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# Relationships between Lithospheric Flexure, Thrust Tectonics and Stratigraphic Sequences in Foreland Setting: the Southern Apennines Foreland Basin System, Italy

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## 1. Introduction

We discuss here tectonics and sedimentation processes occurring during continent-continent collision and relationships between accretionary processes on overplate, flexural lithosphere on underplate and related controls on clastic sedimentation in developing foreland basin systems. This paper focuses on and clastic sedimentation developed during the sequential history of an orogenic system, in the Mediterranean Region. These clastic trends, covering a large time span from Early Mesozoic to the present, may contribute: (1) to the paleogeographic and paleotectonic reconstructions of the southern Italy portions of the western Mediterranean orogen, and (2) to the general models of complex relationships between clastic sedimentation and paleotectonic history of other major orogens.

The evolutionary record of Earth's processes preserved in the form of sedimentary rocks has been pivotal in paleogeographical and paleotectonic reconstructions of source/ basin systems. Compositional trends of clastic strata through space and time are used to infer the structural history of adjacent mountain belts and to monitor the key geodynamic changes during orogenic processes (*e.g.* Dickinson, 1985, 1988; Critelli & Ingersoll, 1994; Critelli, 1999).

The controls on the composition and dispersal pathways of clastic strata along the convergent plate margins have long been debated (*e.g.* Dickinson, 1988; Ingersoll *et al.*, 1995). Clastic infilling of sedimentary basins in orogenic systems have been used as important indicators of tectonic activity and climatic changes. In the orogenic systems, clastic sedimentation may record the accretionary processes, the accommodation of the thrust units, and the flexural features of the foreland plate.

The development of an orogenic wedge during continental collision results in thickening of the crust. The excess mass of this thickened crust acts as a load on the underthrust plate, causing it to be flexed downwards close to the load, so developing a foreland basin (*e.g.* Beaumont, 1981; Sinclair and Allen, 1992). During plate convergence, the vertically acting load of the mountain belt migrates over the foreland plate, thus resulting in the migration of the associated foreland basin.

The foreland is the region between the front of a thrust belt and the adjacent craton (*e.g.* Dickinson, 1974; Bally and Snelson, 1980; Allen *et al.*, 1986; Miall, 1995). Large volumes of clastic sediment are derived from erosion of the thrust belt and deposited in the foreland

basin. The foreland basin generally is defined as an elongate trough that forms between a linear contractional orogenic belt and the stable craton, mainly in response to flexural subsidence caused by thrust-sheet loading in the orogen (fig. 1).

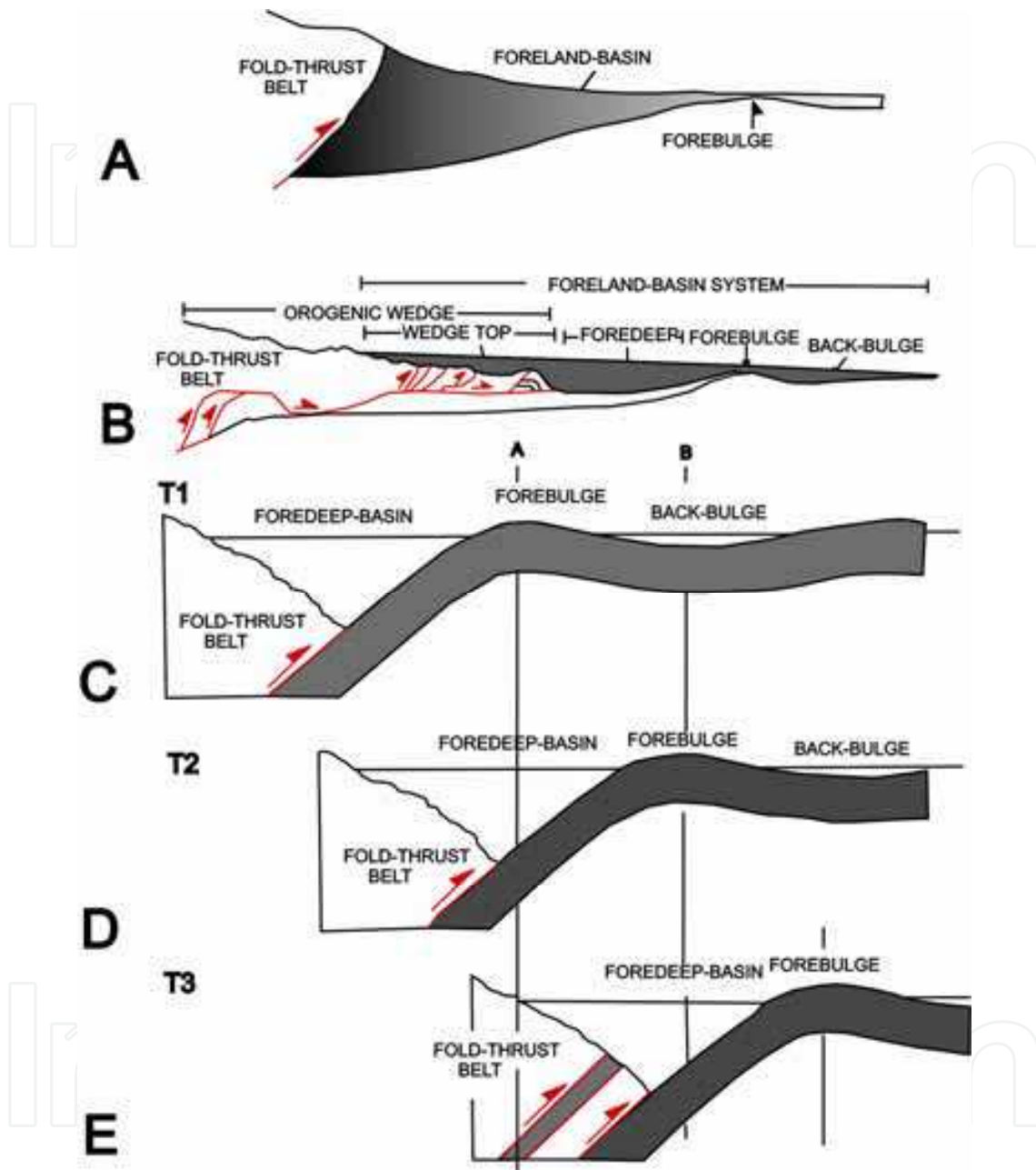


Fig. 1. Diagrammatic cross sections showing the generally accepted notion of foreland-basin geometry (a, b) and the relationships of lithospheric flexure to accommodation space in foreland systems (c to e) (e.g. Giles and Dickinson, 1995; De Celles and Giles, 1996). a) General relationship between fold-thrust-belt, foreland basin and forebulge; b) foreland-basin geometry and depozones: wedge-top, foredeep, forebulge and back-bulge depozones; c to e) relationship of the flexural features in times; c) is the initial (Time1) foreland system; d) foreland evolution during accretion of the fold-thrust-belt at Time 2; forebulge is migrated cratonward; e) previous forebulge is assembled within the fold-thrust-belt. Modified after Giles and Dickinson (1995), and DeCelles and Giles (1996)

Foreland basin stratigraphy records tectonic, eustatic, and climatic changes at convergent plate margins (e.g. Miall, 1995). The formation of unconformities is the results of the interplay of temporal variations in the erosion and lateral progradation rates of the orogenic wedge, as well as tectonic and eustatic sea-level changes (e.g. Beaumont, 1981; Jordan, 1981; Schedl and Wiltschko, 1984; Peper *et al.*, 1995).

## 2. The clastic infill of Foreland basin systems

### 2.1 Depositional zones and geometries

In foreland settings, subsidence and uplift are profoundly affected by lithospheric flexure. Foreland basin subsidence is primarily controlled by downflexing of the lithosphere in response to thrust accommodation and loading (e.g. Jordan, 1981, 1995; Beaumont, 1981). Subsidence rate gradually decreases away from the thrust front producing an asymmetrical depression. Flexure uplift (forebulge) occurs as an isostatic response to downwarping and forms the distal margin of the foreland basin. Cratonward of the forebulge flexure, a broad shallow downwarp or intrashelf basin forms, the back-bulge basin (Fig. 1; e.g. Quinlan and Beaumont, 1984; DeCelles and Giles, 1996).

The dimension and amount of flexural subsidence and uplift produced by the flexural features (i.e., foreland basin, forebulge, back-bulge basin) primarily depend on the geometry and density of the tectonic load, rheology of the lithosphere, density and volume of the sediment infill, and amount of thrust wedge and forebulge erosion (e.g. Beaumont, 1981; Jordan, 1981; Vai, 1987; DeCelles and Giles, 1996; Sgroso, 1998). The interrelationships between lithospheric flexure, single thrust accommodation within the accretionary wedge and flexural subsidence experiences geometrically complex entities within the foreland region. The foreland basin system may be divided into four depozones, the wedge-top, the foredeep, the forebulge, and the back-bulge depozones (Fig. 1; e.g. DeCelles and Giles, 1996). Boundary between depozones may shift laterally through time following the deformation propagation. The longitudinal dimension of the foreland basin system is roughly equal to the length of the adjacent fold-thrust belt (e.g. DeCelles and Giles, 1996).

*Wedge-Top Depozone* - Large amounts of syntectonic sediment cover the frontal part of the fold-thrust-belt. The sediment that accumulates on top of the frontal part of the orogenic wedge constitutes the wedge-top depozone. Its extent toward the foreland is defined as the limit of deformation associated with the frontal tip of the underlying orogenic wedge. The main distinguishing characteristics of wedge-top deposits are the abundance of progressive unconformities (e.g., Riba, 1976) and various types of growth structures (folds, faults, cleavages; Boyer and Elliot, 1982; Cello and Nur, 1988; Srivastava and Mitra, 1994; DeCelles and Giles, 1996; Zuppetta and Mazzoli, 1997). Aerially extensive aprons of alluvial sediment or shallow shelf deposits commonly drape the upper surface of the orogenic wedge during periods the wedge is not deforming in its frontal part, and large, long-lived feeder canyons may develop and fill in the interior parts of orogenic wedges (e.g. Ori *et al.*, 1986; DeCelles and Giles, 1996). Sediments of the wedge-top depozone typically reflect the erosion and unroofing of the thrust-belt (e.g. Critelli and Le Pera, 1994, 1995a, 1998; Trop and Ridgway, 1997).

*Foredeep Depozone* - It is the mass of sediment that accumulates between the frontal tip of the orogenic wedge and the forebulge. Foredeeps are typically 100-300 km wide and 2-8 km thick (e.g., DeCelles and Giles, 1996). Sediment is derived predominantly from the fold-thrust-belt, with minor contributions from the forebulge and craton (e.g. Schwab, 1986; DeCelles and Hertel, 1989; Critelli and Ingersoll, 1994; Critelli and Le Pera, 1998). Foredeep depozones have frequently recorded transitions from early deep-marine sedimentation

("Flysch") to late coarse-grained, nonmarine and shallow-marine sedimentation ("Molasse") (e.g. Covey, 1986; Crook, 1989; Sinclair and Allen, 1992). The transition from "Flysch" to "Molasse" most likely reflects the fact that foreland basin systems originate as oceanic trenches (or remnant ocean basin) and later become shallow marine or nonmarine as continental crust enters the subduction zone (e.g. Ingersoll et al., 1995; DeCelles and Giles, 1996).

*Forebulge Depozone.*- It consists of the region of potential flexure uplift along the craton side of the foredeep. Because of forebulge depozone is a positive, and potentially migratory, feature, which may be eroded, its potential of preservation is low. One signal of the presence of an ancient forebulge may be the erosional unconformity surface. The forebulge generally is considered to be a zone of nondeposition or erosion, and the resulting unconformity may be used to track its position through time (e.g. Vai, 1987; Bosellini, 1989; Crampton and Allen, 1995; DeCelles and Giles, 1996; Sgrosso, 1998; Critelli, 1999).

In subaerial foreland basin systems (in which foredeep is not filled to the crest of forebulge) the forebulge may be a zone of erosion, with streams draining both toward and away from the orogenic belt (Crampton and Allen, 1995). If sediment derived from thrust-belt progrades into the forebulge, a thin condensed fluvial and aeolian sediment is deposited (DeCelles and Giles, 1996; Critelli, 1999).

In subaqueous foreland basin systems (in which foredeep is not filled up to the crest of forebulge), local carbonate platforms may develop in the forebulge depozone; extensive forebulge carbonate platforms and ramps can connect the foredeep with the back-bulge depozone (Giles and Dickinson, 1995; Critelli, 1999).

*Back-bulge Depozone.*- It constitutes the sediment that accumulates between the forebulge depozone and the craton. Sediment contributions from the craton and development of carbonate platforms may be significant in submarine systems (e.g., DeCelles and Giles, 1996). Stratigraphic units in the back-bulge are generally much thinner than those in the foredeep, and consist of dominantly shallow-marine and nonmarine sediments.

