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# The Ligurian Units of Western Tuscany (Northern Apennines): insight on the influence of pre-existing weakness zones during ocean closure

Giuseppe Nirta [a] , Gianfranco Principi [a] & Paola Vannucchi [a]
[a] Dipartimento di Scienze della Terra , Università di Firenze , Via G. la Pira, 4, 50121 , Firenze , Italia
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# The Ligurian Units of Western Tuscany (Northern Apennines): insight on the influence of pre-existing weakness zones during ocean closure

Giuseppe Nirta*, Gianfranco Principi, Paola Vannucchi

*Dipartimento di Scienze della Terra, Università di Firenze, Via G. la Pira, 4 – 50121 Firenze, Italia.*

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**Abstract**

Most of the tectonic units cropping out in Western Tuscany are fragments of the Jurassic oceanic crust, ophiolitic successions, overlaid diachronously by Upper Cretaceous-middle Eocene carbonate and siliciclastic flysch successions with their Cenomanian-lower Eocene shaly-calcareous basal complexes. These units, so called Ligurian, have been emplaced during the closure of the Ligurian-Piedmont Ocean.

Ophiolite bearing debris flows are common in the flysch basins and their relationship with ophiolitic tectonic slices points to a strong relation between tectonics and sedimentation from the early compressive events of the Late Cretaceous. The tectonic activity reflects in a rough morphology of the ocean floor. It progressively influences the distribution and sedimentology of the turbidites. During middle Eocene this relationship begun very important and a paleogeographic reconstruction with prominent linear ophiolitic reliefs that bounded some turbiditic basins can be done.

In our reconstruction the sedimentary and structural evolution can be framed in the context of strain partitioning, developed during the ocean closure, between subduction processes and ancient weakness zones crosscutting both the ocean and the Adria continental margin and reactivated in compressive regime. These weakness zones can be interpreted as transform faults of the Ligurian-Piedmont Ocean with prolongations in the Adria passive margin.

The weakness zones crosscut the oceanic lithosphere and the Adria continental margin and interfered with the subduction processes. The activity of the weakness zones is reflected in the Ligurian Units architecture where two main structural strike trends of thrusts and folds axial planes occur. The first trend is WSW-ENE oriented and it is connected with the reactivation of the weaknesses zones. This first orientation developed progressively from Late Cretaceous to Pliocene, from oceanic to ensialic convergence (D1, D2, and D4 deformation phases). The second trend is NNE-SSW oriented and is related to the late Eocene continental collision and the subsequent translation to the NE of the oceanic units onto the Adria continental margin (D3 deformation phase).



*Keywords:* lithospheric lineaments, fracture zone, oblique convergence, strain partitioning, transpression, ligurian units, northern apennines

## 1. Introduction

In the Middle Jurassic the Ligurian-Piedmont Ocean, a part of the Western Tethys, started to open between the European and Adriatic plates as a NE prolongation of the Central Atlantic basin [1-7]. The Western Tethys was characterized by extensive exposure of mantle ultramafics and ophiolitic breccias, while effusive products were minor

* Corresponding author.
*Tel:* +39 055 2757524 - *Fax:* +39 055 218628
*E-mail address:* giuseppe.nirta@geo.unifi.it

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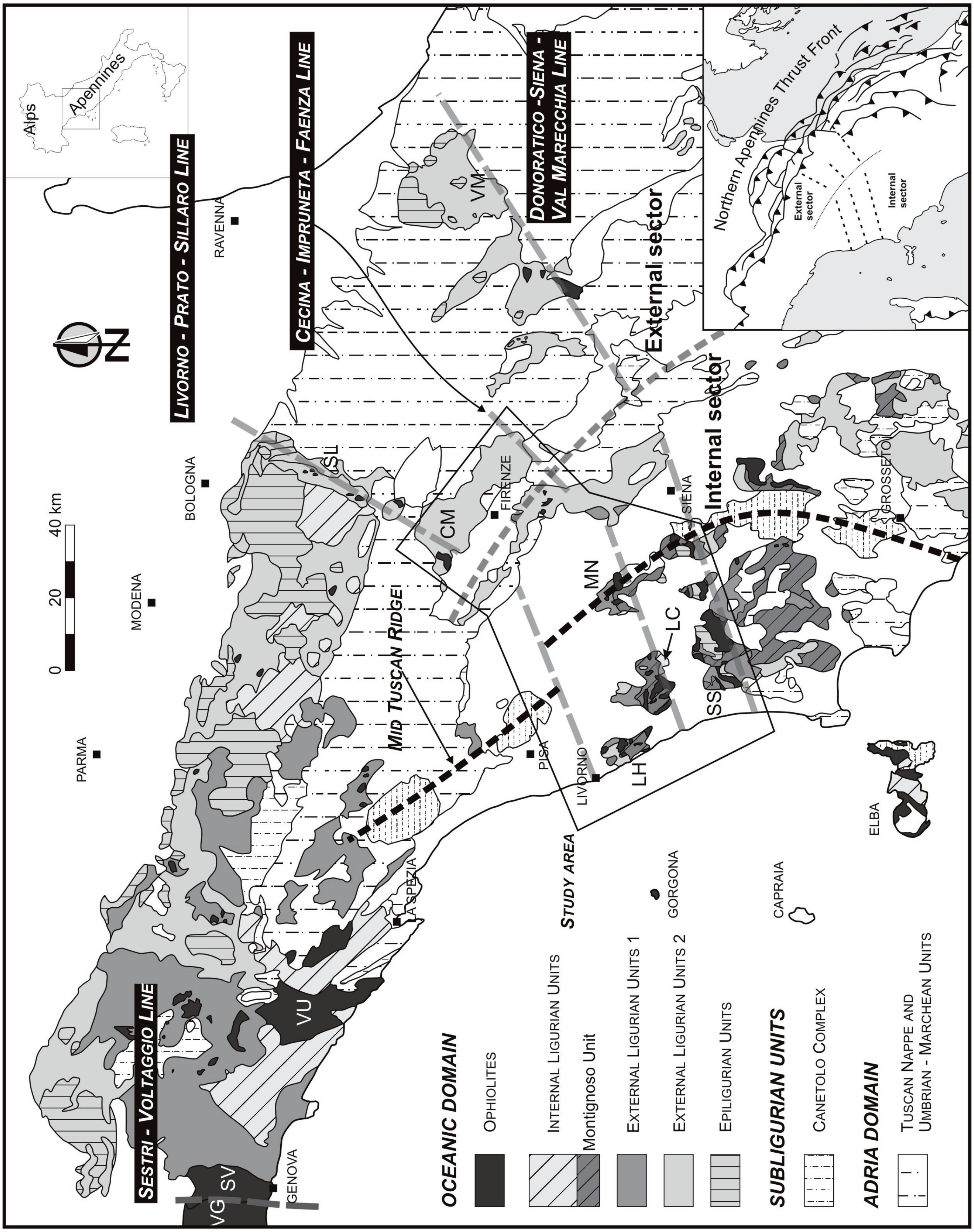

Alps
Apennines
N
0 20 40 km
Livorno - Prato - Sillaro Line
Cecina - Impruneta - Faenza Line
Donoratico - Siena - Val Marecchia Line
Sestri - Voltaggio Line
Mid Tuscan Ridge
External sector
Internal sector
Northern Apennines Thrust Front
External sector
Internal sector
PARMA
MODENA
BOLOGNA
RAVENNA
FIRENZE
SIENA
GROSSETO
PISA
LIVORNO
LA SPEZIA
GENOVA
GORGONA
CAPRAIA
ELBA
VG
SV
VU
SL
CM
VM
MN
LC
LH
SS
Study area
Oceanic Domain
Ophiolites
Internal Ligurian Units
Montignoso Unit
External Ligurian Units 1
External Ligurian Units 2
Epiligurian Units
Subligurian Units
Canetolo Complex
Adria Domain
Tuscan Nappe and Umbrian - Marchean Units

**Fig. 1** – Tectonic sketch map of the Northern Apennines.
The Northern Apennine Thrust Front is traced according to [35].
Calvana mountains area **(CM)**, Lupicaia Creek Tectonic Unit **(LC)**, Livorno hills area **(LH)**, Montaione area **(MN)**, Sillaro area **(SL)**, Sassa area **(SS)**, Sestri-Voltaggio area **(SV)**, Voltri Group **(VG)**, Val Marecchia area **(VM)**, Vara Tectonic Unit **(VU)**.

[6, 8, 9]. These characters have been interpreted as indicative of a slow spreading ocean dominated by amagmatic processes [10-13]. In this scenario, the extension that brought to the growth of the ocean has been thought to be accommodated by the development of significant discontinuities, as transform faults, cutting through the lithosphere [14, 15]. During the Late Cretaceous the ocean opening had stopped and extension was replaced by contraction. The switch from the extensional to the compressional regime is inferred to have caused reactivation of the transform faults as they represented weakness zones [16]. The sedimentation of high amounts of turbidites deposited on top of the deep water shaly-calcareous successions testifies continental uplift triggered by the beginning of subduction processes [17]. Stratigraphic, sedimentologic and petrographic studies carried out in the Northern Apennines on the turbidites reveal heterogeneous sources for the terrigenous material [18-30] and these characters suggest that they represent different sedimentation areas of the ocean that followed, also, distinct paths as they were involved in the subduction complex. In fact, simple paleogeographic reconstructions, with cylindrical back-deformation of the oceanic units in a direction orthogonal to the chain axis, bring several problems due to stratigraphic incongruence. Indeed a pure compressional model for the shortening of the Ligurian-Piedmont Ocean fails to explain the observed juxtaposition of the Ligurian Units, that, as it will be shown in this paper, have very different tectono-stratigraphic evolution. In our kinematic model, the emplacement of the Ligurian Units in Western Tuscany requires a significant transcurrent component active during the build up of the Ligurian subduction complex [31]. This hypothesis agrees with the kinematic reconstruction of the Late Cretaceous oblique convergence between Africa and Eurasia [5, 32].

## 2. Geological setting

The Northern Apennines orogenic belt is formed by a pile of thrust units verging to the NE-NNE - i.e. toward the Adria foreland [33]. The orogenic front is convex toward the NE (Fig. 1) with directions that vary from WNW-ESE in the northern sector (Ligurian-Emilian Apennines) to NNW-SSE in the southern sector (Umbrian-Marchean Apennines) [34, 35]. However the convex shape is not involving the whole belt, but it is most evident in the external sector, where compression is presently active. In fact, the SW sector, apart from the virgation correspondent to the Mid-Tuscan-Ridge (MTR), maintains straight directions of principal structures (Fig. 1). Because of this different geometry, that has been developed starting from the late Miocene, the Cretaceous-Eocene structures described in the present paper do not maintain a constant trend from the internal to the external sector.

The Northern Apennines Thrust Front is well defined by seismic reflection profiles and well data in the Po Plain subsurface and in the Adriatic Sea and is characterized by several undulations convex toward the foreland separated by transfer zones [34-36].

Within the Apennines pile the lower units are those deposited on the continental Adria crust, while the top units are the so called "Ligurian Units" representing sediments and portions of igneous oceanic crust. The Ligurian Units have been distinguished in Internal and External Ligurian Units on the basis of their lithostratigraphic and tectonic setting [37] (Fig.2).

The Internal Ligurian (IL) Units, generally located to the west of the chain, are formed by ophiolitic successions with a basement represented by serpentinized mantle peridotites intruded by gabbroic bodies and topped by both N-MORB lavas [38 and reference therein] and sedimentary ophiolitic breccias [39]. The extensive exposure of ultramafic rocks and ophiolitic breccias and the poor development of effusive products imply slow spreading and amagmatic processes with development of rough seafloor topography [13]. This basement is covered by the Jurassic-Lower Cretaceous pelagic sequence that includes cherts - Mt Alpe Chert - Limestones - Calpionella Limestone – and shales - Palombini Shale [6]. The mantle peridotites and gabbroic rocks have been exposed on the ocean floor as denudated ophiolitic basement. This exposure was associated with a widespread, high temperature (600°-750°C) ocean floor metamorphism that produced ductile deformations as isoclinal folds and mylonitic bands [39, 40]. The ductile structures have been later cut by fractures associated to sea water circulation and amphibolite to greenschist facies overprint [41, 42].

In the "classic" IL unit succession cropping in the Ligurian-Emilian Apennines, the ophiolitic basement and related pelagic sediments are overlaid by siliciclastic turbidites of Campanian-early Paleocene age (Manganesiferi Shale, Mt Verzi Marl, Zonati Shale, Mt Gottero Sandstone [43]).

The External Ligurian (EL) Units are most common on the eastern sector of the Apennines and in Western Tuscany where they are placed beneath the IL Units. The EL Units generally lack the oceanic basement and consist mainly of detached pre-flysch units and calcareous-marly turbidites called Helminthoid Flysches. A trace of the ophiolitic basement is preserved as olistoliths and olistostomes [9] within the flysch sequences. The detachment is regionally cutting through pelagic, pre-flysch units and there are evidences of the detachment occurring before accretion on the Adria passive margin [16]. The pre-flysch-units, also called basal complexes, consist of clay-rich pelagic and hemipelagic successions of Early Cretaceous - early Eocene age. The Helminthoid Flysches of the EL units are progressively younger moving both from western Tuscany, Campanian-middle Eocene ages, to the Liguria-Emilia Apennines, Maastrichtian-middle Eocene, and from west to east where their age of inception is early Eocene [3, 17, 28, 44, 45].

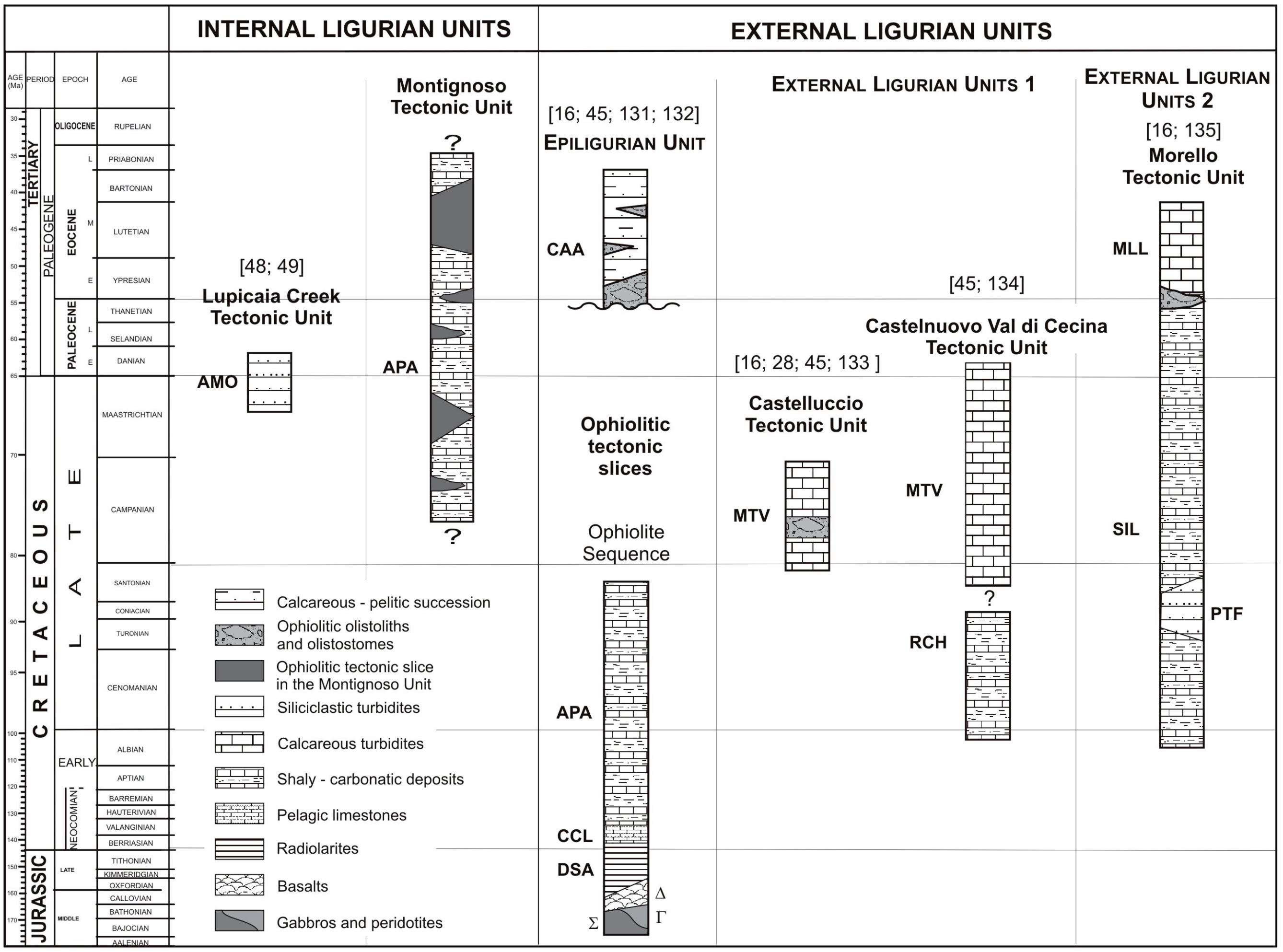



**Fig. 2** – Schematic stratigraphic columns and age of the Ligurian Units of Western Tuscany. **AMO:** Montecatini Sandstone; **APA:** Palombini Shale; **CAA:** Lanciaia Formation; **CCL:** Calpionella Limestone; **DSA:** Monte Alpe Chert; **Δ:** Basalts; **Γ:** Gabbros; **Σ:** Serpentinites; **MTV:** Monteverdi Marittimo Formation; **RCH:** Poggio Rocchino Formation; **MLL:** Monte Morello Formation; **SIL:** Sillano Formation; **PTF:** Pietraforte Formation.

The Ligurian Units are unconformably covered by slope and apron sediments called Epiligurian Successions. They either postdate many of the tectonic contacts present in the Ligurian Units or are deformed together during their NE thrusting on top of the Adria continental margin [46]. Also the Epiligurian Successions have a younging direction to the NE, so that they are late Paleocene-early Eocene in western Tuscany, early Eocene in the Liguria-Emilia Apennines, and early Miocene in the easternmost sector of the chain [16; 47 with reference therein].

