cleavage in quartz–arenite (Fig. 4b), the product of diffusive mass transfer, indicates burial depths exceeding 4 km but not greater than roughly 16 km in the southern Apennines, as the quartz–arenite does not contain dynamically recrystallized quartz grains which normally occur at greater depths (Hobbs *et al.* 1976). In the most internal zones very low temperatures and a depth of burial exceeding 15 km (Turner 1981) are indicated by the presence of lawsonite, the local presence of glaucophane and crossite (and jadeite and quartz in northwest Calabria) in metabasic rocks, and of aragonite in metasediments (Spadea 1982). Illite crystallinity data (Fig. 11) also favour low temperatures (*ca* 200°C) in the internal zone during underplating.

The presence of brittle faults during D_c thrusting and the absence of mylonitic shear zones further implies that temperatures (and/or shear strains) were relatively low. Folding within thrust sheets and the paucity of thrust-parallel packages of bedding indicate that duplex formation was accompanied by significant ductile deformation. However this is due to the predominance of calcareous pelitic and psammitic sedimentary rocks which have a lower plastic yield strength than quartzose lithologies (Hobbs *et al.* 1976). Deformation took place at or near the brittle–ductile transition zone. This was probably at a depth exceeding 15 km given the low geothermal gradient suggested by metamorphic mineral assemblages, illite crystallinity, a rapid accretion rate (Dewey *et al.* 1989), and subduction of old (Late Jurassic), cool, oceanic lithosphere.

Progressive NE-directed thrusting (Fig. 8b) matches well the kinematic pattern of this region represented in the plate reconstruction (Fig. 12).

Crustal thinning (D_d)

In the internal zone the two generations of contractional structures, D_b and D_c, related here to underplating and underthrusting, are truncated by semi-ductile extension faults. The most easily recognizable semi-ductile extension faults occur at the base of the large blocks of oceanic and continental material and along the contact between the lower and upper ophiolitic nappes. Minor extension faults occur at outcrop-scale in the metasedimentary rocks both within and outside the shear zones at the base of the blocks.

Metasedimentary rocks along the major extension faults which envelope both the serpentinite blocks and blocks of continental affinity are mylonitic. The presence of mylonites along fault zones during this deformation, when prior to this brittle faulting dominated, indicates either, a relative rise in temperature and therefore geothermal gradient, or an increase in strain and/or strain rate. An increase in temperature and geothermal gradient is to be expected during extension and crustal thinning. A general greenschist facies overprint of the high pressure assemblages and transformation of aragonite to calcite in the internal zone (Spadea 1976, 1982) (Fig. 10) further support the inference that rising tem-

Not only metamorphic and stratigraphic arguments favour regional extension and thinning, but also the style of structures above and below the main extension contact. Above this contact in the upper ophiolitic nappe, a simple deformation of late ESE-directed contractional imbricate fans is the main structural style. Below the contact, in the lower ophiolitic nappe, polyphase deformation including transposition, isoclinal folding and thrust duplexes occurs. This implies that regional extension led to the emplacement of relatively undeformed oceanic crust, probably from the fore-arc region, onto previously underplated, multiply-deformed high *P/T*-ratio rocks originating from deeper within the wedge (see also Platt 1986, Wallis *et al.* 1993) (Figs. 13a & b). The ESE-directed imbrication post-dates the extensional contact.

East-directed extension does not correlate with the plate boundary kinematics predicted by the reconstruction (Fig. 12). This may be due either to plate motions different to those shown in the reconstruction, or to development of structures by body forces within the wedge rather than plate boundary forces. The latter explanation is preferred as NE-directed thrusting resumes after the extensional episode, implying plate boundary forces were roughly constant in orientation before and after extension. Further evidence of extension in the upper plate at this time are the syntectonic terrigenous sediments of the Oligocene to Miocene Capo d'Orlando Formation, which filled extensional basins that developed on top of the wedge during this episode of crustal thinning (Fig. 13b).

Collision (sensu lato) (D_e)

Structures formed during collision (*sensu lato*) are related to frontal accretion, out-of-sequence thrusting and backthrusting. Frontal accretion structures occur mainly in the external zone where sedimentary rocks do not show the effects of metamorphism. Structural styles here are dominated by recumbent isoclinal folds lacking penetrative cleavage, ramp–flat thrust geometries and imbricate fans. In the higher thrust sheets there are no major duplexes, instead there is widespread evidence for thrust-tip emergence shown by wildflysch deposition (Oligocene to Miocene Albidona Formation, and Bifurto Formation: Fig. 13b).