## 2.2 Compositional signatures of Foreland clastics

*Lithologic Provenance Models.* — Foreland regions are one of the typical settings in which huge volumes of clastic sediments are rapidly accumulated. Provenance studies in this tectonic setting have long been used to contribute to the complex history of the basin evolution, sediment dispersal pathways, dating major thrust events, and the unroofing history of the thrust-belt (e.g. Wiltschko and Dorr, 1983; Graham et al., 1986; Dickinson, 1988; Jordan et al., 1988; Steidtmann and Schmitt, 1988; Critelli, 1999). Tectogenic sediments may be shed as alluvial fans in front of rising thrust sheets, and the age of these sediments may indicate the time of motion on some faults. In this setting, the uplift-erosion-transport-deposition system are genetically and intimately related to the style of deformation in thin-skinned, thrustured terrains. Transport of clastic sediment in the same direction as tectonic transport is the commonly assumed setting for the clastic-wedge/ thrust association (e.g. Graham et al., 1986; Jordan et al., 1988; Steidtmann and Schmitt, 1988), that is named as «synthetic dispersal». However, opposite sediment dispersal pathways with respect to the tectonic transport is named as «antithetic dispersal» (e.g. Steidtmann and Schmitt, 1988). The result is that sediment dispersal pathways in foreland basin systems are controlled by geometries within the thrust sheet system, as frontal ramps, lateral ramps, diverse hanging-wall beds dip. In settings where distinct source-rock compositions are eroded sequentially, as in the case of predominantly vertical uplift of a stratigraphic section, «unroofing sequences» are commonly formed in the resultant clastic wedge (e.g. Graham et al., 1986; DeCelles, 1988; Steidtmann and Schmitt, 1988). This erosional inverted clast stratigraphy can provide valuable

information about the evolving source and the identification of specific source areas (e.g. DeCelles, 1988; Colombo, 1994; Critelli and Ingersoll, 1994; Critelli *et al.*, 1995; Critelli and Le Pera, 1998). In the case of thin-skinned thrust terrains (where horizontal transport dominates), layered rocks having different lithologies are exposed to erosion as they pass over a ramp providing a blended clastic dispersal of the exposed rock types. The resulting clastics may show no unroofing sequences, but include the same blended clast composition for relatively great thickness. These blended clastics may indicate that the source rocks were formed by tectonic transport over a ramp (e.g. Steidtmann and Schmitt, 1988). In thin-skinned thrust belts, both «unroofing sequences» and «blended clastics» can result in combinations.

*Large and Regional-Scale Models based on Sand(stone) Petrofacies.* — Numerous studies have demonstrated that sand (stone) from foreland basins are characterized by high framework percentages of quartz and unstable sedimentary and metamorphic lithic fragments, and the mean composition is quartzolithic (e.g. Dickinson, 1985, 1988; Schwab, 1986; DeCelles and Hertel, 1989; Critelli and Ingersoll, 1994; Critelli, 1999; Critelli *et al.*, 2003). These studies provide a basis for interpretations of tectonic setting from sand (stone) composition. Sandstone petrofacies can be considered a reliable general guide to the overall tectonic settings of most sediment provenances; and although many processes may modify the composition of sedimentary detritus, the fundamental imprint of provenance tectonics is preserved in the final sedimentary products. The key petrofacies of various sedimentary basins occur most typically when transport is short and direct. In other cases sandstone petrofacies in many sedimentary basins have multiple sources showing complex paleotectonic and paleogeographic relationships to the basins (e.g. Dickinson, 1988). The foreland basin systems are a typical basin-setting in which multiple sources can be active in the same time, and the derivative sandstones may show mixed petrofacies (fig. 2; Dickinson *et al.*, 1986; Schwab, 1986; Critelli, 1999). Schwab (1986), in a general statement of foreland-basin sandstone petrofacies, testifies the complex pattern of provenance relationships during the foreland basin evolution. Quartzose sand is typical during the early stage of foreland infill, where the thrust-belt has low elevation and consequently supplies low amounts, whereas cratonal region is flexing and supplies more amounts (Dickinson *et al.*, 1986; Cazzola and Critelli, 1987). Subsequent petrofacies is typical quartzolithic, when the thrust-belt is growing. Local provenances from magmatic arcs, uplifted subduction complexes or uplifted carbonate rocks of the forebulge only represent small amounts of the clastic record within foreland basin system. If thrust belt has severe uplift exposing the crustal basement, petrofacies can evolve to quartzofeldspathic sand (e.g. Critelli and Ingersoll, 1994; Garzanti *et al.*, 1996; Critelli and Le Pera, 1998; Critelli and Reed, 1999; Critelli, 1999; Critelli *et al.*, 2007; Barone *et al.*, 2008).

### 2.3 Present-day morphotectonic zones, Foreland basin system

From south to north the Southern Italy is subdivided into the following morphotectonic belts (Ippolito *et al.*, 1975): (1) The northern Calabrian Arc, including ophiolites, crystalline basement rocks and Mesozoic sedimentary sequences; (2) the Cilento and Calabro-Lucanian Ranges, having ophiolitic, metasedimentary and sedimentary rocks. The Ranges include a Paleogene Subduction Complex (the Calabro-Lucanian Flysch Unit or Liguride Complex of southern Italy), the middle Miocene foreland strata of the Cilento Group and younger sequences, and the Mesozoic to Miocene carbonate platform and slope (inner platform or Alburni-Cervati-Pollino Units and the Monti della Maddalena Unit); (3) the Campano-Lucanian Ranges, including Mesozoic to upper Miocene deep-sea sequences of the Lagonegro and Sicilide units, the outer platform sequences (Monte Alpi Unit), and the

Miocene foreland strata; (4) the Lucanian-Apulia lowland, including the Pliocene to Quaternary foreland clastics; and (5) the Apulian Swell, a Mesozoic to Quaternary carbonate platform (external platform).

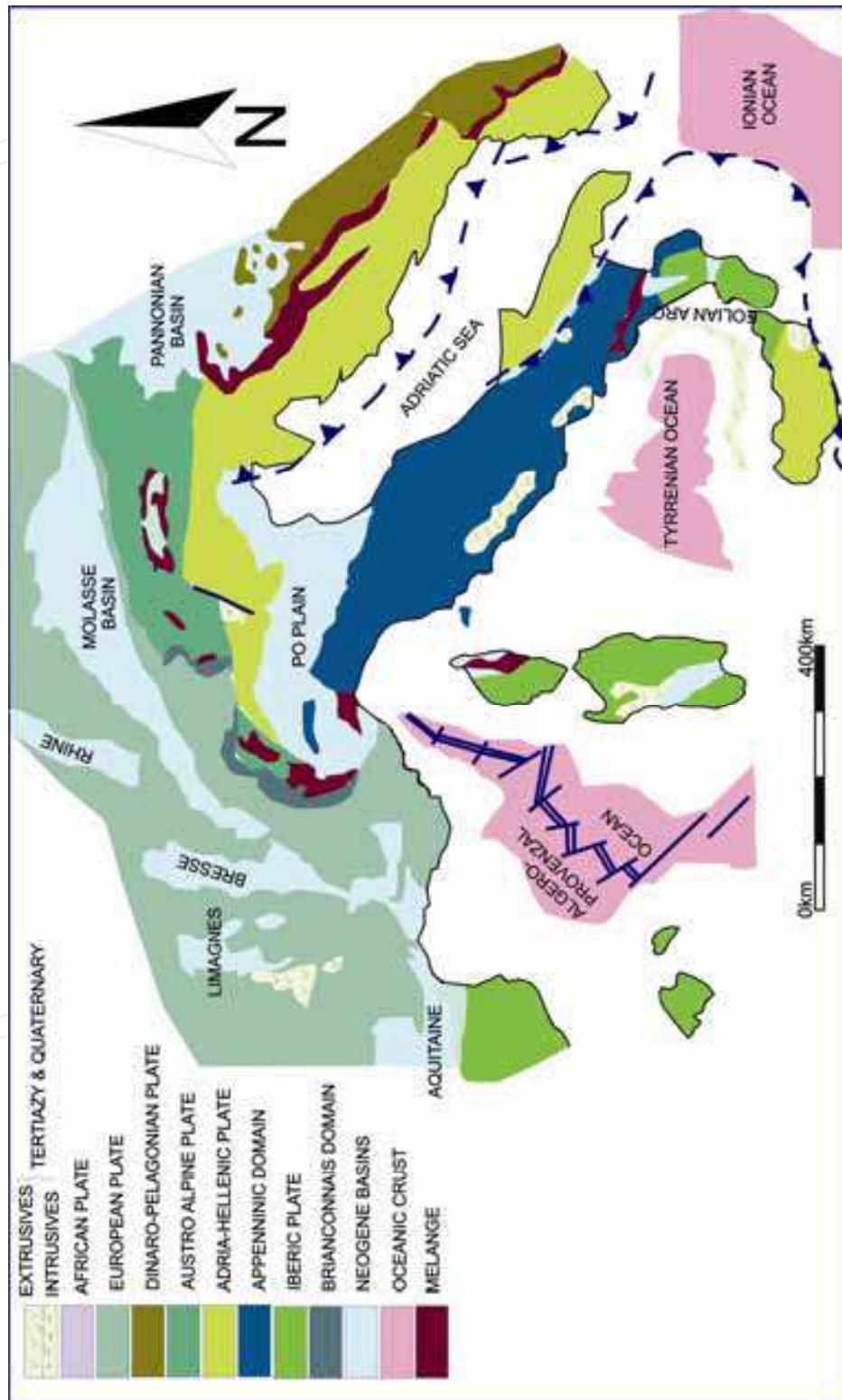


Fig. 2. Present day distribution of the main geodynamic domains of the Alpine region. Modified after Stampfli and Marchant (1997)

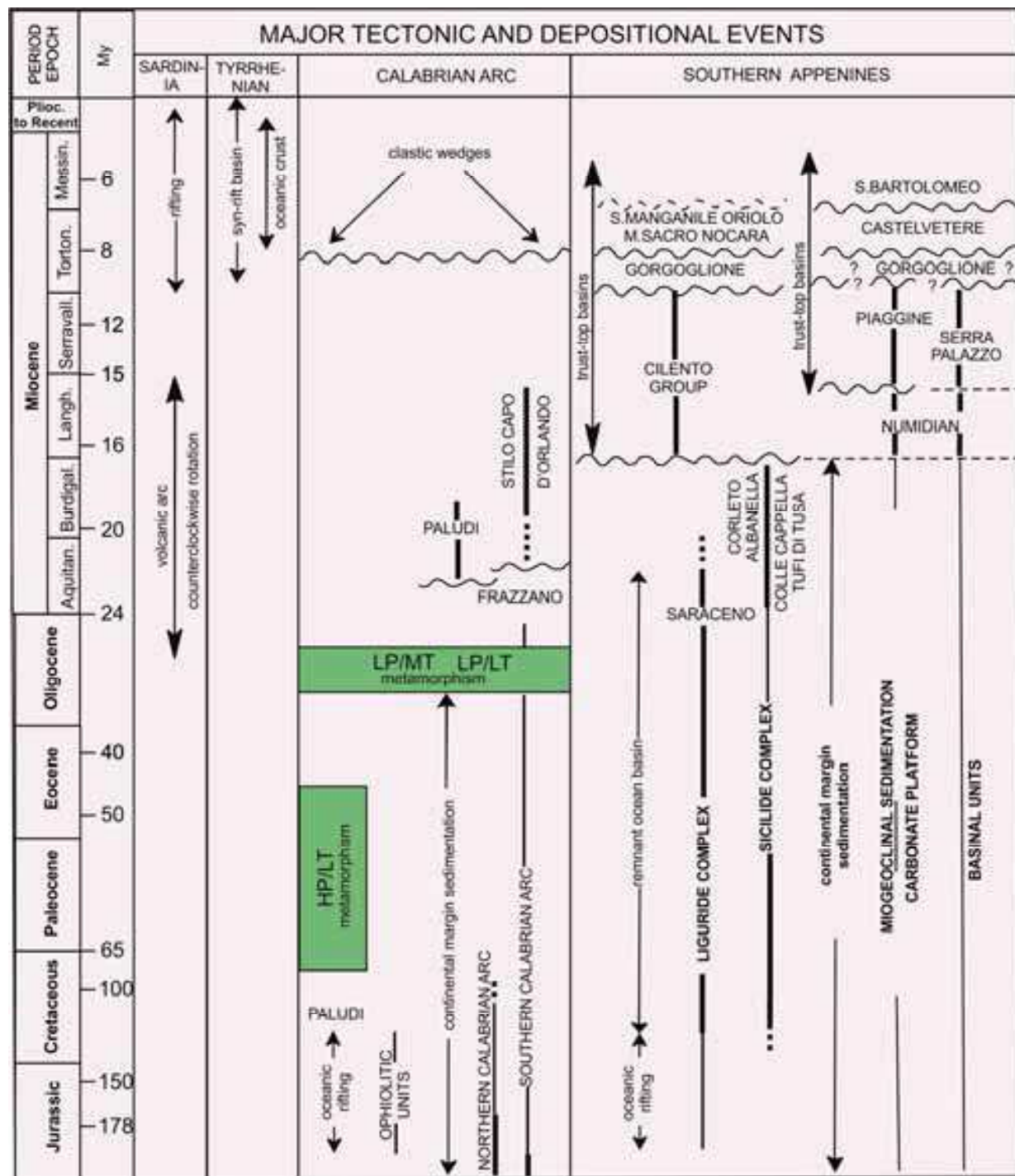


Fig. 3. Chart showing major Mesozoic-Cenozoic tectonic and depositional events in southern Italy sedimentary assemblages. Modified after Critelli and Le Pera (1995a), Critelli et al. (1995b) and Critelli (1999)

**2.4 Approach and scope**

This paper presents results of regional, structural, stratigraphic and provenance relationships that constraint the post-Oligocene tectonic history of the southern Apennines foreland basin system (Fig. 2). The paper focus on the effects of tectonic deformation during sequential history of the growing orogen in southern Italy.