## 3. Lithostratigraphy of the Ligurian Units in Western Tuscany

In Western Tuscany the IL units are represented by the Lupicaia Creek Tectonic Unit and the Montignoso Tectonic Unit (Table 1). The Lupicaia Creek Tectonic Unit is formed by isolated outcrops of Maastrichtian-Paleocene sandstone, related to the IL units flysches of the Ligurian-Emilian Apennines, and here called Montecatini Sandstone (Fig. 2). The Montignoso Tectonic Unit is formed by the Lower Cretaceous clay-rich formation of the Palombini Shale. This latter unit is extensively cropping out in Western Tuscany, while it is totally absent in other sectors of the Northern Apennines (Fig. 1).

The EL units can be divided in three sub-groups: EL1 units, which includes Campanian to Danian carbonatic turbidites deposited in the western portion of the Ligurian basin and associated with middle Campanian ophiolitic olistoliths and breccias; EL2 units, characterized by lower-middle Eocene carbonatic turbidites and upper Paleocene–lower Eocene ophiolitic debris in the basal complexes; and ophiolitic units, structured as a tectonic pile of thrust sheets of variable sizes lying on top of the EL1 units and unconformably covered by the Epiligurian Successions, as the Lanciaia Formation.

### *3.1 Internal Ligurian Units*

**Lupicaia Creek Tectonic Unit.** The Montecatini Sandstone of the Lupicaia Creek Tectonic Unit [16, 48] occupies the topmost geometric position in the Ligurian tectonic pile. The

| LIGURIAN NAPPE | | TECTONIC UNITS | synonyms |
|---|---|---|---|
| proposed nomenclature | | | |
| Internal Ligurian Units (IL) | | Lupicaia Creek Unit | Montecatini Sandstone Subunit [139] |
| | | | Palombini Shale Ophiolitiferous Unit [140] |
| | | Montignoso Unit | Upper Ophiolitic Unit [55; 139] |
| | | | Palombini Shale Ophiolitiferous Unit [135] |
| External Ligurian Units (EL) | | Ophiolitic tectonic slices | Vara Unit [16] |
| | | | Lower Ophiolitic Unit [55; 139] |
| | | | Monteverdi M.mo-Lanciaia Ophiolitiferous Unit [135] |
| | EL1 | Castelnuovo Val di Cecina Unit | Marly-Calcareous Flysch Unit [141]; Monteverdi M.mo-Lanciaia Ophiolitiferous Unit [135] |
| | | Castelluccio Unit | Montaione Ophiolitiferous Unit [135] |
| | EL2 | Morello Unit | Calvana Supergroup [23]; Santa Fiora Unit [142] |

**Table 1** – Proposed nomenclature for the Ligurian Units in western Tuscany.

Montecatini Sandstone, in fact, shows evidence of thrusting on top of the Montignoso Tectonic Unit, even though the exposure is so poor that the Montecatini Sandstone could also represents a tectonic wedge within the Montignoso Tectonic Unit.

The Montecatini Sandstone, Maastrichtian- late Paleocene [48, 49], is formed by frequently amalgamated strata of medium to coarse-grained arkosic sandstones, with siltstones, shales and a lower amount of marls. Clay chips and slumping events are common.

**Montignoso Tectonic Unit.** The Montignoso Tectonic Unit is formed by the Lower Cretaceous Palombini Shale. This formation is characterized by a block-in-matrix style of deformation of the original layered sequence of alternating shales and fine-grained calcareous or siliciclastic turbidites [50]. They also contain dismembered portions of the ophiolitic sequence which rarely maintains stratigraphic contacts with the Palombini Shale themselves (Fig. 3a).

### *3.2 External Ligurian Units*

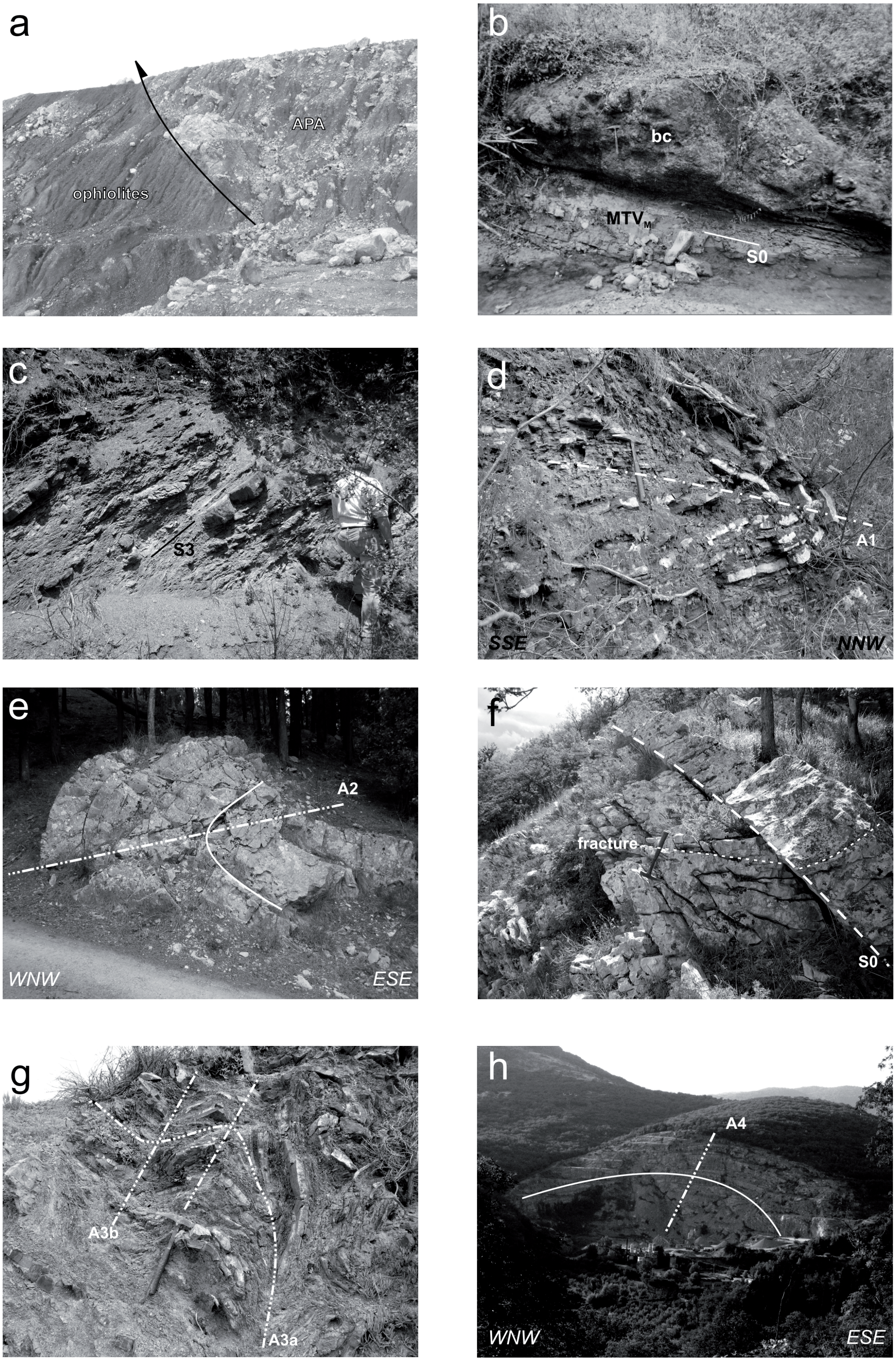
**External Ligurian 1 – EL1 – units.** They include the Castelnuovo-Val di Cecina and the Castelluccio tectonic units. These units have the same stratigraphy, but they are divided by a regional out-of-sequence-thrust into two distinct tectonic units [16]. They are both formed by an upper Santonian-lower Paleocene Helminthoid Flysch, Monteverdi M.mo Formation, that lies on a Cenomanian–lower Turonian pelitic-calcarenitic basal complex, Poggio Rocchino Formation. The age hiatus between the flysch and the basal complex (Fig. 2) could be explained as a sedimentation gap, even though the poor exposure of contacts hampers a precise age calibration [45].

The flysch succession contains cohesive and granular debris flow deposits, hyperconcentrated flows and slide blocks of Campanian age. These deposits are represented by polymictic breccias (Fig. 3b) and coarse grained arenites and rudites with clasts and matrix from the ophiolitic succession (ultramafics, basalts, cherts and limestones). The slide blocks are generally monolithologic, while the biggest bodies are commonly polylithologic.

Some of these bodies, that can have sizes of few km, can also represent blocks emplaced by tectonic mechanisms, as thrust sheets. These blocks were able to modify the depositional systems in the sedimentary basin acting as obstacles, and suggesting a hypothesis to the end of sedimentation of the Monteverdi M.mo Formation as related to the diachronous (late Maastrichtian-Paleocene) emplacement of ophiolitic tectonic slices.

**External Ligurian 2 – EL2 – units.** EL2 units are here represented only by the Morello Tectonic Unit. From the bottom to the top this is formed by: a Upper Cretaceous-lower Tertiary clay-rich, basal complex, the Sillano Formation, an Upper Cretaceous turbidites, commonly intercalated in the Sillano Formation, and called Pietraforte Formation, and an Eocene Helminthoid Flysch, the Monte Morello Formation [51, 52].

Ophiolitic debris flows deposits and slide blocks, similar to those in the EL1 units, are present also here and located between the Sillano and Monte Morello formations, implying deposition during the early Eocene. As in the EL1 units, some


a
APA
ophiolites
b
bc
MTV$_M$
S0
c
S3
d
A1
SSE
NNW
e
A2
WNW
ESE
f
fracture
S0
g
A3b
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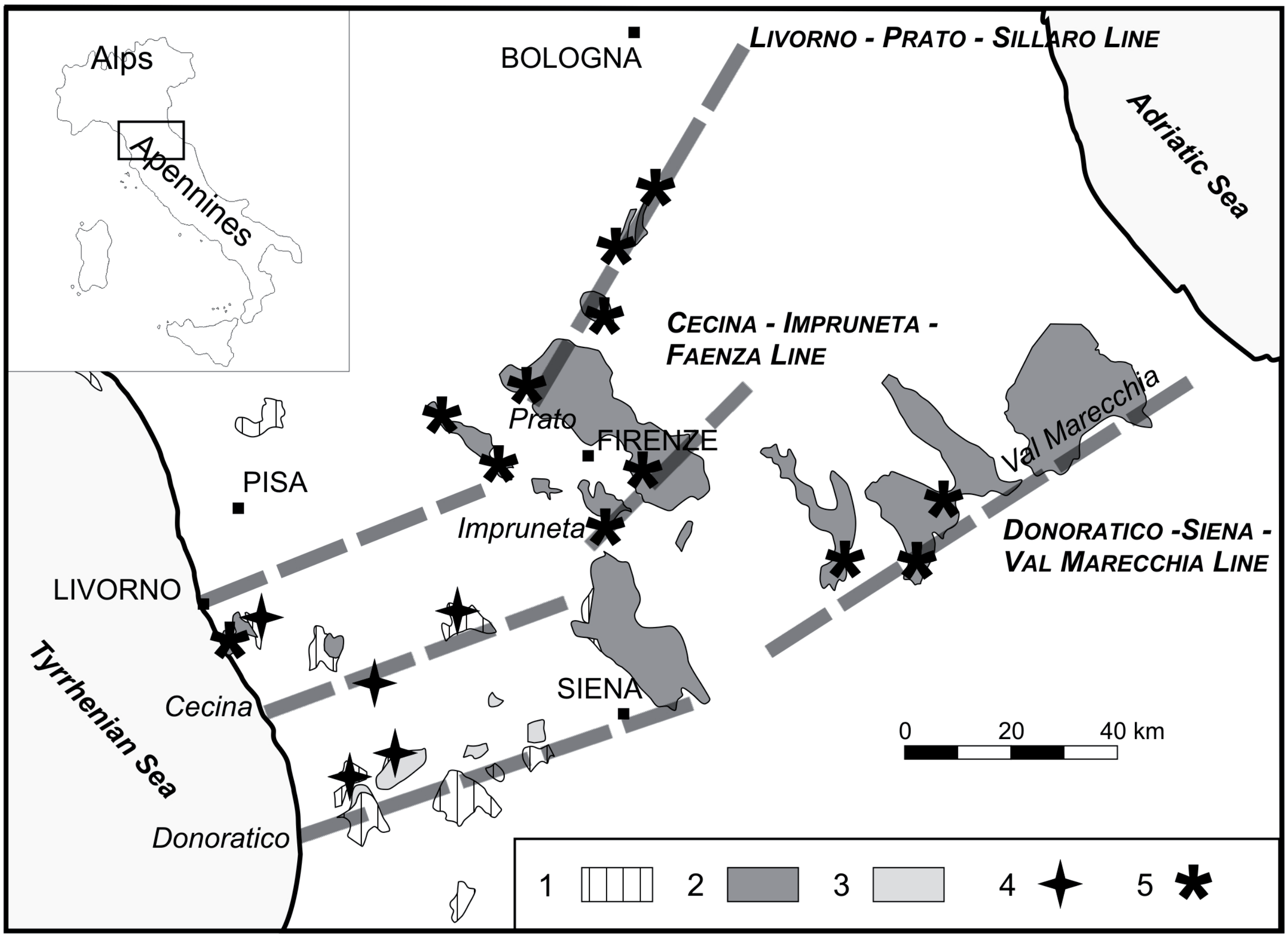
h
A4
WNW
ESE




of these slide blocks have sizes and geometrical setting that suggest tectonic mechanisms of emplacement.

**Ophiolitic tectonic slices.** The ophiolites of Western Tuscany crop as discontinuous assemblages of tectonic slices on top of EL1 units. They are represented by: serpentinized peridotites (spinel lherzolites), olivine bearing gabbros, covered by ophiolitic breccias interlayered with MOR-type basalts, and by the pelagic sedimentary succession with radiolarites, limestones and shaly-calcareous formation (Mt. Alpe Chert, Calpionella Limestone, Palombini Shale, respectively).

The attribution of the ophiolites of Western Tuscany to the EL is due to their geometrical and tectonic setting rather different from the Vara Unit, located on top of the tectonic pile as coherent remnants of the original oceanic crust.

**Fig. 4** – Distribution of the ophiolitic slide blocks and breccias in the Monte Morello Formation (EL2) and in the Monteverdi Marittimo Formation and relationships with the Livorno – Prato – Sillaro (LPS), Cecina – Impruneta – Faenza (CIF) and Donoratico – Siena – Val Marecchia (DSM) tectonic lines. **1:** Monteverdi Marittimo Formation outcrops; **2:** Morello Unit outcrops; **3:** Lanciaia Formation outcrops; **4:** ophiolitic slide blocks and breccias in the Monteverdi Marittimo Formation; **5:** ophiolitic slide blocks and breccias in the Monte Morello Formation.

**Fig. 3** – **a:** The Montignoso Tectonic Unit; tectonic contact between the Palombini Shale (APA) with block-in-matrix style of deformation and a portion of the ophiolitic sequence; **b:** ophiolitic breccia (bc) interbedded in the Monteverdi Marittimo Formation; **c:** tectonized slump in the Sillano Formation; the scaly cleavage developed during the D3 deformation phase; **d:** fold of the D1 deformation phase in the Poggio Rocchino Formation; **e:** parallel fold of the D2 deformation phase in the Monte Morello Formation; **f:** decimetric spaced fractures associated with the angular hinge of a pluri-decametric fold in the Monte Morello Formation; **g:** interference pattern between D3a and D3b folds developed during the D3 deformation phase in the Sillano Formation; h: pluri-decametric open fold developed during D4 deformation phase in the Monte Morello Formation.

On the regional scale, the outcrops of the ophiolitic debris of the EL2 units are organized on WSW-ENE trends (SW-NE in the external sector of the chain) following defined tectonic lineaments that cross cut the Northern Apennines from the Tyrrhenian to the Adriatic sea (Fig.4). These lineaments are also the northern and the southern boundaries of the Monte Morello Formation cropping area (Fig.4).