Out-of-sequence thrusts in the Liguride Complex formed in two ways. First, they probably initiated during collision of the Liguride Complex accretionary wedge with Adria. Late brittle thrusts within the internal zone (D_e) are related here to a wave of contractional deformation passing through the wedge during collision (*sensu lato*).

Second, some out-of-sequence thrusts in the internal zone may be due to wedge dynamics and therefore formed independently of collisional processes. Accretion of the Apennine Platform (Cetraro and Verbicaro thrust sheets) (Fig. 13c) in Miocene time would have increased the length of the wedge and thus reduced the angle of taper. Out-of-sequence thrusting at the front of the wedge would have facilitated the uplift of this area thus returning the wedge to an equilibrium profile (Fig. 13d). Emplacement of the varicoloured clay backthrust nappe would also have contributed to uplift at the front of the wedge (Fig. 13d).

DISCUSSION AND CONCLUSIONS

The Liguride Complex of southern Italy is the remains of a Cretaceous to Paleogene, roughly E-directed accretionary wedge which was emplaced onto the African margin during Early Miocene time. Transport directions of thrust sheets and nappes in the internal zone, based on kinematic indicators, are roughly towards the present day northeast. Late Cenozoic block rotations of up to 60° counterclockwise about a vertical axis have occurred in the southern Apennines during the opening of the Tyrrhenian Sea (Knott & Turco 1991). Kinematic data from the internal zone were collected from an area affected by this deformation. The amount of Late Cenozoic rotation for the sites in this study is unknown, but the general match with the plate reconstruction indicates that it is less than 30°, at least for this part of the southern Apennines.

The northeast transport direction is in conflict with a southwest ('European') vergence in the Liguride Complex that is based on a comparison with the Alps (Haccard *et al.* 1972, Alvarez 1976, Amodio-Morelli *et al.* 1976) and structural data from Monte Reventino, Calabria (Fig. 1) (Alvarez 1978) indicating a southwest transport direction. Alternative interpretations of the southwest transport direction are backthrusting or, more likely, hinterland-directed extension faulting due to gravity spreading of an overthickened wedge (see also northward transport direction in Aspromonte; Platt & Compagnoni 1990). The main transport direction during Late Eocene to Oligocene thrusting in the Liguride Complex was towards the present-day northeast (see also Knott 1987, 1988 and Dietrich 1988).

The detailed structure and deformation history of the internal zone of the Liguride Complex is considered here to be the product of extensive underthrusting followed by underplating of already lithified sediments, ophiolitic rocks and continental basement to the base of an accretionary wedge to depths in places exceeding 30 km. Subduction initiation probably occurred near the European continental margin, given the inclusion of continental basement slices in the upper part of the LON, and the occurrence of quartz–arenitic turbidites of continental provenance as part of the sedimentary cover to the ophiolitic thrust slices (Knott 1987). The contact between the lower ophiolitic nappe (high *P/T*-ratio metamorphism) and the upper ophiolitic nappe (unmetamorphosed) is an extensional fault zone that was active during Late Paleogene gravity-driven extension when the accretionary wedge was overthickened by massive underplating of the Neotethys carapace. Transport directions during this event do not match slip vectors derived from local plate motions, suggesting that this part of the orogen was decoupled from the moving plates and that the gravitational potential of the wedge dominated the kinematics of deformation. Although the data presented come from a poorly exposed and relatively small region, evidence for crustal thickening followed by pre-collision extension from elsewhere in the Calabrian arc (Platt & Compagnoni 1990) indicates that these phenomena are of regional extent.

The ophiolitic blocks in the internal zone are tectonic slices (Knott 1988), rather than olistoliths as was previously suggested. In previous paleogeographic reconstructions (e.g. Bonardi *et al.* 1982) a major part of the shortening since Eocene time was considered to have been taken up by imbrication of the African continental margin. This hypothesis does not fit available estimates of shortening in the African margin during this time (Malinverno & Ryan 1986 and references therein). It is proposed here that most of the shortening (roughly 300 km) (Fig. 13) between Calabria (Europe) and Adria (Africa) during the rotation of the Calabria–Sardinia–Corsica block (Oligocene to Miocene time) was taken up by subduction of oceanic lithosphere, underplating and frontal accretion of the oceanic crust of Neotethys to form an accretionary wedge, the Liguride Complex.

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