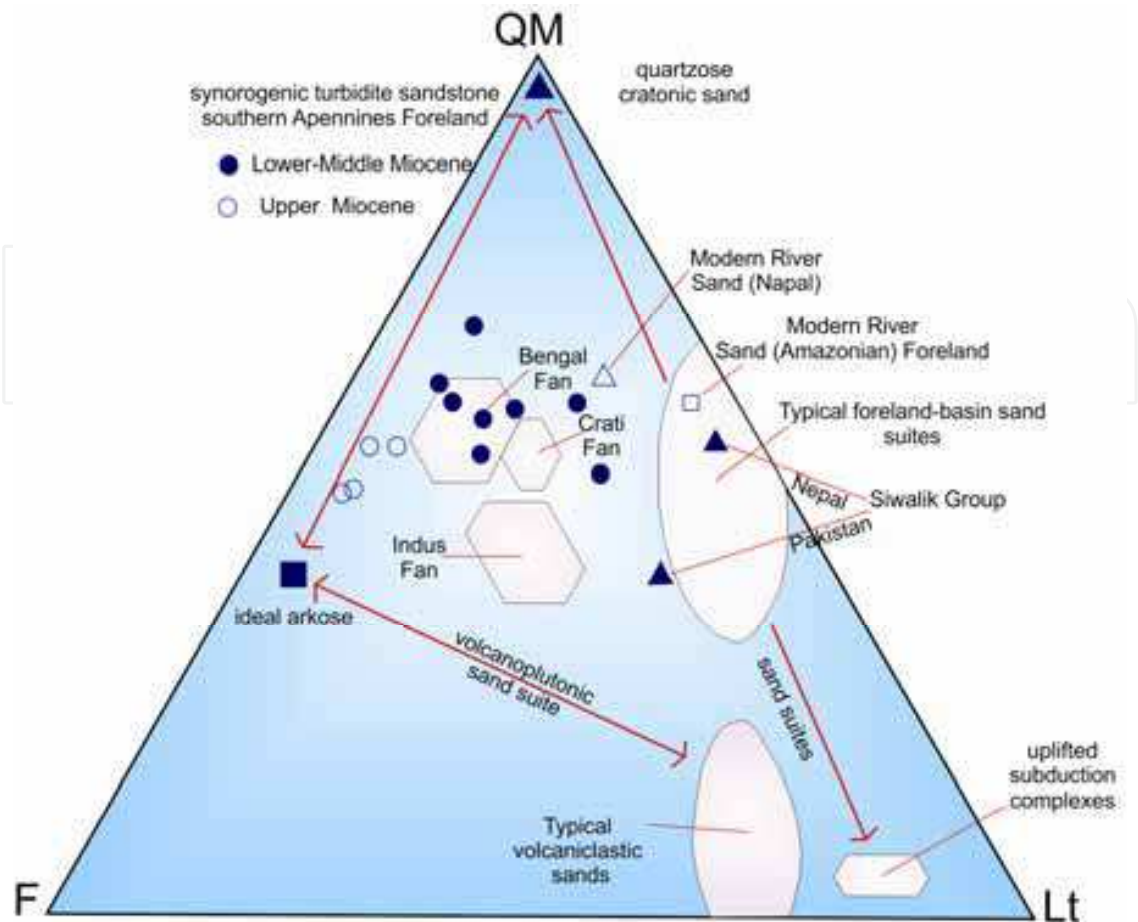


Fig. 4. QmFLt (Qm = monocrystalline quartz, F = feldspars, Lt = aphanitic lithic fragments) diagram to illustrate concept of mixing detritus from different provenance types to produce detrital modes reflecting mixed provenance (from Dickinson, 1988). Typical foreland-basin sand suites were derived from uplifted fold-thrust belts exposing sedimentary and metasedimentary strata. The mixed provenance relations are also typical of some foreland basin systems and remnant ocean basins (i.e. southern Apennines foreland, Indus and Bengal fans of the Himalayan belt). During early stage of foreland infill, sand may derive from cratonic areas, generating quartzose sand. Subsequent petrofacies is quartzolitic, and during final foreland infill, petrofacies may be mixed and quartzofeldspathic. Foreland sandstone detrital mode reference data plotted are: the modern Amazonian foreland (open square; data from DeCelles and Hertel, 1989), the Himalayan foreland [Siwalik (filled triangle) and modern rivers (open triangle); data from Critelli and Ingersoll, 1994], the Bengal and Indus Fan (polygons; data from Garzanti et al., 1996), and the southern Apennines foreland (filled and open circles for Miocene sandstone, and polygon for Holocene Crati Fan; data from Critelli and Le Pera, 1994, 1995). The arrows, symbols (filled triangle and square) and specific fields within the diagram show the sand suite trends from different generic types of provenance terranes (e.g. Dickinson, 1985, 1988; Critelli, 1999)

The entire stratigraphic, structural and compositional data set are interpreted using new general models of sequential evolution of foreland basin systems.

The Calabrian terranes form an arcuate mountain belt that lies between the thrust belts of the Apennines to the north and the Maghrebides to the west (Fig. 2). The study area is a transect across the Calabria block and Apulia platform (Figs.5,11).

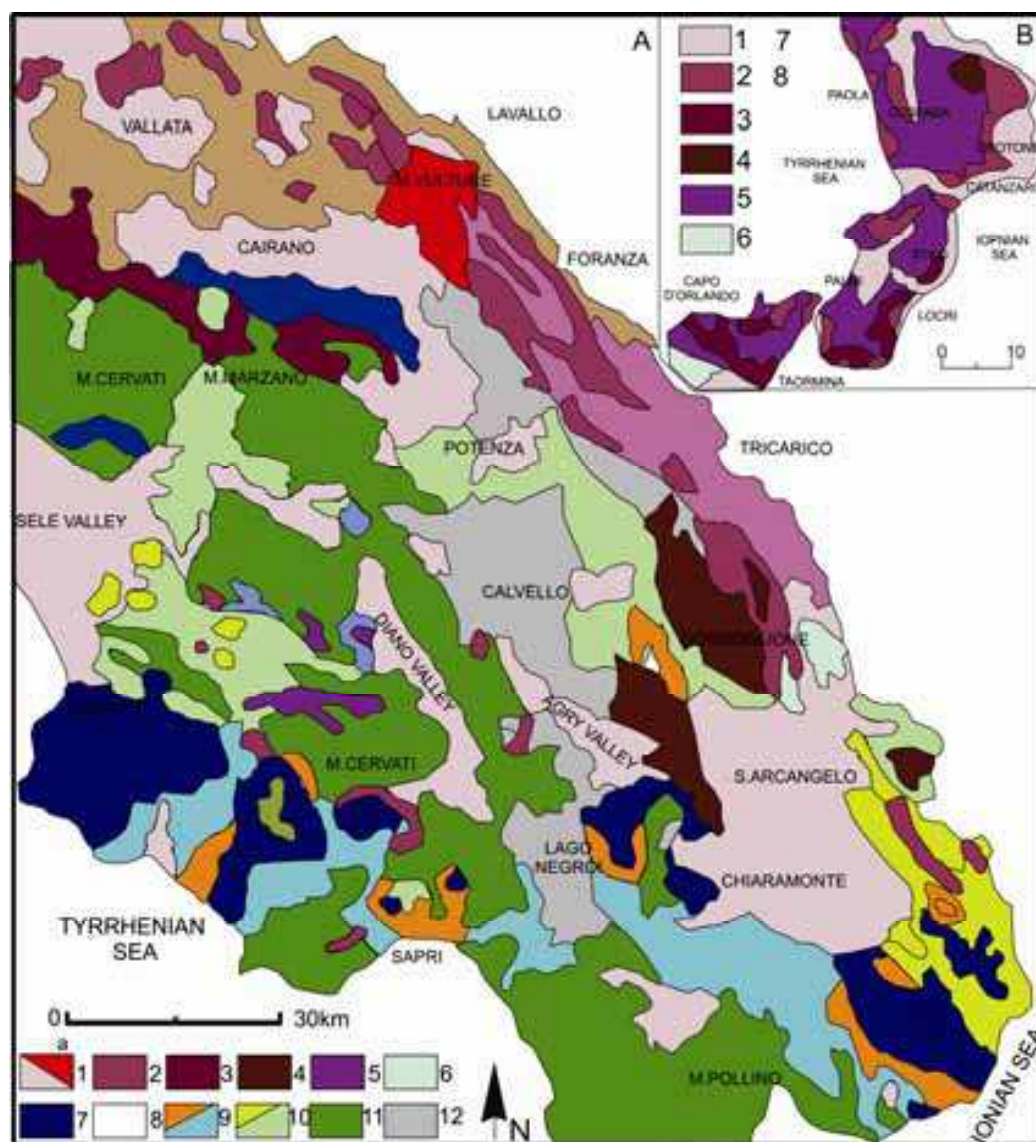


Fig. 5. Geological sketch map of the main tectonostratigraphic units of the Southern Apennines (A) and the Calabria-Peloritani Arc (B). A] 1) Lower Messinian to Holocene sediments (a: Monte Vulture volcanic and volcanoclastic rocks); 2) San Bartolomeo Formation (Messinian); 3) Castelvetere, Oriolo, Monte Sacro, Nocera, Serra Manganile formations (upper Tortonian to lower Messinian); 4) Gorgoglione Formation (Tortonian); 5) Piaggine Formation (Serravallian to Tortonian); 6) Serra Palazzo Formation (Langhian to Tortonian); 7) Cilento Group (Langhian to Tortonian); 8) Numidian Sandstone Formation (Langhian); 9) Liguride Complex (Cretaceous to early Miocene; a: Saraceno Formation); 10) Sicilide Complex (Jurassic to early Miocene; a: Albanella, Corleto, Colle Cappella, Tufiti di Tusa formations); 11) Shallow-water to deep-water carbonate units (Triassic to middle Miocene); 12) Deep-water pelagic sediments (Triassic to early Miocene; Lagonegro, Molise, Sannio units). B] 1) Pliocene to Holocene sediments and recent volcanic deposits; 2) Upper Tortonian to Messinian sediments; 3) Stilo-Capo d'Orlando Formation (early Miocene); 4) Longobucco Group (Jurassic); 5) Jurassic to Cretaceous Ophiolitic units, and Paleozoic metamorphic and plutonic units; 6) Maghrebic units (Mesozoic to Tertiary); 7) Paludi Formation (Late Oligocene to early Miocene); 8) Frazzan`o Formation (Oligocene to early Miocene). Modified after Critelli et al. (1995a,b) and Critelli (1999)

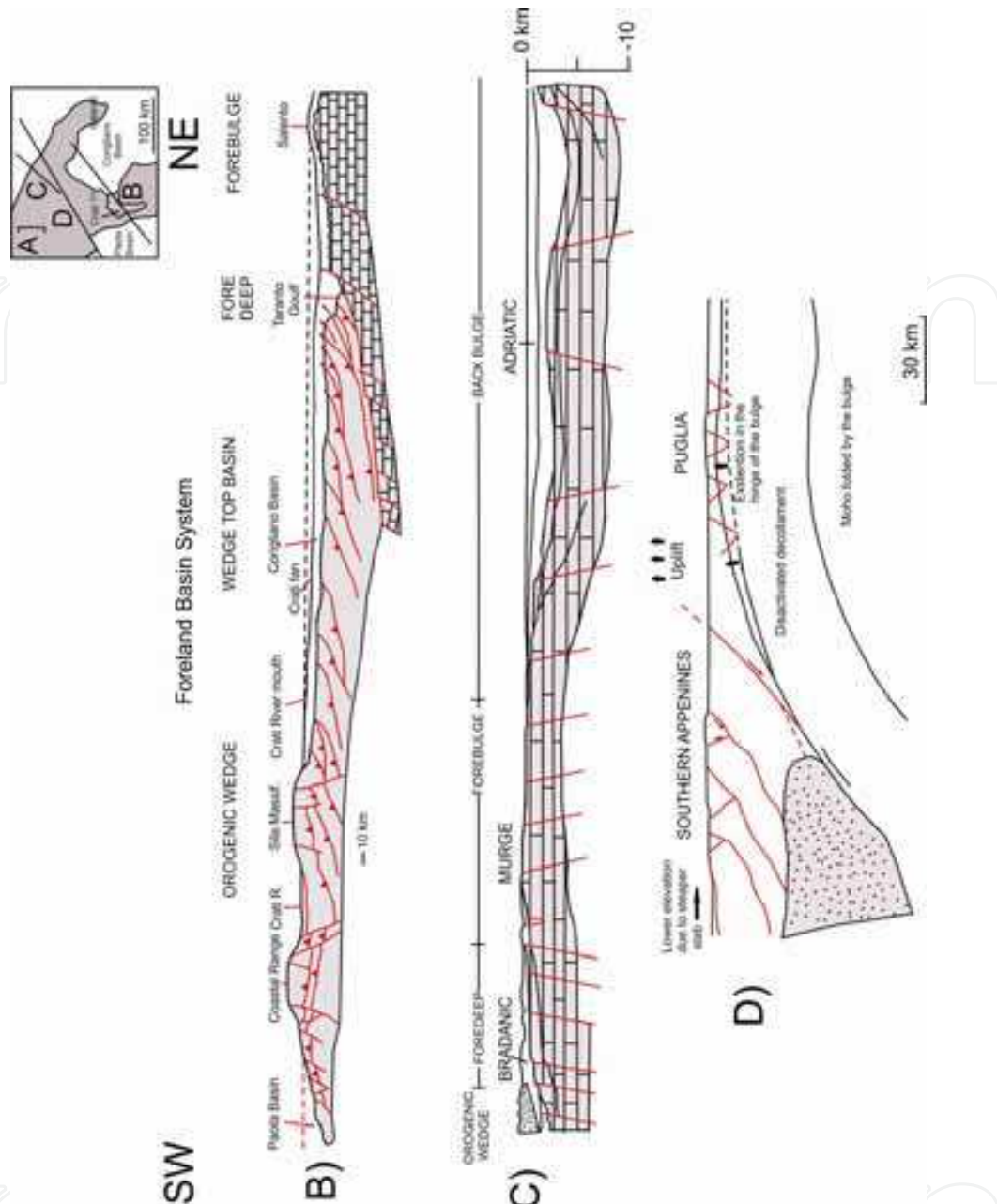


Fig. 6. Cross sections from the Adriatic Sea to the Tyrrhenian Sea, crossing the main depozones of the modern southern Apennines foreland basin system (Critelli, 1999); locations of the cross sections are in a). b) Cross section from the Paola Basin to the eastern Apulia, showing the Paola slope basin (Eastern Tyrrhenian margin), the Calabrian thrust-belt (orogenic wedge), the southern Apennines foreland region (wedge-top and foredeep basins) and the flexed Apulia foreland [Modified after Cello et al. (1981), and Pescatore and Senatore (1986)]. c) Cross section from the outer thrust front of the southern Apennines orogenic wedge to the southern Adriatic Sea; locations of the modern subaerial foredeep (Bradanic trough), forebulge (Murge), and back-bulge (southern Adriatic Sea) depozones are shown [Modified after Ricchetti (1980), and Ricchetti and Mongelli (1980)]. d) General schematic deep cross-section of the southern Apennines orogenic wedge showing formation of the Puglia bulge. During middle Pleistocene, the Bradanic trough was inverted from subsidence to uplift. Modified after Doglioni et al. (1994)