### *3.3 Epiligurian Succession*

The Epiligurian sequences were originally defined by Ricci Lucchi and Ori [46] in the Emilia-Romagna Apennines as the middle Eocene to Miocene sediments uncomforma-

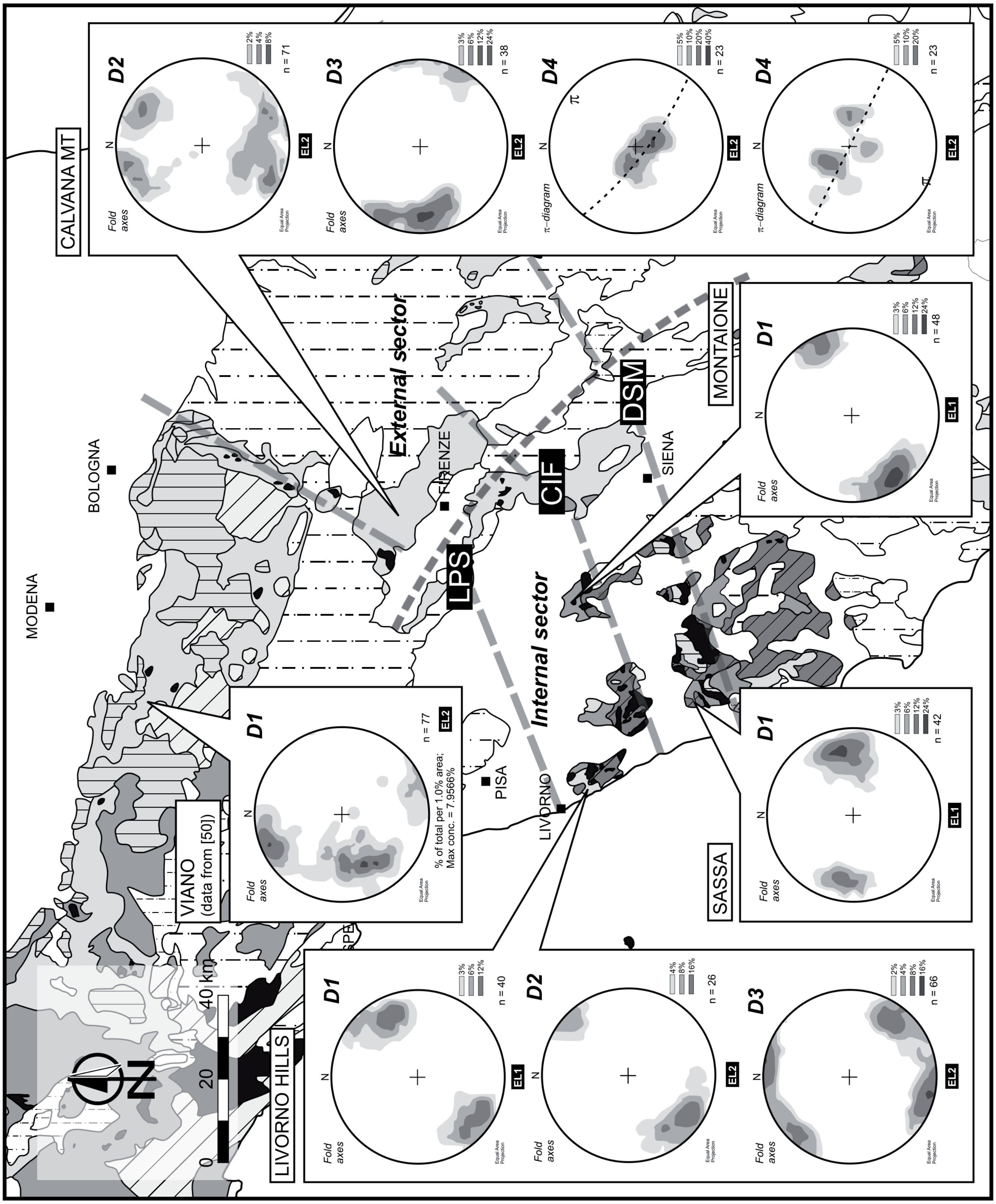



**Fig. 5** - Equal area, lower hemisphere stereographic representation of the D1, D2, D3 and D4 structural data. As a consequence of the post-Messinian differential rotation [60-63], the structures developed during the same deformation phase do not have the same orientation in the internal (western) and external (eastern) sector of the Northern Apennines (see text for details). Data from the Viano area are from [50].

bly deposited on top of the already accreted Ligurian Units during their translation toward the Adria foreland. In Western Tuscany the Lanciaia Formation (late Paleocene to middle Eocene) could be considered the first Epiligurian sequence (i. e. a "proto Epiligurian sequence"), in fact it has been interpreted as a syn-tectonic deposit [33, 53] uncomformably overlying the deformed ophiolitic tectonic complex and the EL1 [54-56]. However, it is worth to point out that the Lanciaia Formation and the more external Epiligurian sequences have been deposited in very different tectonic contexts, in fact, while the first were deposited in the ocean after the Late Cretaceous tectonic phase, the second have been deposited after the middle Eocene Ligurian tectonic phase when the ocean was already closed.

The Lanciaia Formation is formed by a basal ophiolitic sedimentary mélange with blocks, breccias and sandstones, that evolves to a thin sequence of calcarenite, marlstone, sandstone and siltstone, irregularly alternating with ophiolite bearing debris flows deposits.

## 4. Deformation history of the Ligurian Units in Western Tuscany

Detailed field mapping and structural analyses on key areas of Western Tuscany allowed the identification of four main deformation phases in the Ligurian Units. The age of the deformation is deduced from both the age of the formations involved and from the relationships with the Epiligurian succession and the late-orogenic sediments, while cross-cutting of structures suggests relative timing. The deformation events will be described in both EL1 units and EL2 units, even though D1, in particular, involved only EL1 units, producing a geometric setting not directly comparable between the two sets of units.

The oldest perturbations detected in the EL1 units are soft-sediment and they are testified by debris flows and slide blocks, diffuse slump folds and intraformational breccias implying Late Cretaceous instability of the ocean floor. This instability can be associated to the emplacement of the restricted ophiolitic tectonic slices due to the beginning of the orogenesis.

### *4.1 D1 deformation phase*

The syn-sedimentary deformation structures of the EL1 units are involved by widespread and well developed D1 folding (Fig. 5). The folds are asymmetric with inclined to recumbent setting, round hinges and class 1B, 1C and 2 geometry according to Ramsay [57] (Fig. 3d). The interlimb angle ranges from 0° to 30° and their size varies from the centimeter to the kilometer scale. Fold trend maintains a SW-NE to WSW–ENE direction and a SE to SSE vergence, even though later deformation locally scatters the D1 fold axes (Fig. 5). A well developed axial plane cleavage is present only in the siltstone – shale succession of the basal complexes. Shear zones parallel to the fold axes (A1) and with tectonic transport toward SSE are associated with this folding event.

The development of D1 is restricted to EL1 units, even though also EL2 units show some perturbation (Fig. 6). In particular D1 has an effect on the sedimentary processes active in EL2 units and produced syn-sedimentary deformation related with the emplacement of local ophiolitic tectonic slices (Fig. 3c).

The direction of the maximum shortening, deduced from the D1 structures in the EL1, is NNW-SSE oriented with tectonic transport toward SSE.

### *4.2 D2 deformation phase*

D2 is associated with the development of a new generation of folds. The D2 phase lead to a D1 coaxial refolding event in the EL1 units while the same tectonic pulse gave the most pervasive deformation identifiable on the EL2 units (Fig. 6).

In EL1 units, D2 produced close to tight [58] asymmetrical folds with generally parallel geometry, upright attitude and a SSE vergence (Fig. 3e). The superposition of this event on the D1 folds produced interference patterns referable to the type 3 of Ramsay & Huber [59].

In EL2 units, D2 is represented by asymmetrical (generally recumbent) tight to isoclinal folds with roughly parallel geometry (classes 1B, 1C and 2 of Ramsay [57]) and rounded hinges. Flexural slip folds with well developed striations on the bedding surfaces are observable in the well bedded calcareous sequences of the Helminthoid flysches. The folds are SW-NE oriented and SE vergent in the internal sector of the chain while are NNE-SSW oriented and ESE vergent in the external sector (Fig. 5). Axial plane cleavage is common in the alternating shale and siltstones of the Sillano Formation, the basal complex of EL2 units.

D2 affected also the Epiligurian Succession, developing the oldest and most widespread deformation detected in these deposits. In the Epiligurian Succession, the deformation is represented by open to close folds with upright to inclined attitude; their axes trend WSW-ENE and axial planes dip toward NNW (Fig. 6).

All the D2 kinematic indicators point to a shortening direction NNW-SSE oriented with tectonic transport toward SSE in the internal sector of the chain while a WNW-ESE shortening direction with ESE tectonic transport is recognizable in the external sector of the chain.

### *4.3 D3 deformation phase*

D3 develops structures quite different in EL1 units with respect to EL2 units. In the EL1 units the structures are represented by large scale (decametric to kilometric) gentle to open folds with vertical axial planes and rounded hinges (Fig. 6). In the EL2 units, the folds have an inclined to recumbent attitude with interlimb angles ranging from 30° to 70° in the Monte Morello Formation, and from 0° to 70° in the Sillano Formation. In the Monte Morello Formation the D3

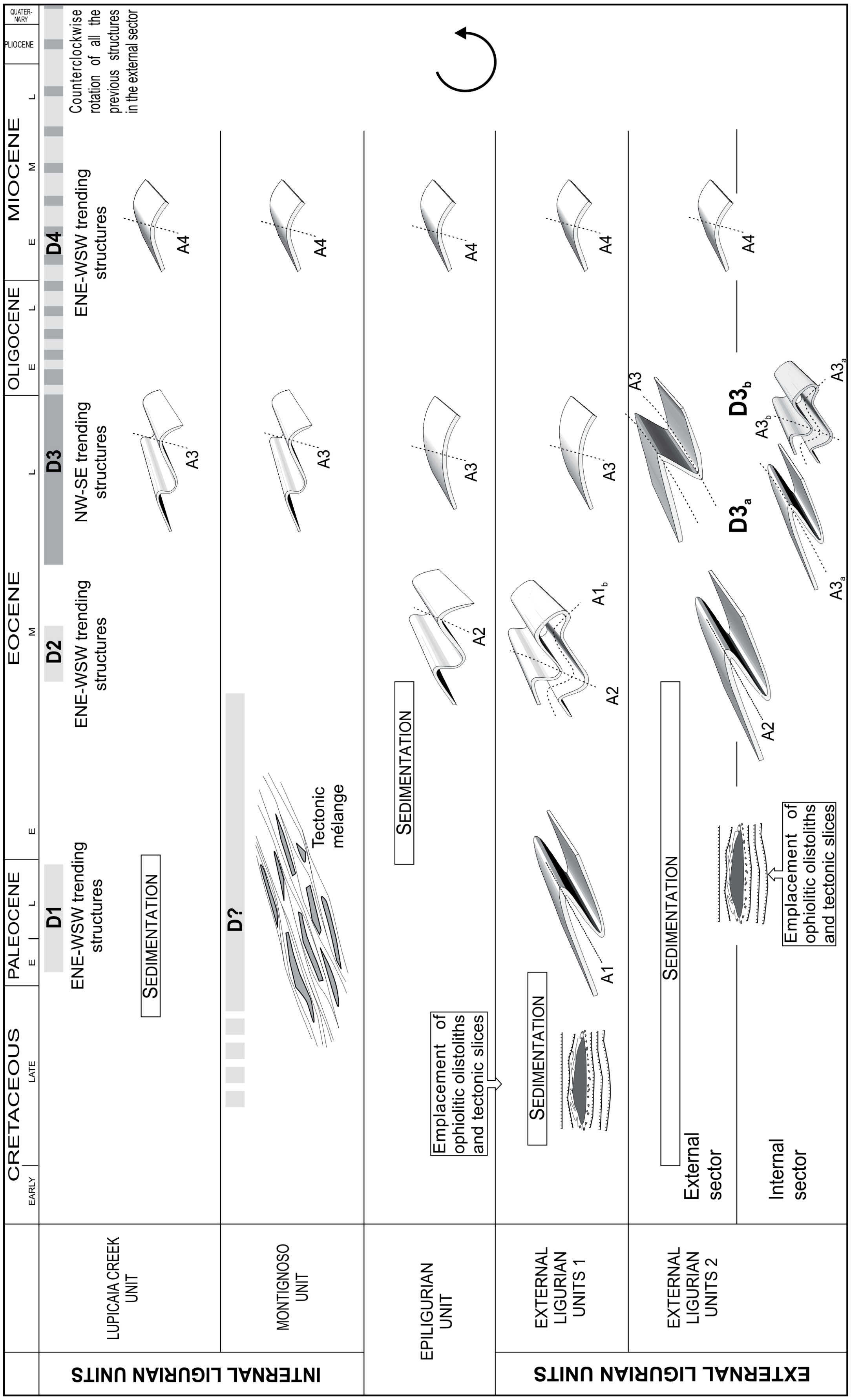



**Fig. 6** – Sketch of the structural evolution of the Ligurian Units in Western Tuscany. D?: pre-middle Eocene deformation phase in the Montignoso Unit; **A1, D1** fold axes; **A2, D2** fold axes; **A3a**, first generation of fold axes of the **D3** deformation phase in the **EL2** cropping out in internal (western) sector of the chain; **A3b**, second generation of fold axes of the **D3** deformation phase in the **EL2** cropping out in internal (western) sector of the chain; **A4, D4** fold axes. The orientations of the structures in the figure refers to the internal (western) sector of the Northern Apennines.

folds are metric to decametric and show a brittle behavior that produced centimetric to decimetric spaced fractures (Fig. 3f) associated with angular hinges. Instead the Sillano Formation D3 folds have a similar geometry and are centimetric to metric in size. D3 folds trend NNW-SSE and have an ENE vergence in the internal sector of the chain while are WNW-ESE oriented with NNE vergence in the external sector (Fig. 5). West dipping shear zones marked by S-C structures are commonly developed at the interface of horizons with strong competence contrast.

The overprinting of the D3 folds onto D2 folds produced interference patterns of the type 2 according to Ramsay & Huber [59]. Mesoscopic evidence of this interference is rare, because of the brittle disruption developed in the D3 hinges that strongly tectonized the zone of interference with the D2 hinge zones.

In the EL2 units of westernmost Tuscany, the progressive deformation lead to a coaxial refolding event within D3 characterized by close to tight asymmetrical similar-parallel folds with axial surfaces steeply dipping toward WSW (Fig. 3g).

In the present study, the Montignoso Tectonic Unit and the Lupicaia Creek Unit (IL) have been involved in the same, D3 deformation event, even though the recognition of structures related to this event is very difficult, especially in the Montignoso Tectonic Unit. In fact the chaotic assemblage of the Montignoso Tectonic Unit records a complex deformation history before the D3 deformation event and also the following D4 phase altered the geometry quite complexly. Nevertheless the NNW-SSE alignment of the ophiolitic body scattered in the chaotic pelitic matrix of the Palombini Shale, the S-C structures and the most evident scaly cleavage can be referred to the D3 event. The tectonic transport as recognizable from these structures points to movements toward ENE.

The structural data collected in the Montecatini Sandstone (Lupicaia Creek Tectonic Unit) are poor due to the scarce and badly exposed outcrops. Nevertheless this Unit shows a little deformation degree and the few fold axes have scattered orientations.

### *4.4 D4 deformation phase*

The D4 deformation phase produced large (km) and widespread gentle to open folds with upright attitude, trending WSW-ENE and with axial planes steeply dipping toward either NNW or SSE (Fig. 3h, Fig. 5). D4 also produced high angle faults, commonly oriented N30°E and cross-cutting the D3 folds. Foliated cataclasites are commonly present along the faults in the Helminthoid flysches. In the cataclastic zones, the kinematic indicators, mainly S-C structures, show an alternation of dip-slip and strike-slip movement.

Since Messinian all the orientations of the above described structures have been modified by the counterclockwise rotations occurred in the eastern/external sector of the Northern Apennines, as visible also from the clear convex shape [60, 61]. During the event that brought to the increasing curvature, all the structures jointly rotated counterclockwise. The internal sector, instead, lacks evident rotation and in fact the Cretaceus-Eocene structures [62, 63] roughly maintain their original parallelism. As a consequence of this differential rotation, the structures developed during the same deformation phase do not have the same orientation in the internal (western) and external (eastern) sector of the Northern Apennines (Fig. 5). This late rotation, furthermore, lead to the amplification and reactivation of the D4 faults, even though a precise reactivation rate is never clearly observable.

## 5. The evidence for transverse lithospheric tectonic lineaments

Thrusts, folds and the pervasive deformation recorded in the Ligurian Units show persistent WSW–ENE orientations (SW-NE in the external sector). Also the outcrops of the ophiolitic debris in the EL1 and EL2 units, and both the northern and the southern boundaries of the Monte Morello Formation cropping area are aligned on WSW-ENE directions (SW-NE in the external sector; Fig. 4). These structural and stratigraphic characters of the Ligurian Units in Western Tuscany, well match deep crust geophysical data [35, 64, 65] that allow the identification of three lineaments cutting through the study area: the Livorno-Prato-Sillaro (LPS) Line, the Cecina-Impruneta-Faenza (CIF) Line and the Donoratico-Siena-Val Marecchia (DSM) Line (Fig 1). WSW–ENE (SW-NE in the external sector) discontinuities are detected in the Northern Apennines as lithospheric tectonic lineaments that cross cut orthogonally the chain from the Tyrrhenian to the Adriatic sea. Tectonic lineaments oriented perpendicularly to the Northern Apennines chain axis have been described by many authors [66-72], which interpreted them with different kinematics (e.g. as normal faults, transfer faults, thrusts lateral ramps and strike-slip faults). As matter of facts some of these lineaments probably experienced different kinematic behaviours during the evolution of the Northern Apennines controlling strain localization in both oceanic and continental units, with evidence that they were active since the Late Cretaceous and during the oceanic and the collisional orogenic phases. The Cretaceous/Eocene deformation of the Ligurian Units has been concentrated along these oceanic transverse lineaments and, quite puzzling, the same lineaments persisted onto the Adria lithosphere, implying that they are inherited structures [73-75]. This situation will be discussed through the new stratigraphic and structural data acquired from the Ligurian Units of Western Tuscany integrated by geological and geophysical data already available for the Northern Apennines.