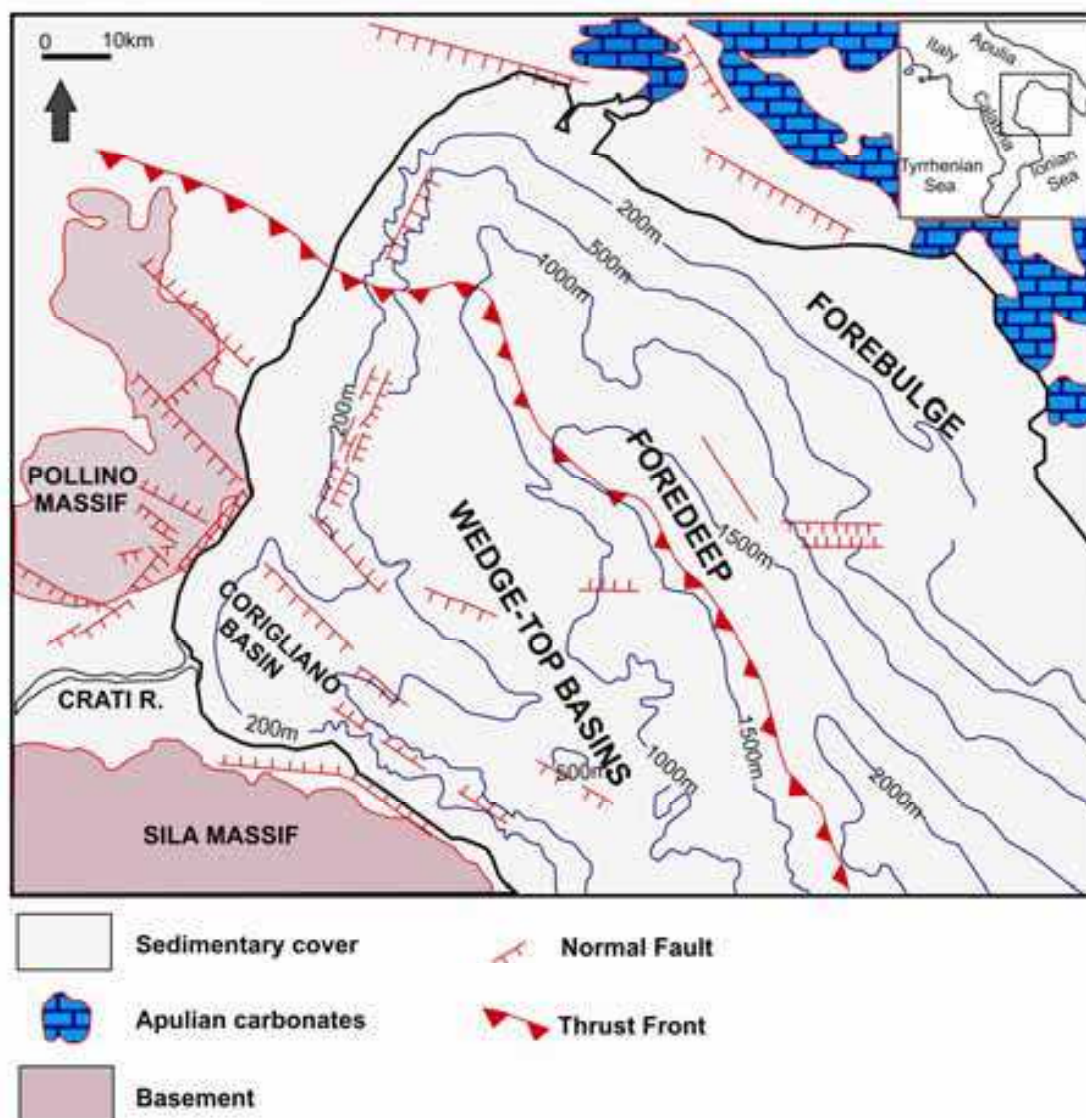


Fig. 7. Bathymetric and structural map of the Gulf of Taranto, which is the deep-marine part of the modern foreland basin system along the south-east side of the Apennines orogenic belt. The Corigliano basin represents the wedge-top depozone of the foreland region, in which the Crati Fan has been developing during the last 6000 y. Modified after Critelli and Le Pera (1998, 2003) and Critelli (1999)

The modern physiography and geology of Calabria are the results of post-30 Ma geodynamic processes in which synchronous accretionary processes were active along the eastern flank (northern Ionian Sea), and rifting processes along the western flank (Eastern Tyrrhenian Margin).

The subduction plane, as such as the southern Apennines and Calabrian accretionary prism, have migrated eastward or southeastward causing the roll-back of the subduction (e.g. Malinverno and Ryan, 1986; Royden et al., 1987; Doglioni, 1991; Gueguen et al., 1997, 1998). The roll-back of the subduction hinge (rate of hinge retreat is 6 cm/ y; Royden et al., 1987; Patacca et al., 1993) appears to have been slowed and buckled during the Late Pleistocene by the interference of the thick continental lithosphere of the Adria Plate (Apulian swell) at the front of the belt (e.g. Doglioni et al., 1994, 1996).

The frontal active accretionary wedge, below sea-level, whereas the main elevated ridge to the west is in uplift and extension instead. The modern basin configuration of this thrust belt is represented by the wedge-top depozone (Corigliano-Amendolara basins), the marine and subaerial foredeep depozone (Gulf of Taranto and the Bradano river basin, respectively), the forebulge (the Gallipoli Basin) and the back-bulge (southern Adriatic Sea) (Figs. 6,7; e.g. Critelli and Le Pera, 1998).

Several Pliocene-Pleistocene basins cross-cut the Apennines and northern Calabria thrust pile, the most important are the Vallo di Diano, Val d'Agri, Potenza Basin, Mercure Basin and Crati Basin (e.g. Turco et al., 1990; Cinque et al., 1993; Colella, 1994; Tortorici et al., 1995; Schiattarella, 1998; Tavarnelli and Pasqui, 1998).

On the backarc area similar fault-controlled Pliocene-Pleistocene basins (Tortorici et al., 1995), as such as the Paola Basin and Gioia Basin, represent the synrift troughs of the eastern Tyrrhenian margin (e.g., Savelli and Wezel, 1980; Barone et al., 1982; Sartori, 1982, 1990).

Respect of low elevation, some calculations (Doglioni et al., 1996; Gueguen et al., 1998) show that the thickness of sedimentary strata in the Apennines exceed 20-25 km, the entire crustal thickness is about 30 km, and a thick pile of synorogenic sediment accumulation (up to 10 km from Miocene to modern) suggesting a delamination of the lithospheric mantle during Apenninic subduction (e.g. Channel and Mareschal, 1989; Doglioni et al., 1996).

The Calabrian ranges is peculiar for their high uplift rates that are 1 mm yr<sup>-1</sup> (e.g. Cosentino and Gliozzi, 1988; Sorriso-Valvo, 1993; Westaway, 1993), where the maximum uplift is toward the frontal part of the accretionary prism (Ionian side) (Cosentino and Gliozzi, 1988).

## 2.5 Plate-tectonic evolution

The study area shows rocks which experienced a large series of geodynamic events occurred between early to middle Paleozoic orogenesis to actual.

The key geodynamic events into the Mediterranean region can be summarized as follow:

a) The Mesoalpine (Eocene to early Oligocene) tectonic phase in southern Italy corresponds with the subduction of the Adria-Ionian oceanic lithosphere beneath the Iberia plate (Fig. 8). This tectonic stage is responsible for the initial flexure, a general erosional processes of both the inner platform (Alburni-Cervati-Pollino-Bulgheria; Boni, 1974; D'Argenio, 1974) and outer platform (Monte Alpi-Apulia).

The Mesoalpine tectonic phase caused regional metamorphism at around 38 Ma (e.g. Steck and Hunziker, 1994), and intra-orogenic magmatism along the Periadriatic zone.

The Middle Oligocene (32-30 Ma) is characterized by intense magmatic activity, part of which is directly linked to the Algero-Provençal rift (Provence and Sardinia), part along the Insubric line and part along the periadriatic domain. In the Alps and northern Apennines, the Eocene and Oligocene siliciclastic sedimentary sequences record provenance from (a) Iberic plate (Corsica-Sardinia-Briançonnais), (b) Adria plate (austroalpine domain), (c) European plate, (d) syneruptive magmatic activity, and from (e) both European and Adria forebulges.

In the southern Italy domain, the Calabro-Lucanian Flysch Unit and the Sicilide Complex strata represent deposition in the remnant ocean basin related to the western subduction of the Adria oceanic lithosphere beneath the Iberia plate (Fig. 8; e.g. Knott, 1988; Dewey et al., 1989; Critelli, 1993; Guerrera et al., 1993; Critelli and Le Pera, 1998; Critelli, 1999). The subduction has been active for all the Paleogene and lower Miocene, producing an accretionary prism, the calabro-Lucanian Flysch Unit and the Sicilide p.p. Complex, and a diffuse calcalkaline volcanism in Sardinia. The Liguride Complex records the accretionary processes along the Adria margin and the consumption of the oceanic crust.

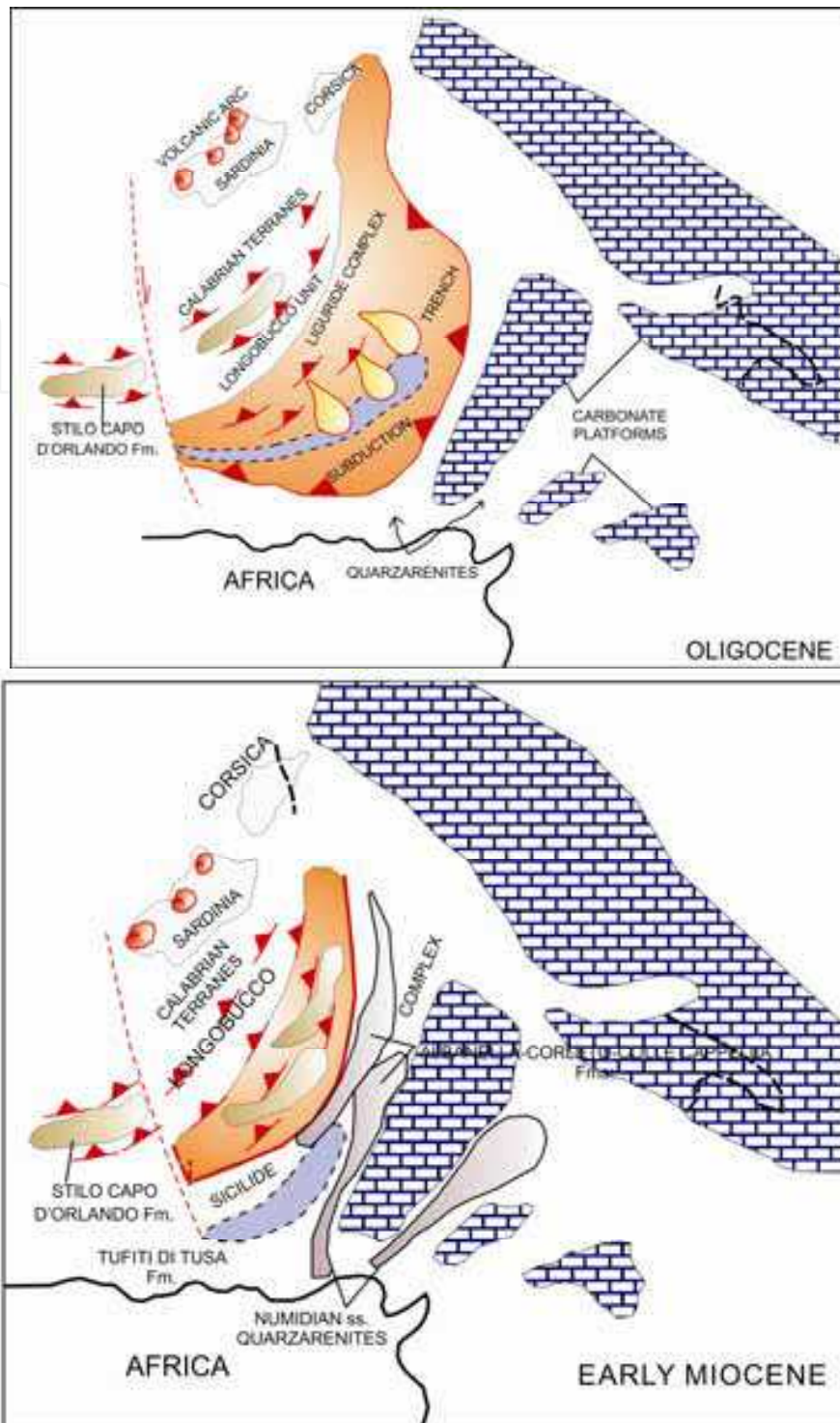


Fig. 8. Paleogeographic reconstructions (Oligocene to early Miocene) of the western Mediterranean during progressive closure of the remnant ocean basin and onset of collision in southern Apennines. Modified after Dewey et al. (1989), Critelli (1993, 1999) and Perri et al. (2011)

f) During early (Fig. 8) to middle (Fig. 9) Miocene the Apenninic domain is the place where immense volume of turbiditic sedimentation is in response of E-NE accretionary processes along the Adria plate (e.g. Ricci Lucchi, 1986; Patacca and Scandone, 1987; Boccaletti et al.,

1990). Here, the foreland basin system is developed over deformed Liguride Complex, during the early-middle Miocene, over Sicilide, Lagonegro and inner platform units during the upper Miocene (Fig. 10), over the previous units and the western margin of the Apulia platform during the Pliocene to Quaternary. The foreland basin system (wedge-top, foredeep, forebulge, back-bulge depozones) migrated in time, and siliciclastic and carbonatoclastic deposits, filling the wedge-top and the foredeep, where derived from progressive unroofing of the Calabrian crustal block or from erosion of the forebulge (e.g. Critelli and Le Pera, 1998).

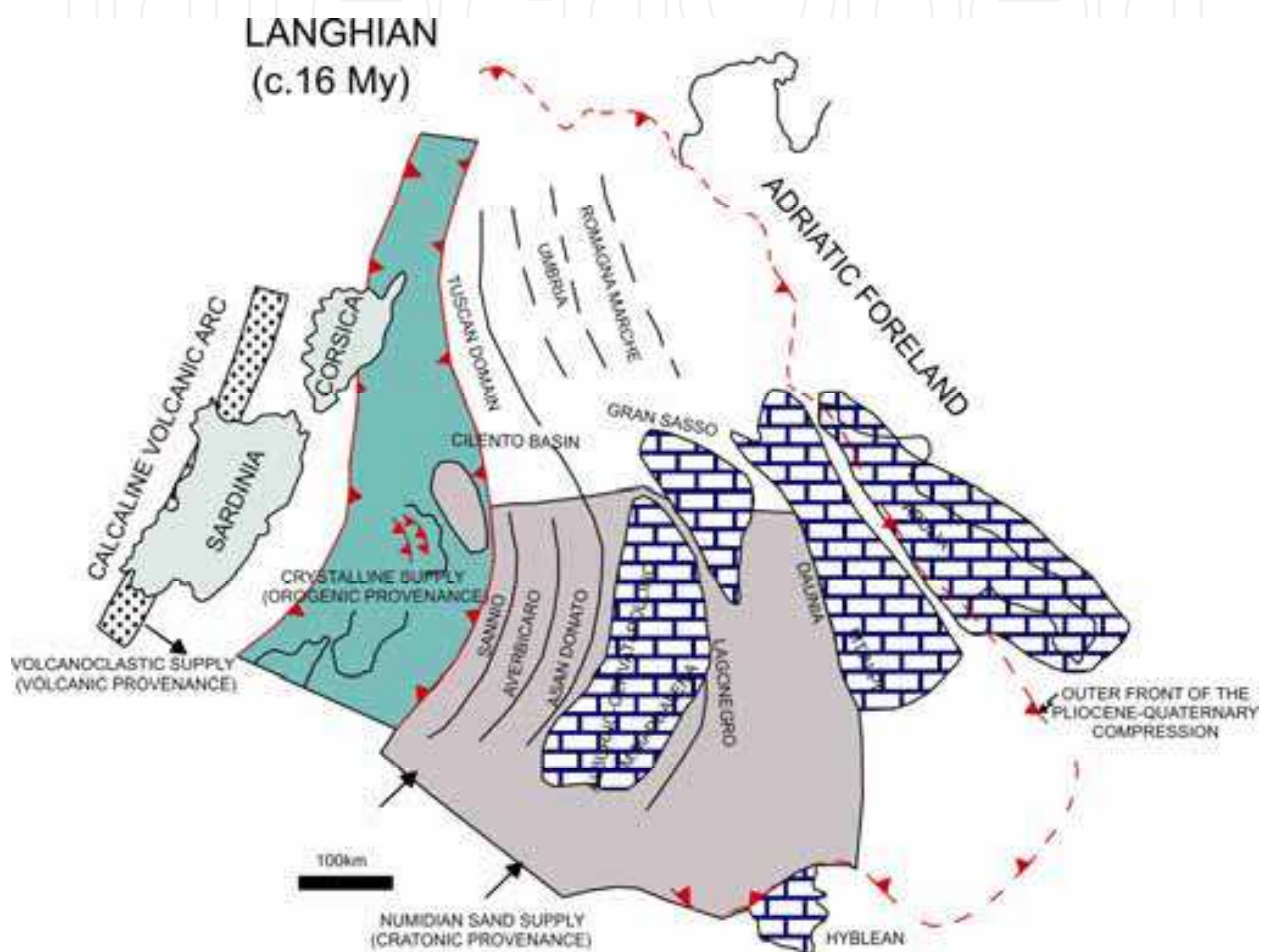


Fig. 9. Palinspastic restoration of the Apenninic domains during Langhian. Modified after Patacca et al.(1992) and Critelli (1999)

g) the geodynamic events of the last 10 My, in the western-central Mediterranean is named the Tyrrhenian phase (15-0 Ma) (Fig. 10). The Tyrrhenian phase (or back-arc extension) is responsible for the fragmentation and dispersion of pieces of the Iberian and European plates (Calabria, Sardinia, Corsica), increased the displacement of the accretionary prism over the Adria plate, the eastward migration of the magmatic arcs, and the roll-back of the Adriatic lithosphere (Malinverno and Ryan, 1986; Patacca et al., 1990, 1993; Argani et al., 1995; Doglioni et al., 1996; Gueguen et al., 1997, 1998). The Tyrrhenian backarc basin migrated eastward (northeastward in the northern Apennines and southeastward in Calabria and Sicily) at velocities of up to 5-7 cm/ yr in the most arcuate parts of the arc (Doglioni, 1991; Gueguen et al., 1998).

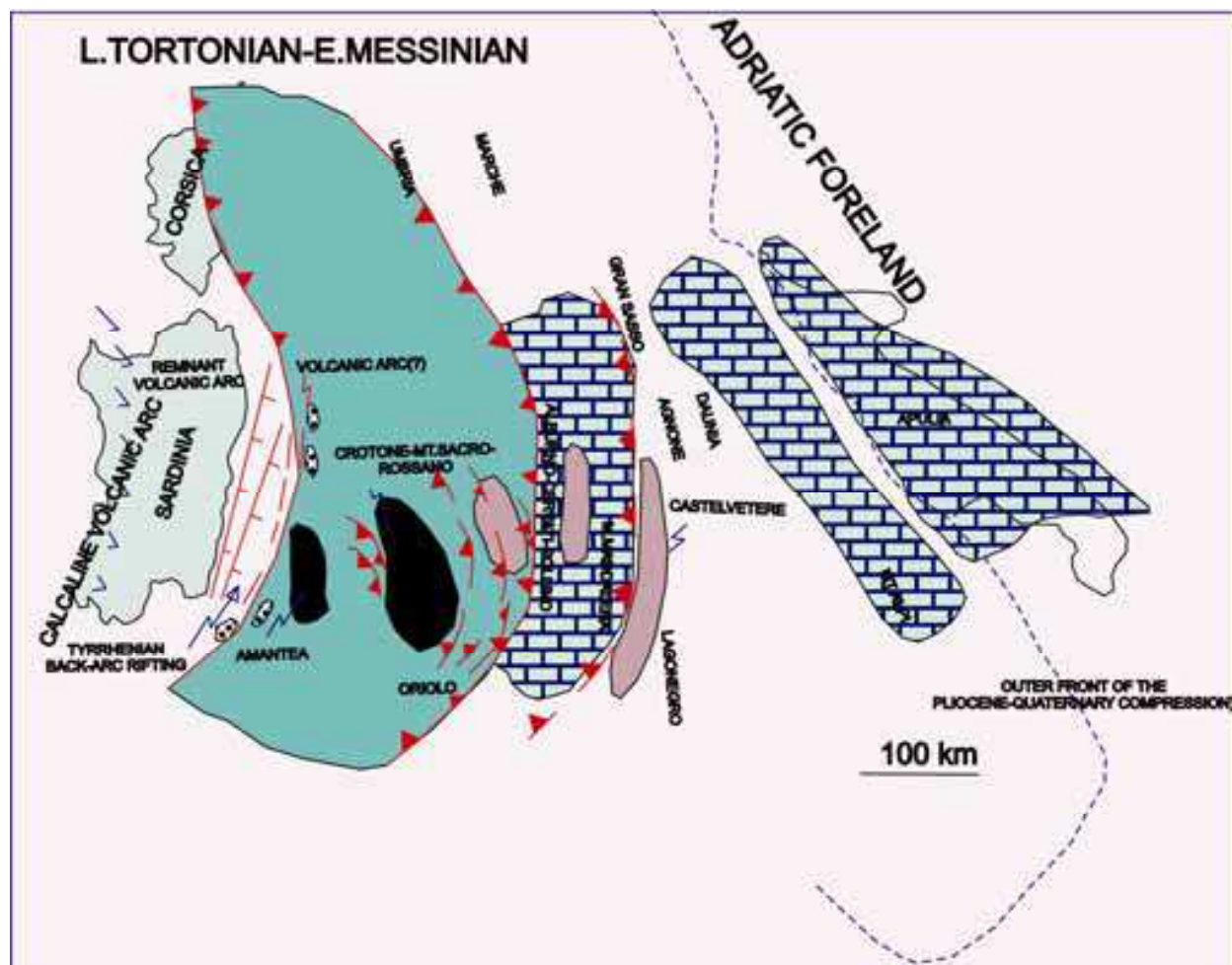


Fig. 10. Palinspastic restoration of the Apenninic domains during the late Tortonian to early Messinian (time interval 8-6 my). Initial back-arc rifting of the Tyrrhenian Sea, and diverse wedge-top and foredeep depozones in the southern Apennine foreland region. Modified after Patacca et al. (1990) and Critelli (1999)

## 2.6 Structural evolution and rise of the Calabrian terranes

The Paleozoic metamorphic and plutonic terranes of the Calabrian Arc represent the remnants of Caledonian, Hercynian and Alpine orogens (e.g. Amodio Morelli et al., 1976; Schenk, 1981; Zanettin Lorenzoni, 1982; Atzori et al., 1984; Del Moro et al., 1986; Zeck, 1990; Messina et al., 1994), that are drifted from the southern Iberic plate and accreted since upper Oligocene over the Adria-Africa lithosphere. They are a key tectonic element of the southern Italy orogen.

However, other authors consider the Calabrian basement terranes as a part of the Austroalpine domain of the African Plate (e.g. Haccard et al., 1972; Alvarez et al., 1974; Alvarez, 1976; Amodio Morelli et al., 1976; Scandone, 1979, 1982; Bonardi et al., 1982, 1993; Dercourt et al., 1986). In other alternative interpretations, the nappes of the Calabrian Arc originated from a microcontinent located between the European and African continents (e.g. Wildi, 1983; Guerrera et al., 1993; Critelli and Le Pera, 1998; Critelli, 1999; Mongelli et al., 2006; Perrone et al., 2006; Critelli et al., 2008; Perri et al., 2008; 2010) or the Calabrian-Arc terranes are the result of the amalgamation of three “crustal microblocks” (e.g. Vai, 1992).