**Stratigraphic data**

The tectonic activity of the transverse lineaments in the oceanic lithosphere had influenced both the turbiditic sedimentation and the emplacement of ophiolitic slide blocks, since the Campanian. EL2 units, in particular, show ophiolitic olistoliths only along the LPS line, the CIF line and the DMS line (Fig. 4)

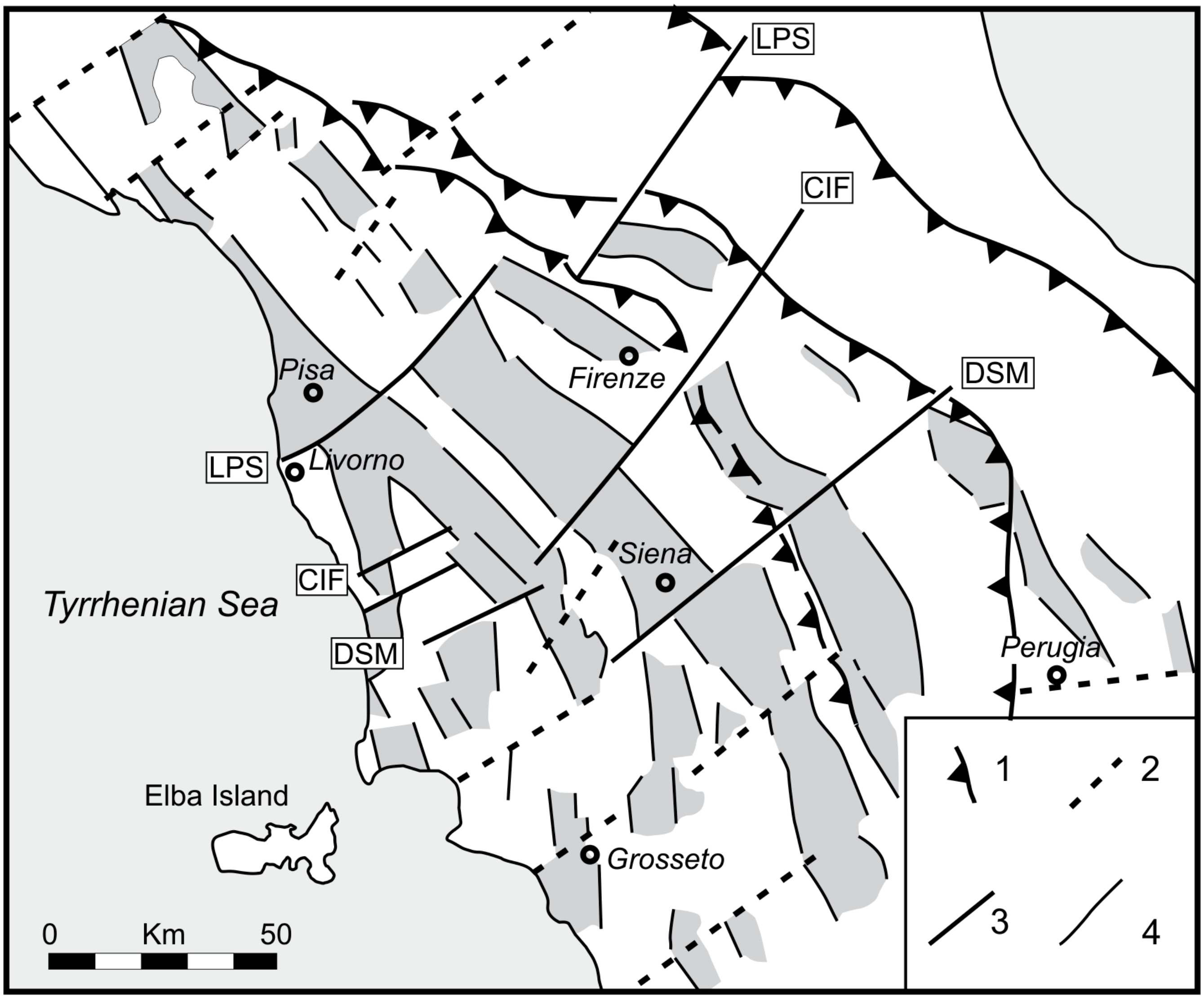



**Fig. 7** – Distribution of the Neogene and Quaternary basins (grey) in the Northern Apennines. 1: main thrust front; 2: main tectonic lines transverse to the chain axis; 3: main tectonic lines transverse to the chain axis partly matching the Livorno – Prato – Sillaro (LPS), Cecina – Impruneta – Faenza (CIF) and Donoratico – Siena – Val Marecchia (DSM) lineaments; 4: main normal faults bordering the Neogene and Quaternary basins (modified and redrawn after [83]).

testifying a tectonic control on the deposition [16]. Moreover the middle Eocene Monte Morello Formation and the Epiligurian Lanciaia Formation crop out clearly confined between the LPS and DSM lines suggesting a bounding role of these lineaments during the last stage of the oceanic closure (Fig. 4). Also, the ophiolitic debris in the Helmintoid flysch of the EL1 "Castelluccio Unit" has been found in outcrops and through drilling [16] only between the LPS and DSM lineaments [Fig. 4].

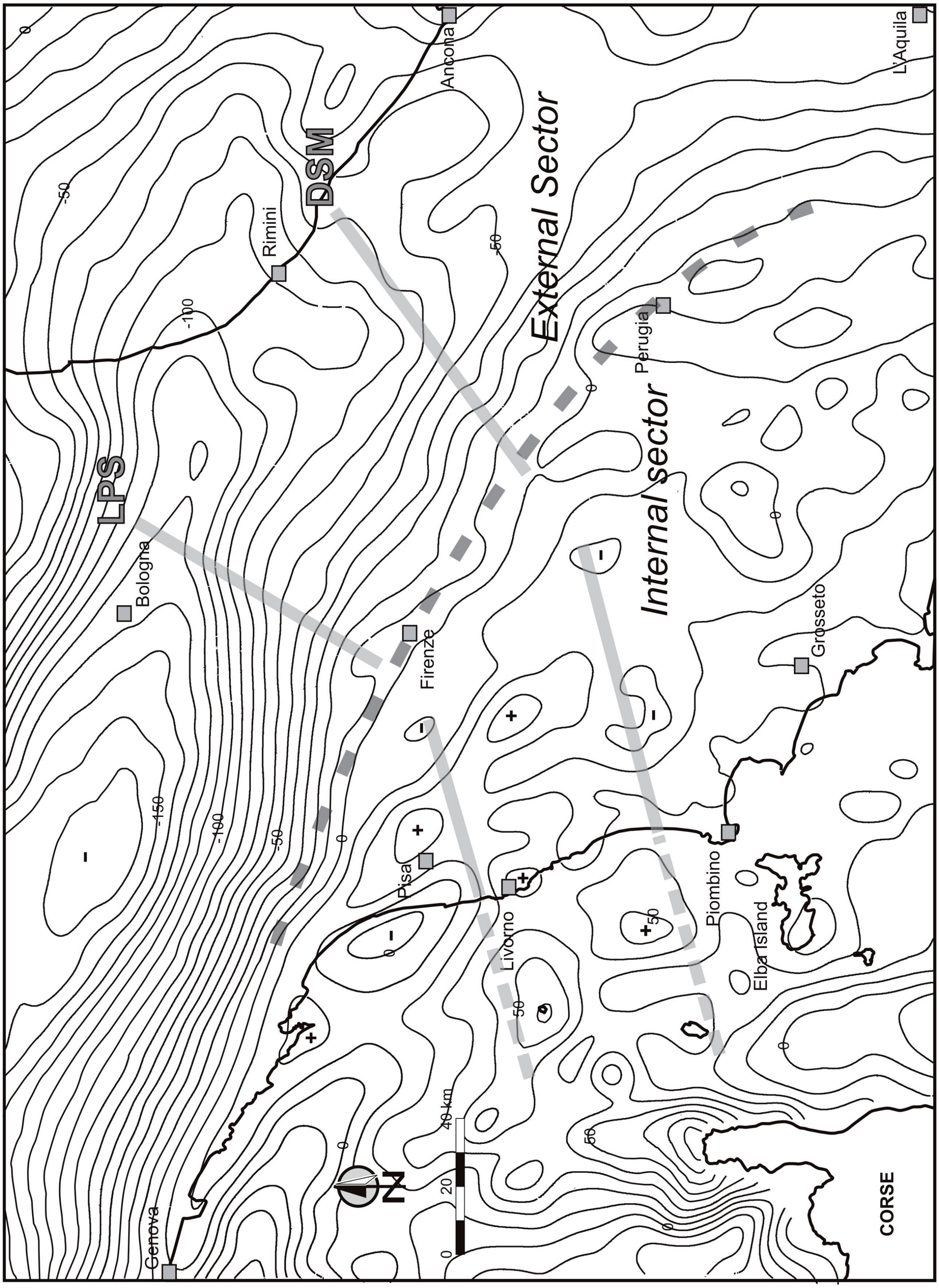
Summarizing, during the subduction phase the tectonic control of the transverse lines on the oceanic sedimentation grew gradually more important till the middle Eocene where the lineaments probably had a morphologic expression working as boundaries of the turbiditic basins.

**Structural data**

The LPS and the DSM lines ubiquitously bound to the north and to the south the area where the Ligurian Units are characterized by WSW–ENE structural trends. Outside from this area, a prominent chain-parallel, NW-SE trend is recognizable in the tectonic structures, and NE-SW trends have been interpreted as the result of strongly non-cylindrical folds [50] (Fig. 5). On the other hand, the continental units of the Adria margin have been interested only partially by the chain-transverse, WSW–ENE, structural trend and only where they are adjacent to the LPS and DSM lines. In particular the bedding of the foredeep deposits and their thrusts are generally oriented NW-SE, but close to the LPS lineament they rotate becoming NNE-SSW, i.e. parallel to the lineament itself, forming lateral ramps [70, 76].

The structural analysis carried out in the area between the LPS line to the north and the DSM line to the south is

**Fig. 8** - Gravity map of the Northern Apennines. Bouguer isoanomalies in mGal. Livorno-Prato-Sillaro (LPS) and Donoratico –Siena-Val Marecchia (DSM) tectonic lines are shown (redrawn after [35]).


Ancona
L'Aquila
DSM
Rimini
External Sector
Perugia
Internal sector
LPS
Bologna
Firenze
Grosseto
Pisa
Livorno
Piombino
Elba Island
Genova
CORSE
N
0
20
40 km
-50
-100
-150
0
50

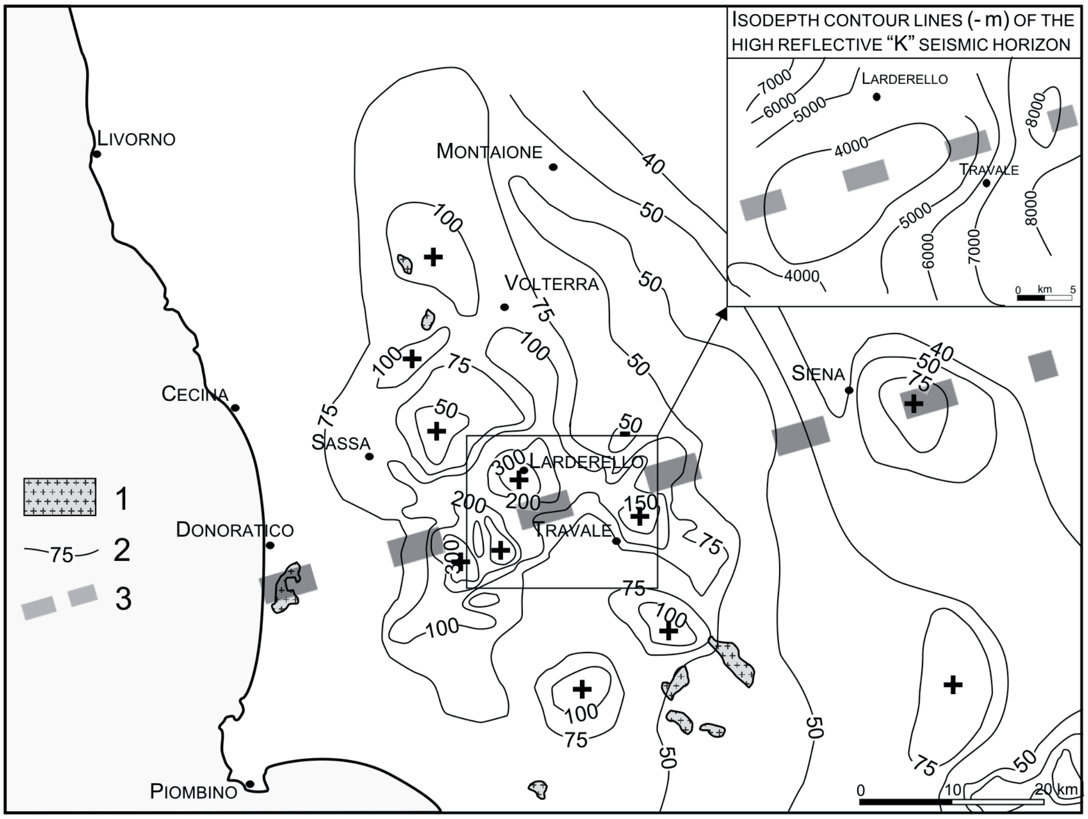



**Fig. 9** – Map of the geothermal gradient in western Tuscany (redrawn after [90]). **1:** outcropping Plio – Quaternary igneous rocks; **2:** geothermal gradient isolines (C°/km); **3:** location of the DSM lineament. The inlet shows the map of the depth of the high reflective seismic horizon (K–horizon [143]) related with the top of the Larderello Pluton (redrawn after [144]). Note how both the highs of the geothermal gradient and the shape of the K-horizon well fit the position of the DSM lineament.

particularly indicative of their interaction with the tectono-stratigraphic evolution of the Ligurian Units. From the Late Cretaceous to the Miocene, the tectonic structures developed in the Ligurian Units show shortening oriented NNW-SSE in the internal sector and WNW-ESE in the external sector.

**Data from other geological studies**

On the basis of geophysical data and regional scale map observation many authors emphasized the presence of transverse tectonic discontinuities in the Northern Apennines [68, 71, 77]. For example, on the Adriatic side of the Apennines, the LPS and the DSM lines are the most evident since they represent the boundaries between the Ligurian Units and the foredeep deposits of the Marnoso-Arenacea Formation [46, 68-70] (Fig.1). A geomorfological evidence of the LPS and DSM lineaments is their matching with first order elevation lows of the Northern Apennines water divide. Moreover, according to Salustri Galli et al. [78] this first order elevation lows of the Apennines water divide could reflect compressive transfer zones formed along inherited Mesozoic structures of the passive continental margin which have been reactivated during the subduction rollback of the Adriatic-Ionian lithosphere [77, 79-82]. On the Tyrrhenian side of the Apennines, instead, the lineaments worked as extensional transfer faults, as they partially match the borders of some Neogene basins [72, 83, 84] (Fig. 7). According to Liotta [72] and Pascucci et al. [84], in the Neogene basins separated by the lineaments the stratigraphy and the thickness of their sedimentary sequences are different. In particular the LPS lineament testifies a post-Miocene strike slip kinematic that locally deformed the sedimentary sequences of the intersected basins and shifted left-laterally the pre-Neogene substrate, (i.e. the MTR) for about 15-20 km [84] (Fig. 1).

These lines can also change their surface kinematic expression, the LPS line, for example, shows a strike slip – transpressive tectonic record of Neogene age in the Adriatic sector [69, 85, 86] and a brittle extensional deformation of early Pleistocene age in the Tyrrhenian sector [64].

**Geophysical data**

Some geophysical observations can be re-assessed in the light of the described field data supporting the lithospheric character of the lineaments. The Apennine surface geology can, in fact, trace deep seated geological processes and the configuration of the descending plate [77]. Recent studies on the Central Apennines, for example [87], describe an important segmentation of the thrust front (Tremiti Transfer zone) as a consequence of a differential slab retreat between two sectors of the subducting Adria plate. In the surface it is also possible to recognize a different tectonic evolution of areas of the upper plate correspondent to this segmentation [87].

The gravity maps of the Northern Apennines [35, 64], for example, show abrupt variations in density and trend of the Bouguer iso-anomalies near the transverse lineaments, and particularly in the vicinity of the LPS and DSM lines (Fig. 8) as already recognized by [68, 71, 88] These variations could reflect a flexure or a step in the basement [70]. However, if this hypothesis is correct, it is difficult to explain the pervasive nature of the WSW–ENE (SW-NE in the external sector) deformation structures in the Ligurian Units as the effect of a vertical offset

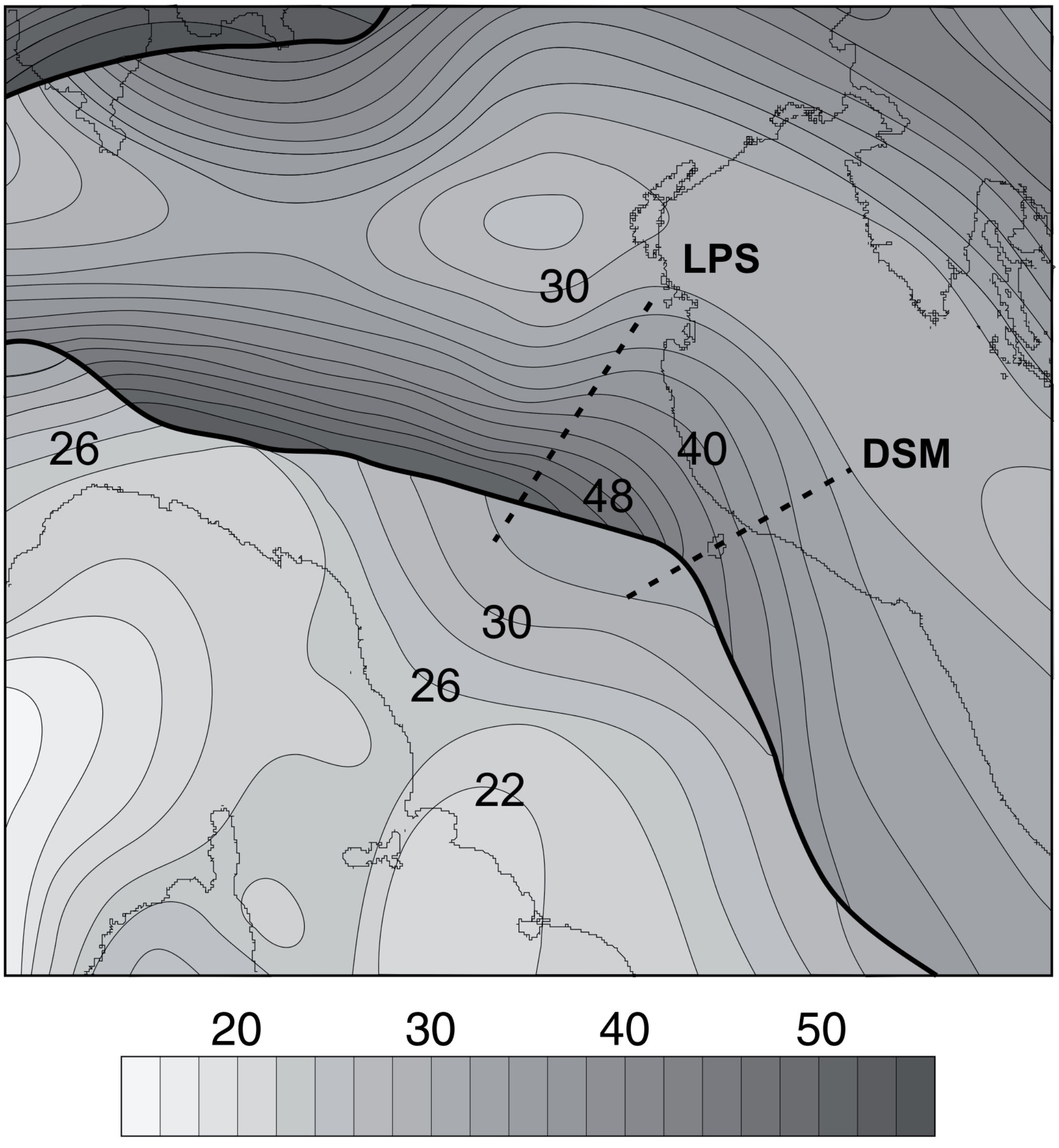



**Fig. 10** - Map of the depth of the Mohorovicic discontinuity in the Northern Apennines. From [96] and reference therein. The location of the Livorno – Prato – Sillaro (LPS) and Donoratico – Siena – Val Marecchia (DSM) lineaments is shown.

in the Adria basement. Moreover the absence of significant post-Oligocene transverse structures in the Adria margin Units (i.e. Tuscan and Umbrian-Marchean Units) near the LPS, CIF and DSM lines lead to reject the hypothesis of a recent evolution of the transverse structures cutting through the Ligurian Units.

Another piece of evidence is coming from the earthquakes distribution in the Italian peninsula [89] showing how along the Apennine axis the depth of the seismogenic layer varies not only in proximity of the southern and northern boundaries of the Northern Apennines (Ancona-Anzio Line and Sestri-Voltaggio Line respectively), but also near the LPS and DSM lines. This behavior suggests a segmentation of the Adriatic lithosphere [77] that influences the dynamic of the large scale surface deformation affecting both sides of the Apennine orogen.

In the Tyrrhenian sector, westernmost Tuscany, the location of both the CIF and the DSM lines fits the occurrence of the maximum heat flow associated to the anomalous geothermal gradient [90]. This widespread heat flow anomaly has been referred to crustal thinning related to the opening of the Tyrrhenian Sea and to the late Miocene – Quaternary magma emplacement at a shallow crustal level [91, 92]. Geological and geophysical data suggest that the emplacement of shallow-level intrusions could have been driven by the activity of strike-slip and normal faults [93]. Recent interpretation of transcrustal seismic lines of the CROP Project [94] clearly shows the Larderello Pluton intruded in the upper crust along a transurrent fault that well fit the position of the DSM lineaments. In this overview the emplacement of the Larderello and nearby intrusions could be related to the DSM line (Fig. 9). This hypothesis is also supported by the analysis of seismic fault plane solutions in the Larderello geothermal field that show the activity of a main strike slip tectonic structure oriented roughly NE-SW [95].

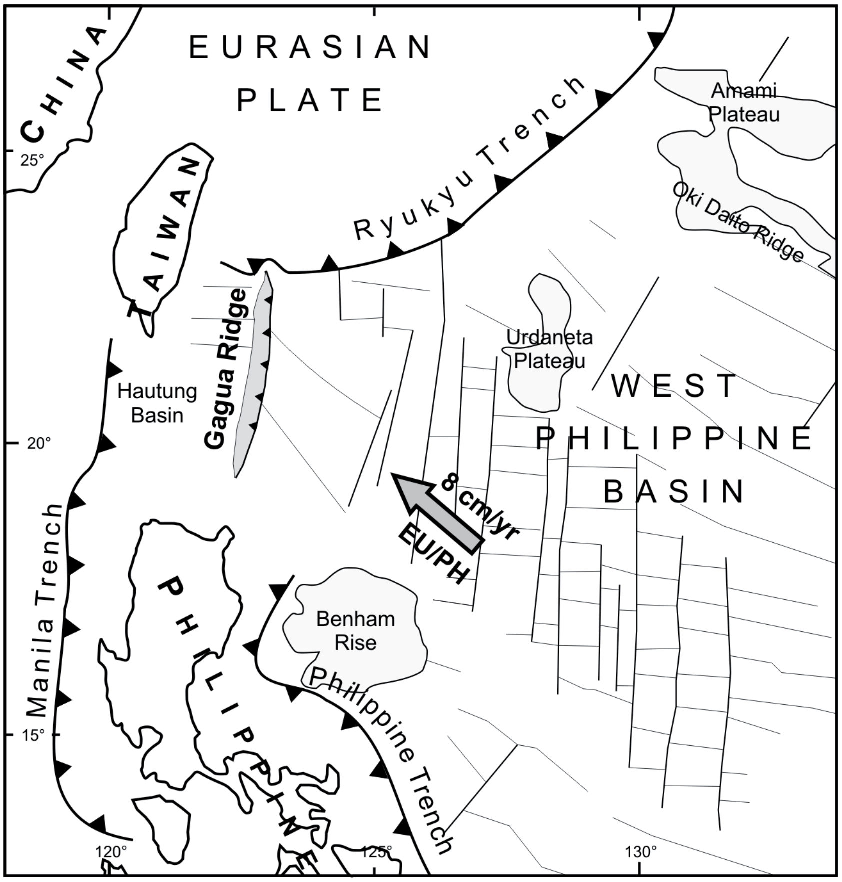



**Fig. 11** – The Gagua ridge and the geodynamic setting of the Philippine Sea Plate (modified after [136]). The thick solid arrow represents the convergence of the Philippine Sea Plate relative to the fixed Eurasia Plate after [137].

The Moho map of the Northern Apennines [96, 97] shows two important steps in correspondence of the eastern portion of the LPS and DSM lineaments (Fig. 10). These steps can also be seen both in the Bouguer anomaly map [35] and in the map of the depth of the Apennine seismogenic layer [89].

## 6. A model for the evolution of the Ligurian Units in Western Tuscany

The geological and geophysical data indicate how the DSM; CIF and LPS lines are lithospheric discontinuities active from the Cretaceous, when they were cutting the crust of the Ligurian-Piedmont Ocean, to the collisional phase of the orogenesis in the continental crust of Adria.

The evolution of the transverse lineaments and their influence on the architecture of the Ligurian subduction complex have to be explained through successive tectonic cycles and through the association of shallow crustal structures reflecting a pervasive deformation of the deeper lithosphere.

Transverse lithospheric lineaments occur in most orogens, [98-101] and some of them have acted throughout a complete Wilson cycle. In particular, Thomas [98] considered the transform faults of the eastern North America margin as long-living wide zones of pervasive deformation in the lithosphere. In this reconstruction they have been acting as weakness zones as they show repeated tectonic inheritance through successive Wilson cycles and so influencing and driving the evolution of oceanic and continental units since the Cambrian time [98].

In the same way, to explain how during subduction the DSM, CIF and LPS lineaments penetrated from the oceanic to the continental downgoing lithosphere, we have to admit their activity already in the continental crust of Adria before the Middle Jurassic opening phase of the Ligurian-Piedmont Ocean. In this scenario the tectonic activity along these lineaments started at least during the Triassic break-up stage of the Pangea both influencing the opening of the ocean and driving the nucleation of transform faults [74]. The successive convergent tectonic phase inherited and used these same transform faults – fracture zones.

Modern analogues of fracture zone reactivation in compressive tectonic regimes are the Gagua Ridge in the west part of the Philippine Sea plate [102, 103] and the Puysegur Ridge, south of New Zealand [104, 105]. Their reactivation is caused by the tectonic reorganization following the Eulerian poles migration of the Philippine sea-Eurasian plates and Pacific-Australian plates respectively [102, 104]. Particularly the Gagua Ridge is a prominent fracture zone reactivated as transpressive fault in the Eocene, that is now oriented almost orthogonally (N-S oriented) to the Ryukyu Trench (E-W oriented) where it subducts in a strongly oblique mode (Fig. 11). In this context the NW-SE motion of the Philippine Sea plate favors the development of thrusts along the Gagua Ridge [102]. The superposition of these structures and the subduction of the ridge itself has a dominant control in the accretionary wedge dynamics [103] resulting in a complex structural evolution of the sediments involved in the subduction zone [106].

The architecture of the Northern Apennines is historically interpreted as the result of "cylindrical" subduction models [33, 107-109]; however these cylindrical models leave many uncertainties as, for example, the polarity of the subduction [33, 108, 110-112]. In our view an overall oblique subduction that lead to a great amount of strain partitioning can explain the complex evolution of the oceanic closure. Many of the debated characters of the Northern Apennines, as the lack of a Late Cretaceous-Eocene volcanic arc and the long residence time of turbidites in the trench, could be associated with the oblique convergence

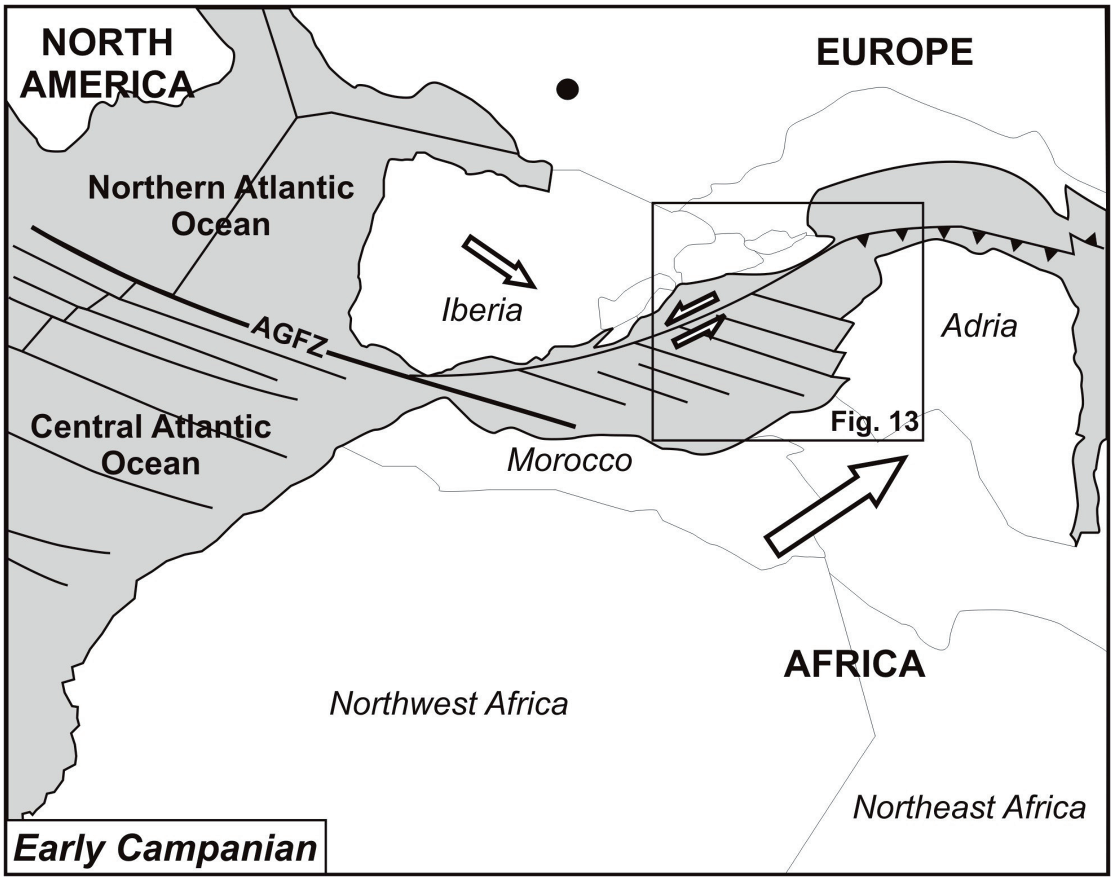



**Fig. 12** - Kinematic reconstruction of the Africa-Eurasia movements during Santonian-Campanian boundary. Arrows indicate the vectors with direction and magnitude of relative velocity between conjugate pairs of plate. Large arrow: movement of Africa relative to Eurasia; small arrow movement of Iberia relative to Africa; small circle: Euler pole of relative convergence between Africa and Eurasia. AGFZ: Azores–Gibraltar Fracture Zone (modified and redrawn after [119]).



scenario [31]. The combination of the oblique convergence and the reactivation of oceanic weakness zones may well explain the dynamic of the Apennine subduction processes. The sedimentary and structural evolution of the units involved in the oceanic closure reflects the strain partitioning connected with the oblique subduction/transverse reactivation (OS/TR).

### *6.1 Reactivation of the Lineaments in the ocean: the D1 deformation phase (Upper Cretaceous – late Paleocene)*

The Atlantic magnetic anomalies, the paleomagnetic data on Adria, Iberia, Eurasia and Africa plates and main geological constrains allowed reconstructions of the major plates movements during the early phases of the Western Tethys opening in the Middle Jurassic [5, 32, 113-116]. From these reconstructions, the continental margins of Iberia and Adria plates are commonly oriented ~NE-SW, while the transform faults were probably parallel to the principal E-W wrench structure of the Azores – Gibraltar fracture Zone (AGFZ), that connected the Western Tethys to the Central Atlantic [4, 7, 9, 74, 117-119] (Fig. 12). In this geometric frame, and starting from the Campanian, the motion of the Africa/Adria plate changed from eastward to northward, causing the beginning of the closure of the Ligurian-Piedmont Ocean [5, 115] (Fig. 12). Paleomagnetic data point to an oblique convergence between the Adria and the Iberia plates from the Late Cretaceous [31]. In this scenario the European margin was presumably activated accompanied by a large amount of strain partitioning expressed by an important role of strike-slip faulting, while, at the same time, some oceanic fracture zones were reactivated in compressive/transpressive regime (Fig. 13). During middle Campanian the tectonic activity along the inner (western) portions of the reactivated LPS, CIF and DSM fracture zones focalized tectonic pulses that produced syn-sedimentary deformations and emplacement of ophiolitic slide blocks in the adjacent flysches of the EL1 units (Fig. 6).

The second strong strain focalization along the fracture zones happened during the late Paleocene (D1) and lead to an effective reactivation of the LPS, CIF and DSM in a compressive regime. This event lead to the development of southward verging thrust sheets formed by portions of the oceanic crust with their sedimentary cover. In the internal sector of the lineaments (north-western portion of the ocean), thrusting of multiple ophiolitic tectonic slices prevent successive sedimentation in the EL1 flysch basin. In the external sector (south-eastern portion of the ocean), the reactivation of the lineaments produced only a localized emplacement of slide blocks in the sedimentary basin of the EL2 flysches.

At the same time, the active margin of the European plate was presumably affected by strain partitioning due to the high oblique plate convergence. According to Martinez et al. [120], an oblique convergence of at least 30° is required before strain

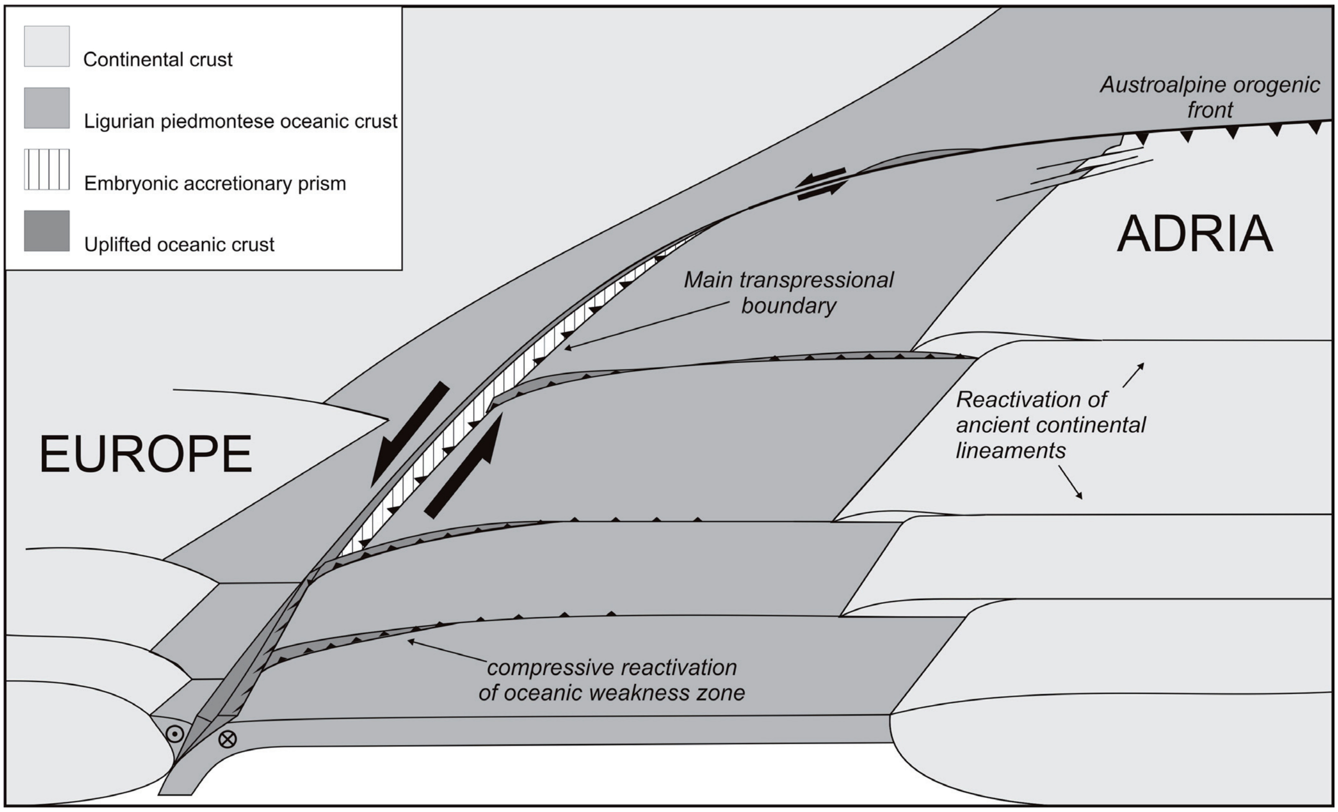



**Fig. 13** – Proposed geodynamic model for the evolution of the Ligure-Piedmontese Ocean and its continental margins during Late Cretaceous. Thick solid arrows show the transcurrent component of movement along the transpressive active margin.



partitioning occurs. With this condition the deformation in an accretionary wedge is accommodated by two main domains: an internal zone of wrenching and an external wedge of imbricate thrusts [120]. Moreover, in our model, the Ligurian accretionary wedge interferes with the reactivated fracture zones and this tectonic combination is traceable in the pre-?middle Eocene structural evolution of the Montignoso Tectonic Unit in the frontal zone of the accretionary wedge.

*6.2 Tectonic and sedimentation between the lineaments (late Paleocene – middle Eocene)*

During the late Paleocene-middle Eocene, the European plate active margin accommodated the major of the tectonic activity, while along the LPS, CIF and DSM lineaments occurred only scattered tectonic pulses.