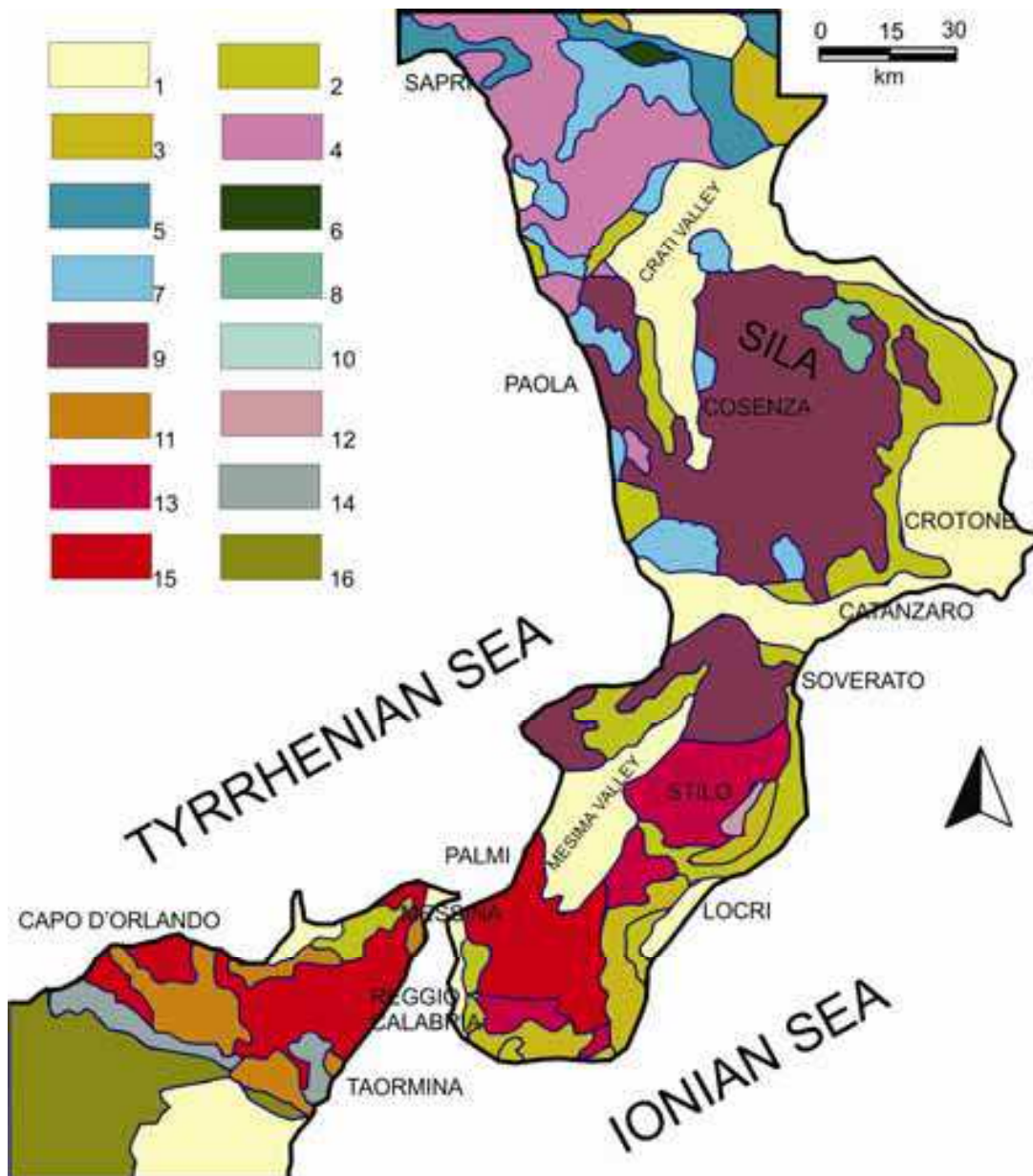


Fig. 11. Tectonic sketch map of the Calabria-Peloritani Arc. 1) Pliocene to Holocene sediments, and volcanic and volcanoclastic rocks; 2) Upper Tortonian to Messinian clastics and evaporites; 3) Cilento Group (Middle Miocene); 4) San Donato, Verbicaro and Pollino Units (Triassic to Miocene); 5) to 7) Liguride Complex: 5. Calabro-Lucanian Flysch Unit (Upper Jurassic to Upper Oligocene); 6. Ophiolitic blocks and Melange; 7. Frido Unit (Upper Jurassic to Upper Oligocene); 8) Longobucco and Caloveto Groups (Lower Lias to Lower Cretaceous) and Paludi Formation (Upper Oligocene); 9) Sila, Castagna and Bagni basement Units (Paleozoic); 10) Malvito, Diamante-Terranova, Gimigliano Ophiolitic units (upper Jurassic to Lower Cretaceous); 11) Floresta Calcarenite (Middle Miocene), Stilo-Capo d'Orlando Formation (Lower Miocene); 12-13) Stilo Unit: 12. Carbonate rocks of the Stilo Unit (Upper Triassic? to Cretaceous) and 13. Basement rocks (Paleozoic); 14) Sedimentary Cover of the Longi-Taormina Unit (Upper Triassic to Oligocene); 15) Basement rocks (Paleozoic) of the Aspromonte, Africo, Mandanici, Fondachelli, Longi, Taormina units; 16) Sedimentary units of the Maghrebain Chain. Modified after Bonardi et al. (1993), and Critelli (1999)

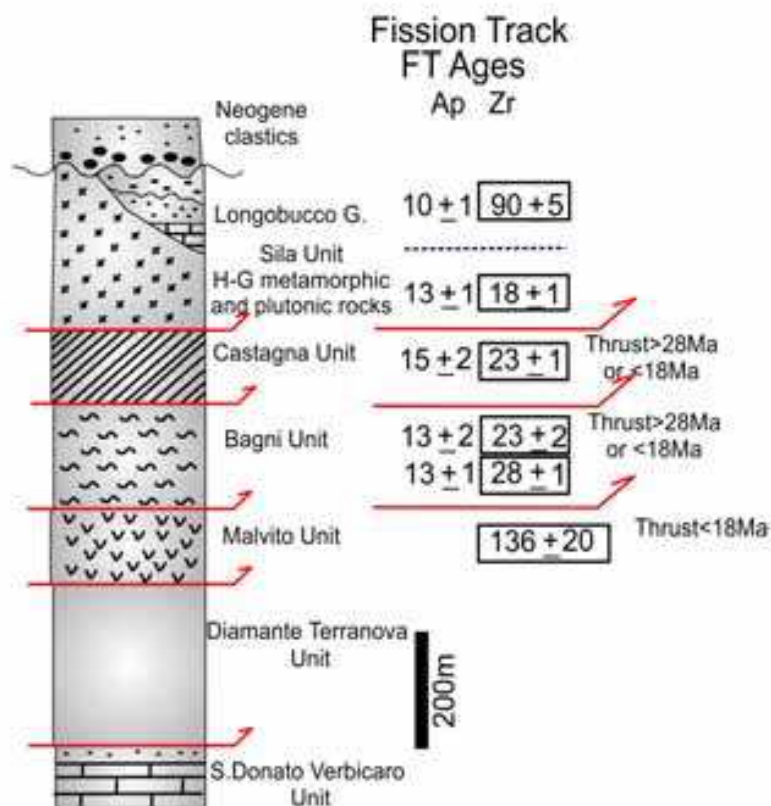
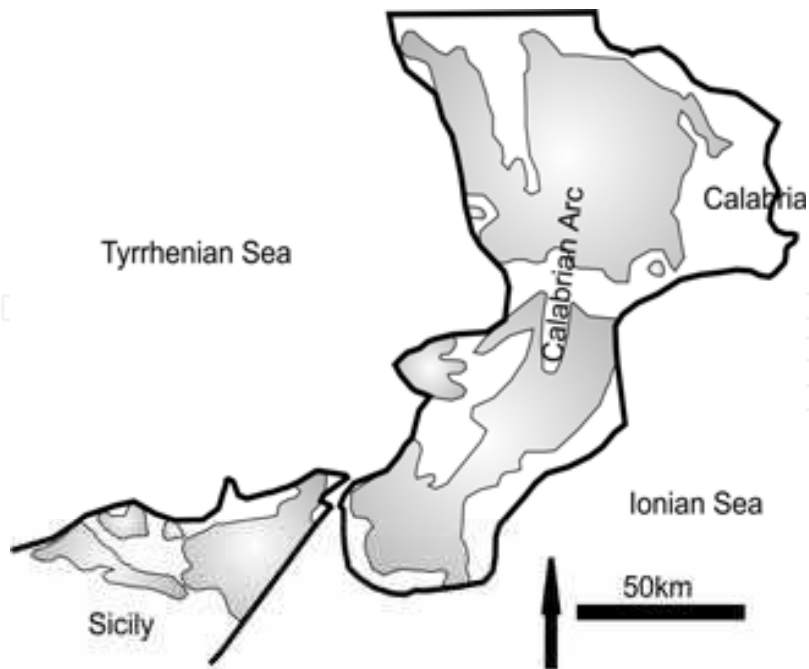


Fig. 12. Sketch map of the Calabria-Peloritani Arc, showing the outcrop of the ophiolitic and crystalline basement rocks, and simplified tectonostratigraphic terranes of the northern Calabrian Arc. Simplified tectonostratigraphy of the nappe sequence is modified after Amodio Morelli et al. (1976). Data of Apatite and Zircon Fission-Track ages is from Thomson (1998). Data on fission-track ages of the key thrust terranes constraints times of tectonic accommodation of the nappe piles. See Thomson (1998) for further details

The northern Calabrian Arc can be divided into three stacked tectonostratigraphic assemblages (Figs. 11, 12; e.g. Amodio Morelli et al., 1976; Bonardi et al., 1976; Scandone, 1979, 1982; Cello et al., 1981; Bonardi et al., 1982; Colonna and Compagnoni, 1982; Colonna, 1998). The lowest is made of mainly carbonate rocks of Mesozoic age (D'Argenio et al., 1973; Ietto and Barillaro, 1993; Iannace et al., 1995; Perrone, 1996; Ietto and Ietto, 1998) that were originally deposited on the continental margins of the Apulia/Adria plate (Channel et al., 1979). These sediments were stripped from their basement during the Early Miocene collision of Calabria with Africa and Adria, and now form part of the Africa-verging Apennine fold-thrust belt. The middle tectonic units are composed of two nappes (Diamante-Terranova, and Malvito units; Fig. 12) of Mesozoic to Cenozoic metasedimentary and ophiolitic rocks, which can be interpreted as the remains of the neo-Tethyan Ocean and related accretionary wedge (De Roever et al., 1974; Amodio Morelli et al., 1976; Lanzafame et al., 1979; Guerrera et al., 1993; Cello et al., 1996). The uppermost tectonic units consists of thrust sheets of Paleozoic igneous and metamorphic rocks (Bagni, Castagna and Sila units) and Mesozoic (Longobucco Group) to Cenozoic sediments (Figs. 11, 12), considered to be the basement and cover, respectively, of the former Iberian/ Europe margin of Neotethys (e.g. Ogniben, 1969, 1973; Bouillin, 1984; Bouillin et al., 1986; Knott, 1987, 1988; Dietrich, 1988; Dewey et al., 1989; Thomson, 1998).

Thomson (1998), with fission track studies, demonstrates that the emplacement of continental basement rocks with Alpine metamorphism over ophiolitic rocks is constrained as a thrust contact of lower-to-middle Miocene age (<23 Ma), and the other major thrust contact of the diverse alpine basement units may be <18 Ma (Fig. 12). The relative cooling ages range from 35 to 15 Ma, where most of this phase of accelerated cooling can be attributed to increased erosion and progressive exhumation since 23 Ma to about 10 Ma (Thomson, 1994, 1998).

### 3. Sequential history of growing orogen in Southern Italy and clastic sediments in space and time

#### 3.1 Pre-collisional and earliest collisional clastic units (Late Cretaceous to early Miocene)

In this large time occurred the main Alpine tectonic phases and final closures of the Piemontese-Ligurian oceanic basin to the north, whereas in the southern Italy, a more external remnant oceanic basin, the Lucanian oceanic basin, divides the Adria margin from the Mesomediterranean Microplate (e.g. Channel and Mareschal, 1989; Dewey et al., 1989; Guerrera et al., 1993; Critelli and Le Pera, 1998; Guerrera et al., 2005; Perrone et al., 2006; Critelli et al., 2008). The Adria plate experienced abrupt surficial and lithospheric changes, as changing nature of the pelagic basins, onset of siliciclastic sedimentation, huge volumes of cratonic quartzose sediments, emersion and erosion of carbonate platform domains, and deformation of the inner carbonate platform to form a forebulge (Patacca et al., 1992; Sgrosso, 1998). The oceanic lithosphere was subducting beneath the Mesomediterranean Microplate, with the Liguride basin representing the oceanic accretionary wedge and a diffuse calcalkaline volcanism was located in Sardinia.

*Mesomediterranean Microplate*— We include in the Mesomediterranean Microplate, the Calabria-Peloritani (CP, Fig. 13), Betic Cordillera (Alpujarrides, Rondades and Malaguides) and Rif (Sebtides, Tetouanides and Ghomarides) (AL, Fig. 13), Tellian Maghrebides (Grande and Petite Kabilies, Eudough, Algiers Massif) (KB, Fig. 13) (Wildi, 1983; Bouillin et al., 1986; Guerrera et al., 1993; 2005; Mongelli et al., 2006; Perrone et al., 2006; Critelli et al., 2008; Perri, 2008; Perri et al., 2008, 2010).

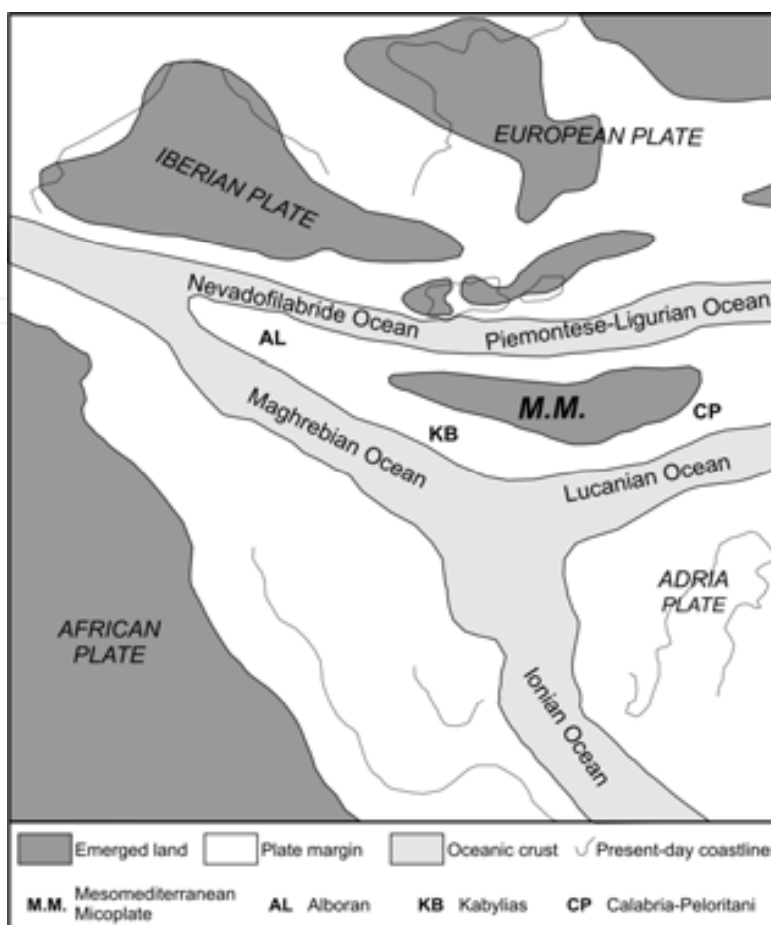


Fig. 13. Simplified Early Cretaceous palaeogeographical sketch of the Western Mediterranean Region (after Perrone et al., 2006 and Critelli et al., 2008, modified). Oceanic basins and margins around the Mesomediterranean Microplate are shown

During Paleogene to early Miocene, sediments over the Calabria-Peloritani Arc include the Upper Oligocene to lower Miocene Frazzanò Formation (de Capoa et al., 1997), in the Peloritani sector of the arc, and the upper Oligocene to lower Miocene Paludi Formation (northern sector) and Stilo Capo d'Orlando Formation (southern sector). These sandstones are quartzofeldspathic (Fig. 14; Zuffa and De Rosa, 1978; Puglisi, 1987; Cavazza, 1989; Nigro and Puglisi, 1993; Critelli et al., 1995b) and reflect their local provenance from crystalline rocks of the Calabrian terranes. The tectonic setting of these basins is complex; the sequences suturing some crystalline thrust units could represent a wedge-top deposition on advancing calabrian thrust-belt (e.g. Weltje, 1992; Patacca et al., 1993; Wallis et al., 1993) or may represent deposition in a forearc setting (e.g. Cavazza et al., 1997). An alternative interpretation is that they could represent remnants of deposition in foreland setting related to the back-thrust belt of the Betics-Alps orogen (e.g. Doglioni et al., 1997; Gueguen et al., 1997, 1998).