The D1 deformation phase (Late Cretaceous-late Paleocene) brought ophiolitic tectonic slices on top of the EL1 flysch, since late Paleocene sedimentation started again with the Lanciaia Formation. The Lanciaia Formation was deposited in small piggy back basins parallel to the LPS, CIF and DSM lineaments (Fig. 14). The boundaries of the Lanciaia Formation were tectonically unstable and fed the basin with ophiolitic olistoliths and olistostromes.

From the early Eocene, the sedimentation of the EL2 units is represented by the Monte Morello Formation. Both the present distribution of the Lanciaia and the Monte Morello formations are confined between the LPS and DSM lineaments and the outcrops are elongated parallel to them (Fig. 4). The control of the lineaments on the sedimentation from the late Paleocene to the middle Eocene can be related to the damming of the turbidity currents exerted by the prominent fracture zones, testifying their morphological elevation also during this period (Fig. 14). As a modern example, this scenario resembles the relationships between tectonics and sedimentation in the West Philippine Sea, where the Gagua Ridge traps most of the sediments from the margins of eastern Taiwan [121].

*6.3 The last tectonic pulsation along the oceanic lineaments: the D2 deformation phase (middle Eocene)*

In the middle Eocene, the last oceanic stage of the orogenesis produced a tectonic pulse localized along the LPS, CIF and DSM lineaments responsible for the D2 deformation phase. This tectonic pulse caused the end of the sedimentation in both the Epiligurian Lanciaia basin and in the EL2 units flysch basins. In the internal sector of the lineaments, in fact, the Lanciaia Formation was locally overthrust by south-verging ophiolitic tectonic slices and then by a regional out-of-sequence thrust represented by EL1 terrain (Castelluccio Unit, [16]). In the external sector the compression and uplifting of the areas between the lineaments was responsible for the end of the sedimentation in the EL2 units basin.

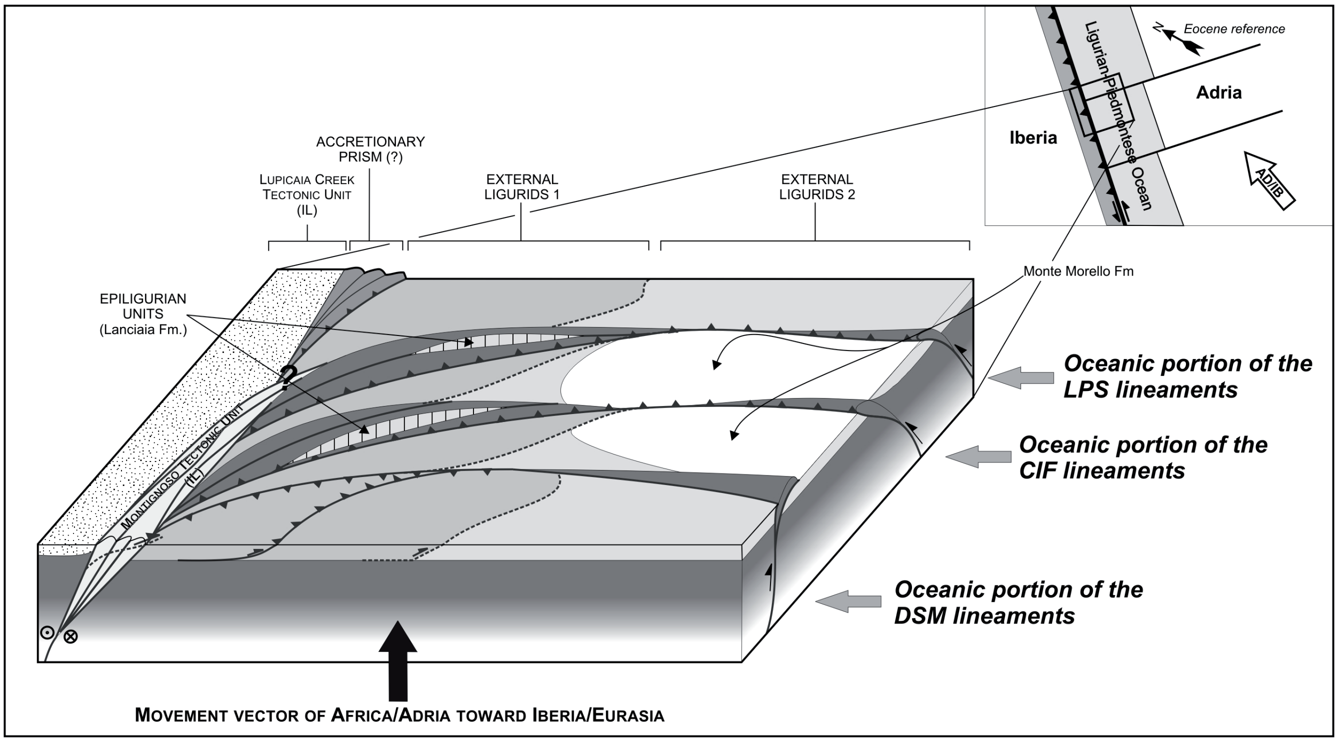



**Fig. 14** - Proposed geodynamic model for the evolution of the Ligure-Piedmontese Ocean during early Eocene. **LPS:** Livorno – Prato – Sillaro lineament; **CIF:** Cecina – Impruneta – Faenza lineament; **DSM:** Donoratico – Siena – Val Marecchia lineament.

The onland prolongation of the LPS, CIF and DSM lineaments within the Adria plate was newly activated in compressive/transpressive regime during middle Eocene, after being active in the Jurassic time [75], because of the spreading processes. This reactivation is associated with the coarse deposits recognized in the continental pelagic deposits of the Adria plate (i.e. Scisti Policromi) during middle-late Eocene [122]. These deposits are represented by coarse calcareous turbidites (i.e. Montegrossi Calcarenites and Dudda Calcarenites) and breccias containing clasts from carbonate and siliceous rocks of the Triassic to lower Tertiary Tuscan-Umbrian lithotypes and locally from low to medium grade metamorphic rocks [122-124].

### *6.4 The accretion of the External Ligurian Units: the D3 deformation phase (middle Eocene – Oligocene)*

During the final, middle-?late Eocene stage of oceanic consumption, the External Ligurian Units were detached from their ophiolitic substratum and accreted onto the European active margin (Fig. 15). The ocean-ward portion of the LPS, CIF and DSM lineaments was subducted. Throughout the stage of oceanic closure and subsequent continental collision, all the Ligurian Units were involved in the D3 deformation phase. This phase, conversely to D1 and D2, is characterized by NNW-SSE trending structures (i.e. parallel to the active margin) testifying the main role carried out by the subduction during the last stages of the Ligurian-Piedmont oceanic closure. During D3 the portion of the lineaments not yet subducted and, later, their continental prolongation acted as passive trails for the progressive westward movement of the accreted Ligurian Units onto the continental margin with the development of localized deformations.

During D3, the EL1 units, the ophiolitic tectonic slices and the overlying Epiligurian unit were emplaced on top of the EL2 flysch units through east-verging thrust. Successive to this configuration, the emplacement of both the Montignoso Tectonic Unit and the other IL onto the more external units (Fig. 14, Fig. 15) testifies the development of east-verging, out-of-sequence thrusts in the internal part of the accretionary wedge. It is noteworthy that, before the D3 phase, only the Montignoso Tectonic Unit was involved in the accretion, while the EL units were deformed only by the reactivation of the oceanic lineaments. On the other hand the D3 deformation phase is the first recorded in the other IL (i.e. Lupicaia creek Unit).

In the external part of the accretionary wedge, the more internal Ligurian Units do not overthrust the EL2 units and the D3 deformation phase developed progressive tectonic emplacement of the whole edifice onto the Adria continental margin.

### *6.5 The counter clockwise rotation of the Apennine Chain and the ensialic reactivation of the lineaments (Oligocene – Present): the D4 deformation phase*

From the late Oligocene, backarc rifting is present in the Apennine chain. According to [80-82] the backarc rifting

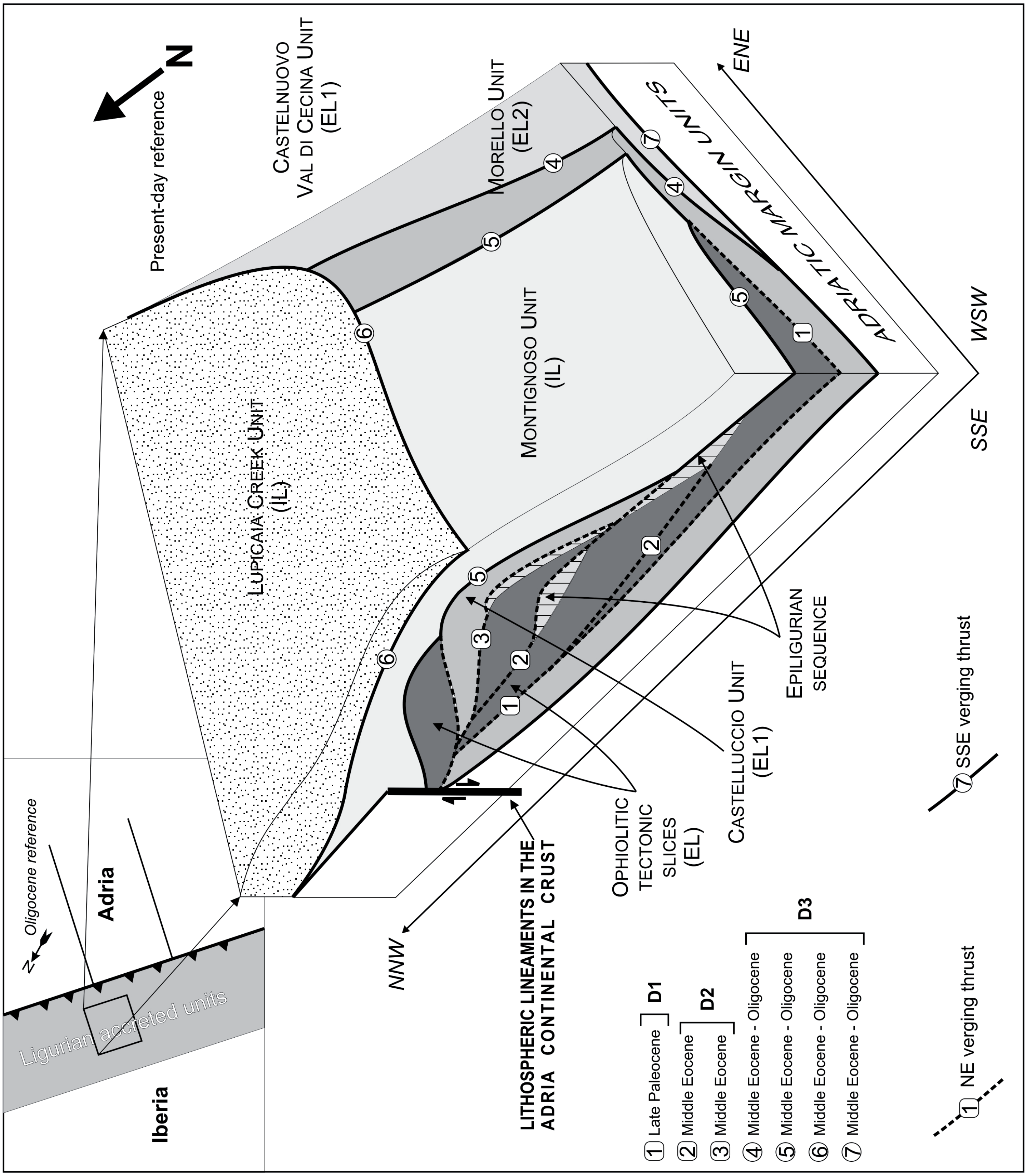



**Fig. 15** – Sketch of the tectonic stack of the Ligurian Units.

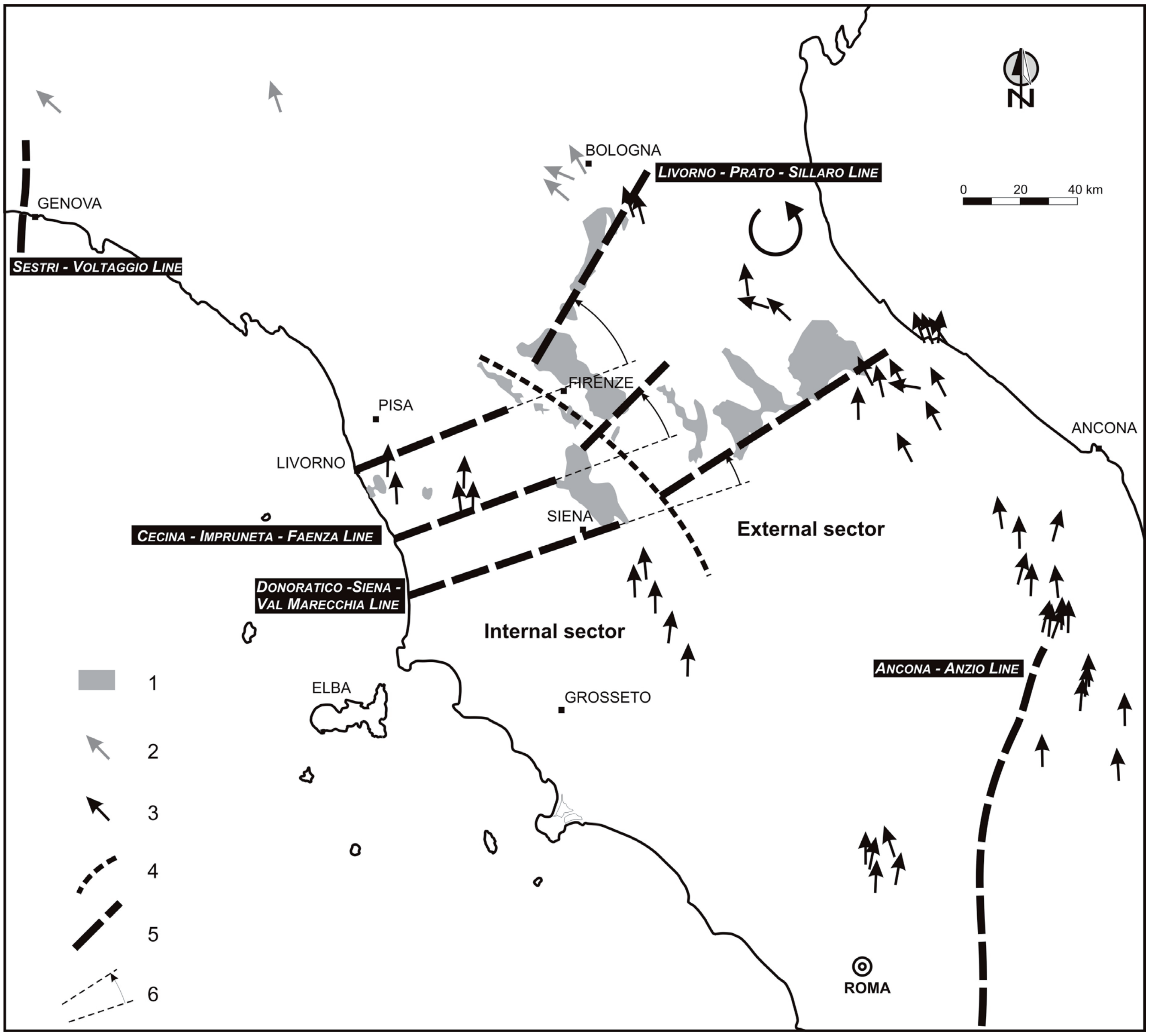



**Fig. 16** – Post-Messinian rotations in the Northern Apennines from paleomagnetic data. **1:** outcrops of the Morello Unit; **2:** tilt corrected paleomagnetic declination from Eocene to late Oligocene sediments, data from [129; 138]; **3:** tilt corrected paleomagnetic declination from late Miocene to Pleistocene sediments, data from [60; 62; 63]; **4:** boundary between the internal sector (no tectonic rotations after late Miocene) and the external sector (tectonic rotations after late Miocene); **5:** location of the Livorno – Prato – Sillaro, Cecina-Impruneta-Faenza and Donoratico–Siena–Val Marecchia tectonic lineaments; **6:** rotations inferred from geological data of the LPS, CIF and DSM lineaments in the external sector respect to the internal sector after late Miocene.

is resulting from the retreat of the NW dipping subduction plane of the Adria lithosphere. Since the early Miocene the rifting was followed by the formation of new oceanic crust [125], which caused a counter-clockwise rotation of the Sardinian-Corsican continental block and of the Apennine orogen [126].

In this context the continental portions of the LPS, CIF and DSM, that in the Triassic age controlled the opening of the Ligurian-Piedmont Ocean, represented once more weakness zones. Their new reactivation accommodated the movement between different lithospheric blocks during rotation. These lines acted as passive elements that focused the deformation producing the D4 folding associated with strike-slip and dip-slip faulting in the adjacent oceanic and continental units. The D4 structural trend is parallel to the lineaments, and D4 structures are clearly recognizable throughout the Northern Apennines from Tyrrhenian Sea to the Adriatic Sea. At the same time and far from the lineaments, the D4 structures are organized as NW-SE trending structures, parallel to D3, associated with the ongoing eastward migration of the thrust front.

From the early-middle Miocene, the progressive eastward migration of the Apennine front caused the opening of the Northern Tyrrhenian Sea. This migration is accompanied by crustal thinning of the internal sector of the Apennine chain [80]. The result is a bipartition of the Apennines in an internal extensional domain and in an external compressive domain. The lithospheric discontinuities of the Adria basement played a role also during this phase. They acted as strike-slip/transpressive faults in the compressive areas [86] and as transfer faults in the extensional areas [72, 83, 84]. The different behavior of the lineaments along strike is clearly recognizable analyzing the activity of LPS during the Neogene. This lineament, in fact, is associated with strike-slip/transpressive deformation in the compressive sector [69, 85, 86] and with brittle extensional deformation in the extensional sector [64]. Where the lineaments worked as transfer faults, they also partially separated the Neogene basins leading to an independent sedimentary evolution for each one of them (Fig. 7).