The oceanic area (Liguride basin) experienced deformation and accretion, involving in a remnant ocean basin (Fig. 8). A tectonic melange (Northern-Calabrian Unit; Critelli, 1993, 1999; Mongelli et al., 2010) was formed in this time frame, including olistholiths and broken formations of oceanic sequences (both basement and its pelagic sedimentary cover) and crystalline rocks (gneiss and granite) (Spadea, 1982). The subduction of the Adria oceanic lithosphere beneath the European plate, producing along the southern-end of the European plate a continental-margin calcalkaline volcanic arc in Sardinia (e.g. Scandone, 1982;

Malinverno and Ryan, 1986; Channel and Mareschal, 1989; Dewey et al., 1989). Cretaceous to Eocene quartzose sandstone (Crete Nere Formation; Bonardi et al., 1988; Monte Soro Unit; Barbera et al., 2011), and late Paleogene (upper Eocene to upper Oligocene) quartzofeldspathic and volcanolithic sandstones are tectonically assembled within the tectonic *mélange* (Northern-Calabrian Unit) (Figs. 14). Quartzofeldspathic sandstone was derived from mixtures of ophiolitic detritus and neovolcanic detritus. Syneruptive volcanolithic sandstones, having basaltic and andesitic fragments, reflect climax of activity of the Sardinia volcanic arc during its initial arc volcanism (late Oligocene, 32-30 Ma; Critelli, 1993). The overlying sequence, the Saraceno Formation (that caps the Liguride Complex), is unconformably over the Northern-Calabrian Unit, is mixture of siliciclastic and carbonatoclastic strata, that are hybrid arenites, lithic and quartzolithic sandstones (Figs. 14). These sandstones reflect a provenance evolution from sedimentary-dominant (both carbonate and siliciclastic fragments) detritus to metamorphic and sedimentary mixtures. This provenance evolution testifies the initial signal of accretion and unroofing of the frontal thrust system of the northern Calabrian terranes (Sila Unit; e.g. Messina et al., 1994).

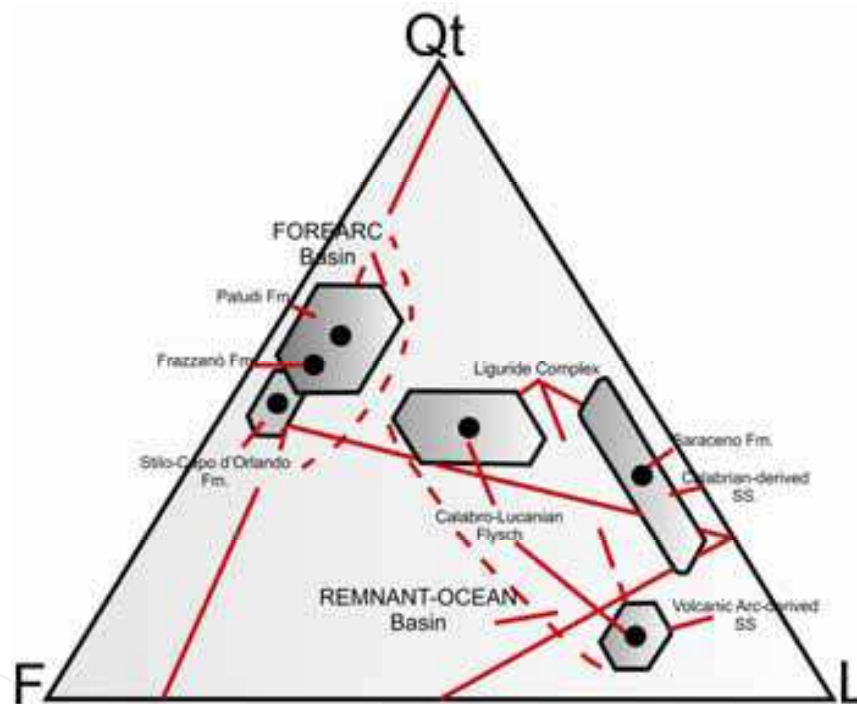


Fig. 14. QtFL plot (with superposed provenance fields of Dickinson, 1985) summarizing sandstone compositions from Paleogene to early Miocene sedimentary assemblages deposited during closure of the Liguride remnant ocean basin. Assemblages include the Calabro-Lucanian Flysch Unit (Tectonic Melange) and the Saraceno Formation of the Liguride Complex (remnant-ocean basin), and episutural (forearc basin) assemblages of the Paludi, Frazzan`o and Stilo-Capo d'Orlando formations. Data from Liguride Complex are from Critelli (1993); data from Paludi, Frazzan`o and Stilo-Capo d'Orlando Formations are from Zuffa and De Rosa (1978), Puglisi (1987), Cavazza (1989), Nigro and Puglisi (1993)

*Adria Margin.*- Siliciclastic sediments are rare or absent within basins located on Adria continental crust. The Paleogeographic scenario of this continental margin includes the outer carbonate domain (Apulia and Monte Alpi platforms), the Lagonegro basin, the inner carbonate platform, the Alburno-Cervati-Pollino units and the Sannio and Sicilide p.p.

basins (Fig. 9). These paleogeographic domains are characterized by thick (several thousands of meters in thickness) carbonate platforms, and their slopes, and pelagic basins. The pelagic basins receive dominantly Cretaceous to Upper Oligocene clays, marls and resedimented carbonate gravity flows (Patacca et al., 1992). There are no important traces of siliciclastic turbidite sandstones, the carbonate platforms record repeated emersions during the Cretaceous (Albian, and Albian-Cenomanian; e.g. Carannante et al., 1988a), showing pervasive karstification and deposition of bauxite deposits (locally up to 10 m). The bauxite deposits seem to be related to an intense weathering of original fine volcanic deposits (Carannante et al., 1988a).

During Paleogene, carbonate platforms and their slopes are characterized by thin stratigraphic sections, repeated emersions and non deposition (hiatus) intervals, that include the Eocene-Oligocene in the Alburno-Cervati-Pollino-Bulgheria units, all the Paleogene in the Monti della Maddalena unit and Apulia platform unit, and the Cretaceous to Early Miocene in the Monte Alpi Unit (Marsella et al., 1995).

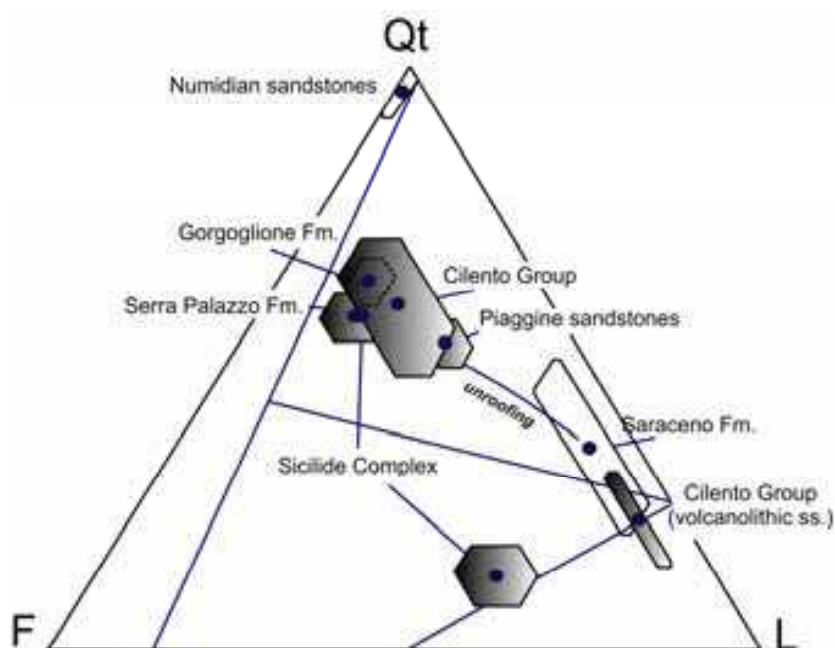


Fig. 15. QtFL plot (with superposed provenance fields of Dickinson, 1985) summarizing sandstone compositions from Burdigalian to Tortonian sedimentary assemblages deposited within the southern Apennines foreland basin system. Detrital modes of the Saraceno Formation shown as reference point of the onset of the unroofing history of the Calabrian terranes. The volcanic provenance from the volcanic arc is also recorded within the foreland basin sequences (Cilento Group and Sicilide Complex). Data from: Sicilide Complex is from Critelli et al. (1990, 1995b) and Fornelli and Piccarreta (1997); Cilento Group and Gorgoglione Formation is from Critelli and Le Pera (1994); Piaggine Sandstones and Serra Palazzo Formation is from Critelli and Le Pera (1995a)

Basinal sequences are dominantly pelagic, consisting of resedimented carbonate gravity flows (turbidite calcarenite to conglomerate debris flows and grain flows), representing carbonate slope aprons in the Lagonegro basin (Flysch Rosso Formation; Pescatore et al., 1988) and the Sicilide basin (Monte Sant'Arcangelo Formation; Lentini, 1979), interbedded with siliceous clays and shales. Rare siliciclastic or hybrid arenites are in the Paleocene

sections of the Sicilide Complex (Monte Sant'Arcangelo Formation; Selli, 1962; Lentini, 1979). The sandstone is quartzolithic, including abundant quartz and metamorphic and sedimentary lithic fragments. Eocene to Oligocene Colle Cappella Sandstone Formation (the lower portion of the Nocera Flysch; Ogniben, 1969; Zuppetta et al., 1984), is a turbidite system including abundant sandstones that are quartzolithic (Critelli and Le Pera, 1998). These sandstones have very abundant low-grade metamorphic fragments, suggesting initial erosion and accretion of the Calabrian terranes. However, the Colle Cappella Sandstone Formation could reasonable be younger than proposed ages, and be considered as early Miocene in age (Aquitainian to Burdigalian; Figs. 8,15) (Critelli et al., 1994, 1995b).

During early Miocene an abrupt paleogeographic and geodynamic change occur along the Adria margin (Figs. 8, 9). Transgressive shallow-water calcarenite sediments were deposited on carbonate platform domains (Fig. 16; Selli, 1957; Carannante et al., 1988b; Patacca et al., 1992; Sgrosso, 1998). Within the Sicilide and Lagonegro basins a thick (up to 1000 m) quartzose turbidite sand, the Numidian Sandstone Formation (Patacca et al., 1992) represent the key signal of a mature quartzose (cratonic) provenance from the northern Africa continental margin (Fig. 9; Wezel, 1970a, 1970b; Patacca et al., 1992). This widespread quartzose material was deposited, during upper Burdigalian (?) to Langhian, within the nascent foredeep of the Sannio-Sicilide, on the forebulge of the Alburno-Cervati-Pollino units, and on the back-bulge Lagonegro depozones (Patacca et al., 1992). Syneruptive andesitic volcanoclastic layers are interbedded with the shallow-water calcarenites (Fig. 16), and quartzose sandstones testifying the volcanic activity on the Sardinia Arc (Patacca et al., 1992), as such as arkosic debris flows (cf. Carbone et al., 1987) recording signals of provenance from accreted crustal block of the Calabrian terranes.

The active volcanic source and the crystalline sources of the Calabria-Peloritani Arc are recorded within the Sicilide foredeep, forming distinct early Miocene siliciclastic turbidite systems having sand compositions ranging from volcanolithic (Tufiti di Tusa Formation) to quartzolithic and quartzofeldspathic (Albanella, Corleto and Colle Cappella Formations) (Fig. 15; e.g. Critelli et al., 1994; Fornelli and Piccarreta, 1997; Critelli and Le Pera, 1998; Perri et al., 2011). Volcanoclastic detritus, interbedded with quartzose sandstone strata, seem to be also deposited within the back-bulge Lagonegro depozone (Pescatore et al., 1988).