The prolongation of the lineaments from the oceanic to the continental lithosphere is the most intriguing point of the proposed model. In fact the today preservation of the oceanic deformation of the Ligurian Units along the lineaments implies a cylindrical shortening that maintains the units between the lineaments through the oceanic and continental deformation phases. Also some stratigraphic characters are preserved in a cylindrical way. The outcrops of the oceanic Eocene sediments, as the Monte Morello Formation and Lanciaia Formation, for example, are confined between the LPS and the DSM lineaments and the upper Paleocene-lower Eocene ophiolitic inputs in the Monte Morello Formation are uniquely aligned along the LPS, CIF and DSM lineaments (Fig. 4). This "conservative evolution" could be explained only admitting a strong long-living control by the LPS, CIF and DSM lineaments on the structure of the Northern Apennines, throughout the oceanic and continental tectonic phases. In the proposed model the lineaments, reactivated in the Adria lithosphere during the collision, produced significant vertical offsets that maintained the Ligurian units in a lithospheric-scale structural low during their thrusting over the continental units (Fig. 15). Later on, these large vertical offsets may have influenced also the sedimentation and structural evolution of the foredeep deposits [77]. However, this hypothesis requires further stratigraphic, structural and geophysical testing along the entire length of the lineaments.

Summarizing, the current surface appearance of the LPS, CIF and DSM lineaments on the Northern Apennines seems to reflect segmentation and tearing of the subducted oceanic slab.

A modern analogue of slab tear mechanism with strong influences on the surface geology is the subduction of the Gagua fracture zone in the Ryukyu trench [127]. Here, the subduction of lithospheric slabs with different dipping angles creates vertical offsets that drive the surface tectonic processes [128]. Similarly, in our model, it is possible to explain the persistence on the continent of the structures evolved in the oceanic domain, postulating the existence of lithospheric vertical offsets that worked as trails for the westward migration of the Ligurian Units onto the Adria continental crust.

### *6.6 The oroclinal bending of the chain (late Miocene – present)*

Since the Messinian thrusting in the compressive Apennine sector was accompanied by complex rotations, resulting in a convex thrust front toward the Adriatic foreland [60, 129] (Fig. 15). In the internal sector the extension went on without significant rotations [62, 63]. During the oroclinal bending the LPS, CIF and DSM lineaments were also involved maintaining a parallel trend in the extensional internal sector, while diverging in the external compressional sector with differential counter-clockwise rotations. Accordingly with the oroclinal bending of the chain, the counterclockwise rotation of the lineaments has angles that increase from the DSM to the LPS. In fact, the amount of rotations deduced from the paleomagnetic data shows a striking correspondence with the differential rotations between the internal and external sector of the LPS, DSM and CIF lineaments inferred from the geological data (Fig. 16).

According to Lucente & Speranza [61], the convexity of the Northern Apennines reflects the bending of the subducting Adria plate beneath the orogene [130]. In this hypothesis the lithospheric nature of the lineaments is confirmed pointing to a genetic relationship between surface deformations and deep subduction processes.

During the oroclinal bending some structures formed in the previous deformation phases could have been reactivated with strike slip/transtensive movement in order to accommodate the convex geometry of the compressive front.

## 7. Concluding remarks

Long living lithospheric lineaments can experience different kinematics through time, according to the tectonic regime in which they are involved. Lithospheric tectonic lineaments characterized the Northern Apennines area from the ? Triassic age, and some of them represents weakness zones which were reactivated during the complex history that characterized this sector of the Western Tethys. For the LPS, CIF and DSM lineaments three main activity phases can be recognized:

1. During the Jurassic opening of the Ligurian-Piedmont Ocean some lithospheric lineaments of Adria have been probably used, as nucleation points for oceanic fracture zones/transform faults. During this phase on the Adria continental margin the development of isopic zones is linked with the activity of the lineaments.

2. The evolution of the Ligurian Units during the Late Cretaceous – middle Eocene closure phase of the Ocean could be framed in a structural setting where longitudinal transpression along the European active margin is coupled with reactivation of the fracture zone of the ocean in compressive regime. The polyphasic structural evolution of the Ligurian Units evidences a large scale strain partitioning between the transpressive subduction complex and the intraoceanic reactivation of the lineaments.

3. With the beginning of the ensialic phase of the orogenesis, the LPS, CIF and DSM lineaments are likely to have been inherited as passive elements in the Adria continental crust. They accommodated the movements among the sectors of the chain with different strain rate during the Miocene rotation of the Apennines Chain and the subsequent Tyrrhenian Sea opening.

During their activity the LPS, CIF and DSM lineaments influenced the sedimentation and the structural evolution of the adjacent unit in both the Ocean and the Adria continental crust. Particularly in western Tuscany the Ligurian Units record the influence of such lineaments in the Upper Cretaceous to middle Eocene turbiditic sequence as emplacement of ophiolitic slide blocks and breccias.

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**References**

[1] Le Pichon X., Sea-Floor spreading and continental drift, J. geophys. Res. 73 (1968) 3661-3697.

[2] Pitman W. C., Talwani M., Sea-floor spreading in the North Atlantic, Geol. Soc. Amer. Bull. 83 (1972) 619-649.

[3] Abbate E., Bortolotti V., Passerini P., Sagri M., The Northern Apennines geosyncline and continental drift, in: G. Sestini (eds.), Development of the Northern Apennines Geosyncline. Sed. Geol. Special Issue 4, 3/4 1970, 251-340.

[4] Abbate E., Bortolotti V., Conti M., Marcucci M., Principi G., Passerini P., Treves B., Appennine and Alps ophiolites and evolution of the Western Tethys, Mem. Soc. Geol. It. 31 (1986) 23-44.

[5] Savostin L. A., Sibuet J. C., Zonenshain L. P., Le Pichon X., Roulet M. J., Kinematic evolution of the Tethys belt from the Atlantic Ocean to the Pamirs since the Triassic, Tectonophysics 123 (1986) 1-35.

[6] Principi G., Bortolotti V., Chiari M., Cortesogno L., Gaggero L., Marcucci M., Saccani E., Treves B., The pre-orogenic volcano-sedimentary covers of the Western Tethys oceanic basin: A review, Ofioliti 29 (2004) 177-211.

[7] Bortolotti V., Principi G., Tethyan ophiolites and Pangea break-up, The Island Arc 14 (2005) 442-470.

[8] Decandia F. A., Elter P., La zona ofiolitifera del Bracco nel settore compreso tra Levanto e Monte Zatta (Liguria Or.), Mem. Soc. Geol. It. 11 (1972) 503-530.

[9] Abbate E., Bortolotti V., Principi G., Apennine ophiolites: a peculiar oceanic crust in: G. Rocci (eds.), Tethyan ophiolites, Western Area, Ofioliti, Special Issue 1, 1980, 59-96.

[10] Barrett T. J., Spooner E. T. C., Ophiolitic breccias associated with allochtonous oceanic crustal rocks in the eastern Ligurian Apennines, Italy - a comparison with observations from rifted oceanic ridges, Earth Planetary Science Letters 35 (1977) 79-91.

[11] Lagabrielle Y., Cannat M., Alpine Jurassic ophiolites resemble the modern central Atlantic basement, Geology 18 (1990) 319-322.

[12] Lagabrielle Y., Ophiolites of the southwestern Alps and the structure of the Tethyan oceanic lithosphere, Ofioliti 19 (1994) 413-434.

[13] Lagabrielle Y., Lemoine M., Alpine, Corsican and Apennine ophiolites: the slow-spreading ridge model, C. R. Acad. Sci. Paris 325 (1997) 909-920.

[14] Mutter J., Karson J., Structural processes at slow-spreading ridges, Sciences 257 (1992) 627-634.

[15] Tucholke B. E., Lin J., A geological model for the structure of ridge segments in slow spreading ocean crust, J. Geoph. Res. 99 (1994) B6 11937-11958.

[16] Nirta G., Pandeli E., Principi G., Bertini G., Cipriani N., The Ligurian Units of Southern Tuscany, Boll. Soc. Geol. It. Special Paper 3 (2005) 29-54

[17] Marroni M., Monechi S., Perilli N., Principi G., Treves B., Late Cretaceous flysch deposits of the Northern Apennines, Italy: age of inception of orogenesis-controlled sedimentation, Cretaceous Res. 13 (1992) 487-504.

[18] Parea G. C., Contributo alla conoscenza dei Flysch a Elmintoidi dell'Appennino Settentrionale. Strutture sedimentarie, modo di deposizione e direzione di apporto, Boll. Soc. Geol. It. 80 (1961) 186–205.

[19] Parea G. C., Le provenienze nei sedimenti nel flysch campaniano – maastrichtiano dell'Appennino settentrionale tra il Passo dei Giovi e la valle del Panaro, Boll. Soc. Geol. It. 84 (1965) 217–222.

[20] Sagri M., Provenienza dei clastici nella Formazione dell'Antola, Boll. Soc. Geol. It. 88 (1969) 81-57.

[21] Sagri M., La Formazione dell'Antola nel versante tirrenico dell'Appennino Settentrionale e della Toscana a Sud dell'Arno, Mem. Soc. Geol. It. 8 (1969) 797-836.

[22] Sestini G., Flysch facies and turbidite sedimentology, in: G. Sestini (eds.), Development of the Northern Appennines Geosyncline, Sed. Geol., Special issue 4, 3/4, 1970, 559–597.

[23] Abbate E., Sagri M., The eugeosynclinal sequence, in: G. Sestini (eds.), Development of the Northern Appennines Geosyncline, Sed. Geol., Special issue 4, 3/4 1970, 251-340.

[24] Wildi W., Heavy mineral distribution and dispersal pattern in Penninic and Ligurian flysch basins (Alps, Northern Apennines), Giorn. Geol. 47 (1985) 77-99.

[25] Sestini G., Bruni P., Sagri M., The flysch basin of the Northern Appennines: a review of facies and of Cretaceous - Neogene evolution, Mem. Soc. Geol. It. 31 (1986) 87-106.

[26] Rowan M. G., The Upper Cretaceous Helminthoid Flysch of the Northern Apennines and Maritime Alps: correlation and provenance, Ofioliti 15 (1990) 305-326.

[27] Fontana D., Spadafora E., Stefani C., Stocchi S., Tateo F., Villa G., Zuffa G. G., The Upper Cretaceous Helminthoid Flysch of the Northern Apennines: provenance and sedimentation, Mem. Soc. Geol. It. 237-250 (1994).

[28] Gardin S., Marino M., Monechi S., Principi G., Biostratigraphy and sedimentology of Cretaceous Ligurid flysch: paleogeographic implications, Mem. Soc. Geol. It. 48 (1994) 219-235.

[29] Daniele G., Plesi G., The Ligurian Helminthoid flysch units of the Emilian Apennines: stratigraphic and petrographic features, paleogeographic restoration and structural evolution, Geodin. Acta 13 (2000) 313-333.

[30] Argnani A., Fontana D., Stefani C., Zuffa G. G., Late Cretaceous carbonate turbidites of the Northern Apennines: Shaking Adria at the onset of Alpine collision, J. Geology 112 (2004) 251-259.

[31] Marroni M., Treves B., Hidden terranes in the Northern Apennines, Italy: A record of Late Cretaceous Oligocene transpressional tectonics, J. Geology 106 (1998) 149-162.

[32] Le Pichon X., Bergerat F., Roulet M., Plate kinematics and tectonics leading to the Alpine belt formation - A new analysis, Geol. Soc. Am. Spec. Pap. 218 (1988) 111-132.

[33] Principi G., Treves B., Il sistema corso-appennino come prisma d'accrezione. Riflessi sul problema generale del limite Alpi-Appennino, Mem. Soc. Geol. It. 28 (1984) 549-576.

[34] Pieri M., Groppi G., Subsurface geological structure of the Po Plain, Italy, CNR, Progetto Finalizzato Geodinamica 414 (1981) 13 p.

[35] CNR, Structural Model of Italy and Gravity Map, Progetto Finalizzato Geodinamica 114 (1992).

[36] Pieri M., Three seismic profiles through the Po Plain, in: A. W. Bally (eds.), Seimic Expression of Structural Styles. A Picture and Work Atlas, AAPG Stud. Geol., 15, 1983, 25-28.

[37] Elter P., L'ensemble Ligure, Bull. Soc. Geol. Fr. 17 (1975) 984-997.

[38] Piccardo G. B., Muentener O., Zanetti A., Alpine-Apennine ophiolitic peridotites: new concepts on their composition and evolution, Ofioliti 29 (2004) 63-74.

[39] Cortesogno L., Galbiati B., Principi G., Note alla "Carta geologica delle ofioliti del Bracco" e ricostruzione della paleogeografia Giurassico-Cretacica, Ofioliti 12 (1987) 261-342.

[40] Molli G., Microstructural features of high-temperature shear zones in gabbros of the Northern Apennine ophiolites, J. Struct. Geol. 16 (1994) 1535-1541.

[41] Cortesogno L., Gianelli G., Piccardo G. B., Pre-orogenic metamorphic and tectonic evolution of the ophiolite mafic rocks (Northern Apennine and Tuscany), Boll. Soc. Geol. It. 94 (1975) 291-321.

[42] Cortesogno L., Grandjaquet C., Haccard D., Contribution a l'étude de la liaison Alpes-Apennins. Evolution tectono-métamorphique des principaux ensembles ophiolitiques de Ligurie (Apennins du nord), Ofioliti 4 (1979) 157-172.

[43] Marroni M., Meneghini F., Pandolfi L., From accretion to exhumation in a fossil accretionary wedge: a case history from Gottero, unit (Northern Apennines, Italy), Geodin. Acta 17 (2004) 41-53.

[44] Rio D., Villa G., Nannofossils dating of the Helminthoid flysch Units in the Northern Apennines, Giorn. Geol. 45 (1983) 57-86.

[45] Marino M., Monechi S., Nuovi dati sull'età di alcuni Flysch ad Helmintoidi cretacei e terziari dell'Appennino Settentrionale, Mem. Soc. Geol. It. 46 (1994) 43-77.

[46] 46. Ricci Lucchi F., Ori G. G., Field excursion D: syn-orogenic deposits of a migrating basin system in the NW Adriatic foreland, in: P. Allen (eds.), Foreland basins, Excursion Guidebook, Intern. Ass. Sedim., Fribourg, 1985, 137–176.

[47] Catanzariti R., Ottria G., Cerrina Feroni A., Tavole Stratigrafiche, in: Regione Emilia-Romagna-CNR (eds.), Carta geologico strutturale dell'Appennino emiliano-romagnolo, S.E.L.C.A., Firenze, 2002.

[48] Maccantelli M., Stratigraphy of the Ligurian formations in the Montecatini Val di Cecina area (Southern Tuscany), Mem. Soc. Geol. It. 48 (1994) 211-215.

[49] Bossio A., Cerri R., Mazzei R., Salvatorini G., Sandrelli F., Geologia dell'area Spicchiaiola-Pignano (settore orientale del Bacino di Volterra), Boll. Soc. Geol. It. 115 (1996) 393-422.

[50] Bettelli G., Vannucchi P., Structural style of the offscraped Ligurian oceanic sequences of the Northern Apennines: new hypothesis concerning the development of melange block-in-matrix fabric, J. Struct. Geol. 25 (2003) 371-388.

[51] Bortolotti V., Contributo alla conoscenza della serie Pietraforte-Alberese, Boll. Soc. Geol. It. 81 (1962) 225-304.

[52] Ponzana L., Caratteri sedimentologici e petrografici della Formazione di Monte Morello (Eocene inferiore-medio, Appennino settentrionale), Boll. Soc. Geol. It. 112 (1993) 201-218.

[53] Bortolotti V., Fazzuoli M., Pandeli E., Principi G., Babbini A., Corti S., Geology of Central and Eastern Elba Island, Italy, Ofioliti 26 (2001) 97-150.

[54] Signorini R., Centamore E., Conato V., La formazione di Lanciaia nella Val di Cecina, Boll. Serv. Geol. It. 82 (1963) 83-100.

[55] Mazzanti R., Geologia della zona di Pomarance - Larderello (provincia di Pisa), Mem. Soc. Geol. It. 5 (1966) 105-138.

[56] Cerrina Feroni A., Mazzanti R., Stratigrafia delle formazioni alloctone della Toscana Marittima. 4) La Formazione di Lanciaia nelle zone di Querceto, Micciano, Libbiano e fattoria di Monterufoli in Val di Cecina, Boll. Soc. Geol. It. 86 (1967) 673-685.

[57] Ramsay J. G., Folding and Fracturing of Rocks, McGraw-Hill New York, 1967, 568 p.

[58] Fleuty M. J., The description of folds, Proc. Geol. Assoc London 75 (1964) 461-492.

[59] Ramsay J. G., Huber M. I., The Techniques of Modern Structural Geology, Volume 2: Folds and Fractures Academic Press San Diego, 1987, 700 p.

[60] Speranza F., Sagnotti L., Mattei M., Tectonics of the Umbria-Marche-Romagna Arc (central northern Apennines, Italy): New paleomagnetic constraints, J. Geoph. Res. 102 (1997) 3153-3166.

[61] Lucente F. P., Speranza F., Belt bending driven by lateral bending of subducting lithospheric slab: geophysical evidences from the northern Apennines (Italy), Tectonophysics 337 (2001) 53-64.

[62] Sagnotti L., Mattei M., Faccenna C., Funiciello R., Paleomagnetic Evidence for No Tectonic Rotation of the Central Italy Tyrrhenian Margin since Upper Pliocene, Geophys. Res. Lett. 21 (1994) 481-484.

[63] Mattei M., Kissel C., Funiciello R., No tectonic rotation of the Tuscan Tyrrhenian margin (Italy) since late Messinian, J. Geoph. Res. 101 (1996) 2835-2845.

[64] Cantini P., Testa G., Zanchetta G., Cavallini R., The Plio–Pleistocene evolution of extensional tectonics in northern Tuscany, as constrained by new gravimetric data from the Montecarlo Basin (lower Arno Valley, Italy), Tectonophysics 330 (2001) 25-43.

[65] Wigger P. J., Die Krustenstruktur der Nordapennins und angrenzender Gebiete mit besondere Berücksichtigung der geotermischen Anomalie der Toskana, Geowiss. Abh. Reihess Heft 9 (1984) 1-87.

[66] Sacco F., Le direttrici tettoniche trasversali nell'Appennino settentrionale, Atti R. Acc. Lincei 2 (1935) 371-375.
[67] Signorini R., Linee tettoniche trasversali nell'Appennino settentrionale, Rendiconti R. Accad. Naz. Lincei 21 (1935) 42-45.
[68] Bortolotti V., La tettonica trasversale dell'Appennino. I. La linea Livorno-Sillaro, Boll. Soc. Geol. It. 85 (1966) 529-540.
[69] Groscurth J., Hemmer C., Die Prato-Sillaro Linie, eine transversee Störugszone im Nordapennin und ihre Bedeutung als synsedimentäre Faziesgrenze, N.Jb. Geol. Paläont. Abh. 144 (1973) 181-205.
[70] De Jager J., The relation between tectonics and sedimentation along the 'Sillaro line', Geol. Ultraiec. 19 (1979) 97 p.
[71] Fazzini P., Gelmini R., Tettonica trasversale nell'Appennino Settentrionale, Mem. Soc. Geol. It. 24 (1984) 299-309.
[72] Liotta D., The Arbia - Val Marecchia Line, Northern Appennines, Eclog. Geol. Helv. 84 (1991) 413-430.
[73] Abbate E., Sagri M., Le unità torbiditiche cretacee dell'Appennino Settentrionale e i margini continentali della Tetide, Mem. Soc. Geol. It. 24 (1982) 115-126.
[74] Boccaletti M., Coli M., Principi G., Sagri M., Tortorici L., Piedmont-Ligurian ocean: An example of the passive tension fissure within a megashear zone, Ofioliti 9 (1984) 353-362.
[75] Fazzuoli M., Sguazzoni G., Jurassic and Cretaceous isopic zones in the Tuscan Domain, Mem. Soc. Geol. It. 31 (1986) 59-84.
[76] Cibin U., Di Giulio A., Martelli L., Catanzariti R., Poccianti C., Rosselli S., Sani F., Factors controlling foredeep turbidite deposition: the case of Northern Apennines (Oligocene-Miocene, Italy), in: S. A. Lomas, P. Joseph (eds.), Confined turbidite systems, Geol. Soc. London Spec. Pub., 222, (2004), 115-134.
[77] Royden L., Patacca E., Scandone P., Segmentation and configuration of subducted lithosphere in Italy: an important control on thrust-belt and foredeep-basin evolution, Geology 15 (1987) 714-717.
[78] Salustri Galli C., Torrini A., Doglioni C., Scrocca D., Divide and highest mountains vs subduction in the Apennines, Studi Geologici Camerti 1 (2002) 143-153.
[79] Faccenna C., Becker T. W., Lucente F. P., Jolivet L., Rossetti F., History of subduction and back-arc extension in the Central Mediterranean, Geophys. J. Int. 145 (2001) 809-820.
[80] Malinverno A., Ryan W. B. F., Extension in the Tyrrhenian sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere, Tectonics 5 (1986) 227-245.
[81] Patacca E., Sartori R., Scandone P., Tyrrhenian basin and Apenninic arcs: kinematic relations since Late Tortonian times, Mem. Soc. Geol. It. 45 (1990) 425-451.
[82] Faccenna C., Jolivet L., Piromallo C., Morelli A., Subduction and the depth of convection in the Mediterranean mantle, J. Geophys. Res. 108 (2003) art. no.-2099.
[83] Bossio A., Cerri R., Costantini A., Gandin A., Lazzarotto A., Magi M., Mazzanti R., Mazzei R., Sagri M., Salvatorini G., I bacini distensivi neogenici e quaternari della Toscana. 1. Guide alle escursioni post-congresso, L'Appennino Settentrionale. 76. Riun. Estiva. Soc. Geol. It. Firenze (1992) 199–200.
[84] Pascucci V., Martini I. P., Sagri M., Sandrelli F., Effects of the transverse structural lineaments on Neogene - Quaternary basins of Tuscany (inner Northern Apennines, Italy), in: G. Nichols, C. Paola, E. A. Williams (eds.), Sedimentary processes enviroments and basins - A tribute to Peter Friend, I.A.S. Special Pubblication, 2007, In press.
[85] Marabini S., Vai G. B., Analisi di facies e macrotettonica della Vena del Gesso in Romagna, Boll. Soc. Geol. It. 104 (1985) 21-42.
[86] Castellarin A., Pini G. A., L'arco del Sillaro: la messa in posto delle Argille Scagliose al margine appenninico padano (Appennino bolognese), Mem. Soc. Geol. It. 39 (1989) 127-141.
[87] Scrocca D., Thrust front segmentation induced by differential slab retreat in the Apennines (Italy), Terra Nova 18 (2006) 154-161.
[88] Gelmini R., Ipotesi sul ruolo della linea tettonica trasversale Follonica-Val Marecchia sull'assetto strutturale della Toscana Marittima, Studi Geologici Camerti Volume speciale 1994/1 (1994) 201-209.
[89] Chiarabba C., Jovane L., DiStefano R., A new view of Italian seismicity using 20 years of instrumental recordings, Tectonophysics 395 (2005) 251-268.
[90] Baldi P., Bellani S., Ceccarelli A., Fiordelisi A., Squarci P., Taffi L., Correlazioni tra le anomalie termiche ed altri elementi geofisici e strutturali della Toscana meridionale, Studi Geologici Camerti Volume speciale 1994/1 (1994) 139–149.
[91] Keller J. V. A., Minelli G., Pialli G., Anatomy of late orogenic extension: the Northern Apennines case, Tectonophysics 238 (1994) 275-294.
[92] Jolivet L., Faccenna C., Goffé B., Mattei M., Rossetti F., Brunet C., Storti F., Funiciello R., Cadet J. P., D'Agostino N., Parra T., Midcrustal shear zones in postorogenic extension: example from the northern Tyrrhenian Sea, J. Geoph. Res. 103 (1998) 123-160.
[93] Acocella V., Rossetti F., The role of extensional tectonics at different crustal levels on granite ascent and emplacement: an example from Tuscany (Italy), Tectonophysics 354 (2002) 71-83.
[94] Finetti I. R., Basic regional crustal setting and superimposed local pluton - intrusion - related tectonics in the Larderello - M. Amiata geothermal province, from integrated CROP seismic data, Boll. Soc. Geol. It. 125 (2006) 117-146.
[95] Albarello D., Batini F., Bianciardi P., Ciulli B., Spinelli E., M. V., Stress field assessment from ill-defined fault plane solutions : an example from Larderello geothernmal field (western Tuscany, Italy), Boll. Soc. Geol. It. Special Issue 3 (2005) 187-193.
[96] Dèzes P., Ziegler P. A., Moho depth map of Western and Central Europe, World Wide Web Address: http://www.unibas.ch/eucor-urgent. (2002).
[97] Panza G., Pontevivo A., Chimera G., Raykova R., Aoudia A., The lithosphere-astenosphere: Italy and surroundings, Episodes 26 (2003) 169-174.
[98] Thomas W. A., Tectonic inheritance at a continental margin, GSA Today 16 (2006) 4-11.
[99] Royden L. H., Late Cenozoic tectonics of the Pannonian Basin system, in: L.H. Royden, F. Horwáth (eds.), The Pannonnian Basin - a study in basin evolution, Am. Ass. Petroleum Geol. Memoir, 45, 1988, 27-48.
[100] Watson J., The ending of the Caledonian orogeny in Scotland, J. Geol. Soc. London 141 (1984) 193-214.
[101] Jacques J. M., A tectnostrastigraphic synthesis of the Sub-Andean basins: inferences on the position of South American intraplate accommodation zones and their control on South Atlantic opening, J. Geol. Soc. London 160 (2003) 687-701.
[102] Deschamps A. E., Lallemand S. E., Collot J. Y., A detailed study of the Gagua Ridge: A fracture zone uplifted during a plate reorganisation in the Mid-Eocene, Mar. Geophys. Res. 20 (1998) 403-423.

[103]Schnurle P., Liu C. S., Lallemand S. E., Reed D. L., Structural insight into the south Ryukyu margin: effects of the subducting Gagua Ridge, Tectonophysics 288 (1998) 237-250.

[104]Collot J.-Y., Lamerche G., Wood R. A., Delteil J., Sosson M., Lebrun J.-F., Coffin M. F., Morphostructure of an incipient subduction zone along a transform plate boundary: Puysegur ridge and trench, Geology 23 (1995) 519-522.

[105]Lamarche G., Lebrun J. F., Transition from strike-slip faulting to oblique subduction: active tectonics at the Puysegur Margin, South New Zealand, Tectonophysics 316 (2000) 67-89.

[106]Dominguez S., Lallemand S. E., Malavieille J., von Huene R., Upper plate deformation associated with seamount subduction, Tectonophysics 293 (1998) 207-224.

[107]Boccaletti M., Elter P., Guazzone G., Plate tectonics model for the development of the Western Alps and Northern Apennines, Nature 234 (1971) 108–111.

[108]Elter P., Pertusati P., Considerazioni sul limite Alpi-Appennino e sulle sue relazioni con l'arco delle Alpi Occidentali, Mem. Soc. Geol. It. 12 (1973) 359 – 375.

[109]Bortolotti V., Principi G., Treves B., Ophiolites, Ligurides and tectonic evolution from spreading to convergence of a mesozoic Western Tethys segment, in: G.B. Vai, I. P. Martini (eds.), Anatomy of an Orogen, the Apennines and the adjacent Mediterranean basins, Kluwer Academic Publisher, Amsterdam, 2001, 151-164.

[110]Boccaletti M., Coli M., Decandia F., Giannini E., Lazzarotto A., Evoluzione dell'Appennino Settentrionale secondo un nuovo modello strutturale, Mem. Soc. Geol. It. 21 (1980) 359–374.

[111]Hoogerduijn Strating E. H., van Wamel W. A., The structure of the Bracco ophiolite complex (Ligurian Apennines) – A change from 'Alpine' to 'Apennine'polarity, J. Geol. Soc. London 146 (1989) 933–944.

[112]Marroni M., Pandolfi L., The deformation history of an accreted ophiolite sequence: The internal Liguride units (northern Apennines, Italy), Geodinamica Acta 9 (1996) 13-29.

[113]Srivastava S. P., Roest W. R., Kovacs L. C., Oakey G., Lévesque S., J. V., Macnab R., Motion of Iberia since the Late Jurassic: Results from detailed aeromagnetic measurements in the Newfoundland Basin, Tectonophysics 184 (1990) 229-260.

[114]Dewey J. F., Helman M. L., Turco E., Hutton D. H. W., Knott S. D., Kinematics of the Western Mediterranean, in: M. P. Coward, D. Dietrich, R. G. Park (eds.), Alpine Tectonics, Geol. Soc. Spec. Publ., 45, London, 1989, 265-283.

[115]Olivet J. L., Bonnin J., Beuzart P., Auzende J. M., Kinematique de l'Atlantic nord et central, Centre national pour l'exploration des oceans (CNEXO), Rapports scientifiques et techniques 54 (1984) 1-108.

[116]Muttoni G., Kent D. V., Channell J. E. T., Evolution of Pangea: paleomagnetic constraints from the Southern Alps, Italy, Earth Planet. Sci. Lett. 140 (1996) 97-112.

[117]Stampfli G. M., Borel G. D., Marchant R., Mosar J., Western Alps geological constraints on western Tethyan reconstructions, in: G. Rosenbaum, G. S. Lister (eds.), Reconstruction of the evolution of the Alpine-Himalayan orogeny, Journal of the Virtual Explorer., 8, 2002, 77-106.

[118]Dercourt J., Zonenshain L. P., Ricou L. E., Kazmin V. G., Le Pichon X., Knipper A. L., Grandjacquet C., Sbortshikov I. M., Geyssant J., Lepvrirer C., Pechersky D. H., Boulin J., Sibuet J. C., Savostin L. A., Sorokhtin O., Westphal M., Bazchenov M. L., Lauer J. P., Biju Duval B., Geological evolution of the Tethys belt from Atlantic to the Pamirs since the Lias, in: J. Auboin, X. Le Pichon, A. S. Monin (eds.), Evolution of the Tethys, Tectonophysics, 123, 1986, 241-315.

[119]Schettino A., Scotese C., Global kinematic constraints to the tectonic history of the Mediterranean region and surrounding areas during the Jurassic and Cretaceous, in: G. Rosenbaum, G. S. Lister (eds.), Reconstruction of the evolution of the Alpine-Himalayan orogeny, Journal of the Virtual Explorer, 8, 2002, 149-168.

[120]Martinez A., Malavieille J., Lallemand S., Collot J.-Y., Partition de la déformation dans un prisme d'accrétion sédimentaire en convergence oblique: approche expérimentale, Bull. Soc. Géol. Fr. 173 (2002) 17-24.

[121]Liu C. S., Deposition of the Orogenic Sediment in the Huatung Basin East of Taiwan, Chapman Conference on Continent-Ocean Interactions within the East Asian Marginal Seas, abstract, San Diego, 11-14 November 2002.

[122]Patacca E., Microfacies dei conglomerati della "Scaglia" e del "Macigno" di alcune serie toscane, Mem. Soc. Geol. It. 12 (1973) 187-225.

[123]Fazzuoli M., Ferrini G., Pandeli E., Sguazzoni G., Le formazioni giurassico-mioceniche della Falda Toscana a nord dell'Arno: considerazioni sull'evoluzione sedimentaria, Mem. Soc. Geol. It. 30 (1985) 159-201.

[124]Fazzuoli M., Pandeli E., Sani F., Considerations on the sedimentary and structural evolution of the Tuscan Domain since early Liassic to Tortonian, Mem. Soc. Geol. It. 48 (1994) 31-50.

[125]Burrus J., Contribution to a geodynamic synthesis of the Provençal basin (north-western Mediterranean), Mar. Geol. 55 (1984) 247–269.

[126]Montigny R., Edel J. B., Thuizat R., Oligo–Miocene rotation of Sardinia: K–Ar ages and paleomagnetic data of Tertiary volcanoes, Earth Planet. Sci. Lett. 54 (1981) 261–271.

[127]Deschamps A., Monie P., Lallemand S., Hsu S. K., Yeh J., Evidence for Early Cretaceous oceanic crust trapped in the Philippine Sea Plate, Earth Planet. Sci. Lett. 179 (2000) 503-516.

[128]Lin J. Y., Hsu S. K., C. S. J., Melting features along the Ryukyu Slab tear, beneath the southwestern Okinawa Trough, Geophys. Research Lett. 31 (2004) L19607.

[129]Muttoni G., Argnani A., Kent D. V., Abrahamsen N., Cibin U., Paleomagnetic Evidence for Neogene Tectonic Rotations in the Northern Apennines, Earth Planet. Sci. Lett. 154 (1998) 25-40.

[130]Lucente F. P., Chiarabba C., Cimini G. B., Giardini D., Tomographic constraints on the geodynamic evolution of the Italian region, J. Geoph. Res. 104 (1999) 20307-20327.

[131]Costantini A., Lazzarotto A., Mazzanti R., Mazzei R., Salvatorini G., Sandrelli F., Note Illustrative della Carta Geologica d'Italia alla scala 1:50.000, Foglio 285-Volterra, L.A.C. Firenze, 2002, 149 p.

[132]Maccantelli M., Mazzei R., Inquadramento bio-cronostratigrafico di due unità riferibili alla Formazione di Lanciaia affioranti nell'area a nord della bassa val di Cecina (Toscana Occidentale), Atti Soc. Tosc. Sci. Nat. Mem. Serie A 100 (1993) 29-43.

[133]Lazzarotto A., Martellini F., Mazzanti R., Mazzei R., Sandrelli F., La Formazione di Lanciaia nelle aree a sud di Micciano e a nord di Montecerboli (Provincia di Pisa), Atti Soc. Tosc. Sci. Nat. Mem. Serie A 102 (1995) 159 -169.

[134]Costantini A., Lazzarotto A., Maccantelli M., Mazzanti R., Sandrelli F., Tavarnelli E., Elter F. M., Geologia della Provincia di Livorno a Sud del Fiume Cecina, Quad. Mus. St. Nat. Livorno 13 (1993) 1-164.

[135]Lazzarotto A., Sandrelli F., Foresi L. M., Mazzei R., Salvatorini G., Cornamusini G., Pascucci V., Note Illustrative della Carta Geologica d'Italia alla scala 1:50.000, Foglio 295-Pomarance, L.A.C. Firenze, 2002, 140 p.

[136]Plesi G., Galli M., Daniele G., The Monti Rognosi ophiolitic unit (cfr. Calvana unit Auct.) paleogeography position in the external Ligurian domain, relationships with the tectonic units derived from the Adriatic margin, Boll. Soc. Geol. It. Special Issue 1 (2002) 273-284.

[137]Hall R., Ali J. R., Anderson C. D., Baker S. J., Origin and motion history of the Philippine Sea Plate, Tectonophysics 251 (1995) 229–250.

[138]Seno T., Stein S., Gripp A. E., A model for the motion of the Philippine Sea Plate consistent with NUVEL-1 and geological data, J. Geoph. Res. 98 (1993) 17941–17948.

[139]Muttoni G., Lanci L., Argnani A., Hirt A. M., Cibin U., Paleomagnetic evidence for a Neogene two-phase counterclockwise tectonic rotation in the Northern Apennines (Italy), Tectonophysics 326 (2000) 241-253.

[140]Bertini G., Cornamusini G., Lazzarotto A., Maccantelli M., Stratigraphic and tectonic framework of the Ligurian Units in the Castellina M.ma Hills (Southern Tuscany, Italy), 119 (2000) 687-701.

[141]Costantini A., Lazzarotto A., Maccantelli M., Sandrelli F., Ligurian units in the Monti della Gherardesca area (Southern Tuscany), Boll. Soc. Geol. It. 110 (1991) 849-855.

[142]Costantini A., Lazzarotto A., Liotta D., Mazzanti R., Mazzei R., Salvatorini G., Note Illustrative della Carta Geologica d'Italia alla scala 1:50.000, Foglio 306-Massa Marittima, L.A.C. Firenze, 2002, 172 p.

[143]Batini F., Bertini G., Gianelli G., Pandeli E., Puxeddu M., Deep structure of the Larderello field: contribution from recent geophysical and geological data, Mem. Soc. Geol. It. 25 (1983) 219–235.

[144]Gianelli G., Manzella A., Puxeddu M., Crustal models of the geothermal areas of southern Tuscany (Italy), Tectonophysics 281 (1997) 221-239.