### 3.2 Early collisional clastic units (Late Burdigalian to Tortonian)

Final closure of the Liguride remnant ocean basin and onset of continental collision in the southern Apennines are dated as early Miocene (Burdigalian). The provenance of the detrital constituents of the Miocene foreland sandstones was dominantly from the Calabrian Arc terranes, the active growing front of the fold-thrust belt (Fig. 8). Nevertheless, folded and thrustured remnant oceanic sequences, active volcanics, and the forebulge of the flexed Adria margin were in time and space important detrital sources of the southern Apennines foreland basin system (Fig. 8).

*The Source Areas of the Southern Apennines Foreland Basin System.* - The key sources of clastics deposited within the foreland basin system include different present-day realms (morphotectonic zones), that are:

- a. the Basement rocks of the northern Calabrian Terranes. Initial signals of the provenance is during final closure of the Liguride Complex (Saraceno Formation). During accretionary processes of, thrust units of the Calabrian terranes along the Adria margin (Figs. 11, 12).
- b. the uplifted subduction complex (the Calabro-Lucanian Flysch Unit), during the mid-late Miocene (Figs. 15,).

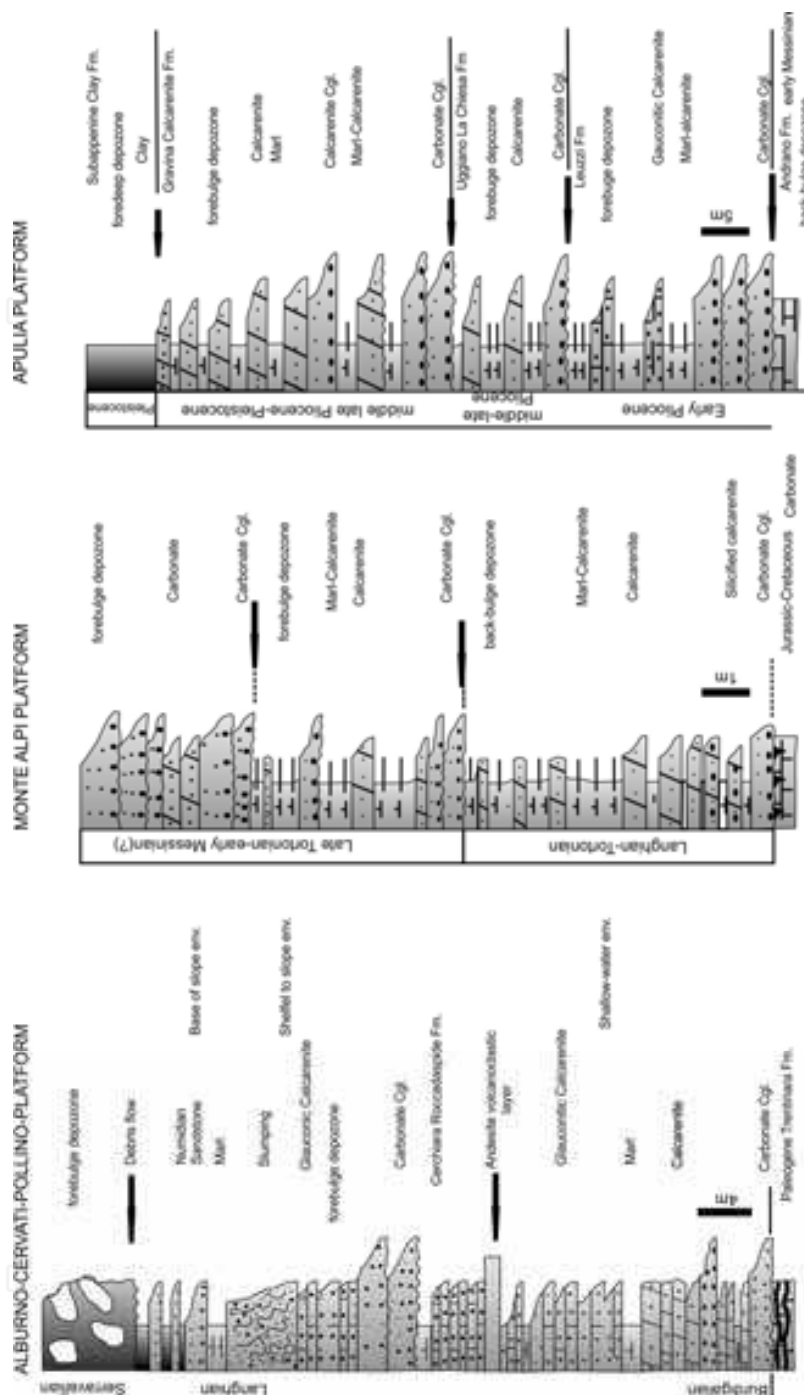


Fig. 16. Schematic columnar sections of the upper portions of the Adria carbonate platform domains, involved in flexural features during southern Apennines foreland basin system. The three sections correspond with the forebulge depozone sedimentation during (a) Burdigalian to early Tortonian (Alburno-Cervati-Pollino units or inner platform domain), (b) late Tortonian to Messinian (Monte Alpi unit), and (c) early Pliocene to the present (Apulia unit). Reference stratigraphic data for: the Alburno-Cervati-Pollino platform are from Carannante et al. (1988b), Critelli (1991), Patacca et al. (1992); the Monte Alpi Unit are from Sgrosso (1988), Taddei and Siano (1992), and personal unpublished data; the Apulia platform are from Ricchetti (1981), Ciaranfi et al. (1988), Ricchetti et al. (1988). Modified after Critelli (1999)



- c. the Mesozoic to Tertiary Apulia/Adria basinal and platform domains (Lagonegro, Sannio, Sicilide units, the Alburno-Cervati-Pollino units, Verbicaro-San Donato-Bulgheria-Monti della Maddalena units, Monte Alpi unit, and the Apulia unit). The forebulge sources to the foredeep depozones, were the Alburno-Cervati-Pollino-Monti della Maddalena units from Burdigalian to Tortonian; the Monte Alpi unit, from early Messinian to lower Pliocene, and the Apulia unit, since Pliocene (Fig. 16).
- d. An additional source of sediment is volcanic, that is mainly related to the calcalkaline volcanic arcs (between 32 and 11 Ma) was widespread along the western side of Sardinia (Cherchi and Montadert, 1982; Assorgia et al., 1986), or more recently to the intraorogenic alkaline volcanism in the Oligocene to Miocene.

### 3.3 Burdigalian-early Langhian Foreland basin system

The upper Sicilide Complex represents the oldest deposits of the foredeep basin (Critelli et al., 1995b). In the forebulge and back-bulge depozones, a widespread quartz arenite, the Numidian sandstone, as well as shallow-water calcarenite and thin volcanoclastic layers were deposited (Selli, 1957, 1962; Perrone, 1987; Carbone et al., 1987; Carannante et al., 1988b; Santo and Sgroso, 1988; Patacca et al., 1992; Sgroso, 1998) (Fig.16).

Siliciclastic strata of the Sicilide Complex (Fig. 8) include quartzolithic, volcanolithic and quartzofeldspathic sandstones. Thick (up to 300m) volcanolithic strata of the Tufiti di Tusa include syneruptive (e.g. Critelli and Ingersoll, 1995) andesite and basaltic andesite sandstones, recording climax of volcanic activity of the calcalkaline volcanic arc (Critelli et al., 1990; Fornelli and Piccarreta, 1997). Interbedded with carbonatoclastic sequences of the Adria forebulge (the Capaccio-Rocadaspide, Cerchiara Formations; Carannante et al., 1988a), similar andesitic volcanolithic sand testifies to wide dispersal of the neovolcanic detritus in the forebulge and back-bulge depozones (Pieri and Rapisardi, 1973; Perrone, 1987; Pescatore et al., 1988; Critelli, 1991; Patacca et al., 1992). Metamorphoclastic quartzolithic and quartzofeldspathic sandstones occur in the lower portions of the Tufiti di Tusa below the volcanoclastic strata (Critelli et al., 1990), and characterize the Corleto, Colle Cappella and Albanella formations (Fig. 15; Critelli et al., 1994; Fornelli and Piccarreta, 1997). They are derived from low to middle grade metasedimentary terranes, and are partly derived from ophiolitic rocks (Fornelli & Piccarreta, 1997). Interbedded thick carbonatoclastic (calcarenite-marl) strata within Tufiti di Tusa, Albanella and Corleto Formations, testify a provenance from the forebulge.

*Langhian to Tortonian Foreland Basin System.*- Since Langhian time, elongate turbidite basins have formed on top of advancing thrust-sheet systems (Fig. 9). The Cilento Group, Serra Palazzo, Piaggine, Gorgoglione, Sorrento, Castelvete, Oriolo, San Bartolomeo formations, Monte Sacro, Serra Manganile and Nocera Conglomerate formations (Figs. 3) are the main turbiditic successions that were deposited after the completion of rotation of Corsica and Sardinia (19±1 Ma; Dewey et al., 1989), in progressively shifting wedge-top and foredeep depozones of the growing foreland basin system. Except for the Serra Palazzo Formation and the Sorrento Sandstone, the base of each succession is everywhere marked by an unconformable contact (Patacca et al., 1990; Sgroso, 1998).

The Cilento Group (Fig. 17) ("*Cilento Flysch*" according to Ietto et al., 1965), Langhian to Tortonian in age (Amore et al., 1988; Russo et al., 1995; Zuppetta and Mazzoli, 1997), ranging from 2000 to 1200 m thick, rests unconformably on the Liguride Complex, and in turn it is unconformably overlain by the upper Tortonian Gorgoglione Formation, and the upper Tortonian to lower Messinian (?) Monte Sacro, Oriolo and Serra Manganile formations (Critelli et al., 1995b). The Cilento Group consists of different turbidite

depositional systems (Valente, 1993). In addition to siliciclastic turbidite beds, the Cilento Group includes numerous carbonatoclastic megabeds (ranging from few meters to 65 m thick; olistostrome beds (ranging from ten to hundreds of meters thick), and coarse volcanoclastic debris flows and turbidites. Sandstones (Fig. 15) of the Cilento Group are quartzolithic, volcanolithic and quartzofeldspathic (Critelli and Le Pera, 1994). Hybrid arenites and calcarenites characterize the carbonatoclastic megabeds. Sandstone strata of the lower portions are metamorphiclastic quartzolithic and quartzofeldspathic, resting on quartzolithic sandstone of the Liguride Complex.

Proportions of volcanic and plutonic detritus increase upward in the upper Pollica Formation and lower San Mauro, Torrente Bruca and Albidona formations. A volcanoclastic interval in the lower San Mauro Formation includes abundant felsic (rhyodacite to rhyolite) calcalkaline volcanic clasts (Critelli and Le Pera, 1994).

Sandstone of the upper Cilento Group is plutoniclastic quartzofeldspathic, consisting of abundant phanerites of plutonic and metamorphic fragments. In the upper Cilento Group thick carbonatoclastic and olistostroma megabeds record major tectonic events on both active thrust belt and forebulge (e.g. Critelli and Le Pera, 1994, 1998). Carbonatoclastic megabeds record huge volumes of sand-sized and mud derived from flexed Adria margin. These beds have impressive volumes and basinal lateral continuity (Colella and Zuffa, 1988; Cieszkowski et al., 1995). Olistostroma beds are siliciclastics, and include mountain-sized blocks of Calabrian terranes and Liguride Complex terranes (including also oceanic crust rocks; Ietto et al., 1965; Cocco and Pescatore, 1968; Carrara and Serva, 1982; Di Girolamo et al., 1992; Valente, 1991, 1993). Liguride-derived detritus appear only in the middle-upper Cilento Group, suggesting initial signals of the Liguride Complex emersion. Clear signals of the Liguride Complex emersion and erosion are recorded within the Piaggine Sandstone (Fig. 15; Serravallian to Tortonian; Sgrosso, 1981, 1998; Castellano et al., 1997). Quartzolithic sandstone of the Piaggine is derived from abundant Liguride Complex detritus (over than 50%;, suggesting that near the Serravallian-Tortonian boundary, the Liguride Complex was in a subaerial position, probably representing the emerged frontal thrust of the mountain belt (Critelli and Le Pera, 1995a, 1998).

The Serra Palazzo Formation (Selli, 1962; Ogniben, 1969) has been interpreted as the foredeep basin of the Langhian to Serravallian (Tortonian?) southern Apennines foreland region (Patacca et al., 1990). It has quartzofeldspathic sandstone (Fig. 15), hybrid arenite, and calcarenite, suggesting provenance from both thrust belt and forebulge. The middle-upper sedimentary succession, includes an olistostromal bed of carbonate clasts (olistoliths) (Loiacono and Sbarra, 1991) recording abrupt flexure along the passive margin.

Sandstone of the Gorgoglione Formation (Selli, 1962), upper Tortonian in age (Patacca et al., 1990), is quartzofeldspathic having similar provenance to that of the upper Cilento Group sandstone. The Cilento and Gorgoglione sandstone modes record accretionary processes of the Calabrian terrane, and initial unroofing of the crystalline terranes (Critelli and Le Pera, 1994, 1995a, 1998).

### 3.4 Late collisional clastic units (Late Tortonian to Messinian)

*Southern Apennines Foreland Basin-* The history of deep erosion of the Calabrian terranes is clearly recorded by Upper Tortonian to Messinian clastics (Fig. 10).

These foreland clastics, including Castelvetero, Monte Sacro, Oriolo, Serra Manganile, Nocara, and San Bartolomeo formations (Fig. 3), abruptly shift sand composition toward "ideal arkose" (e.g. Dickinson, 1985) or continental-block-derived sandstone (Fig. 18), suggesting deeply eroded Calabrian terranes. The previous forebulge of the Alburno-

Cervati-Pollino-Monti della Maddalena units, during the late Tortonian were assembled within the fold-thrust-belt, and the new forebulge of the foreland basin system might be located on the Monte Alpi Unit (Fig. 10; Patacca et al., 1992).

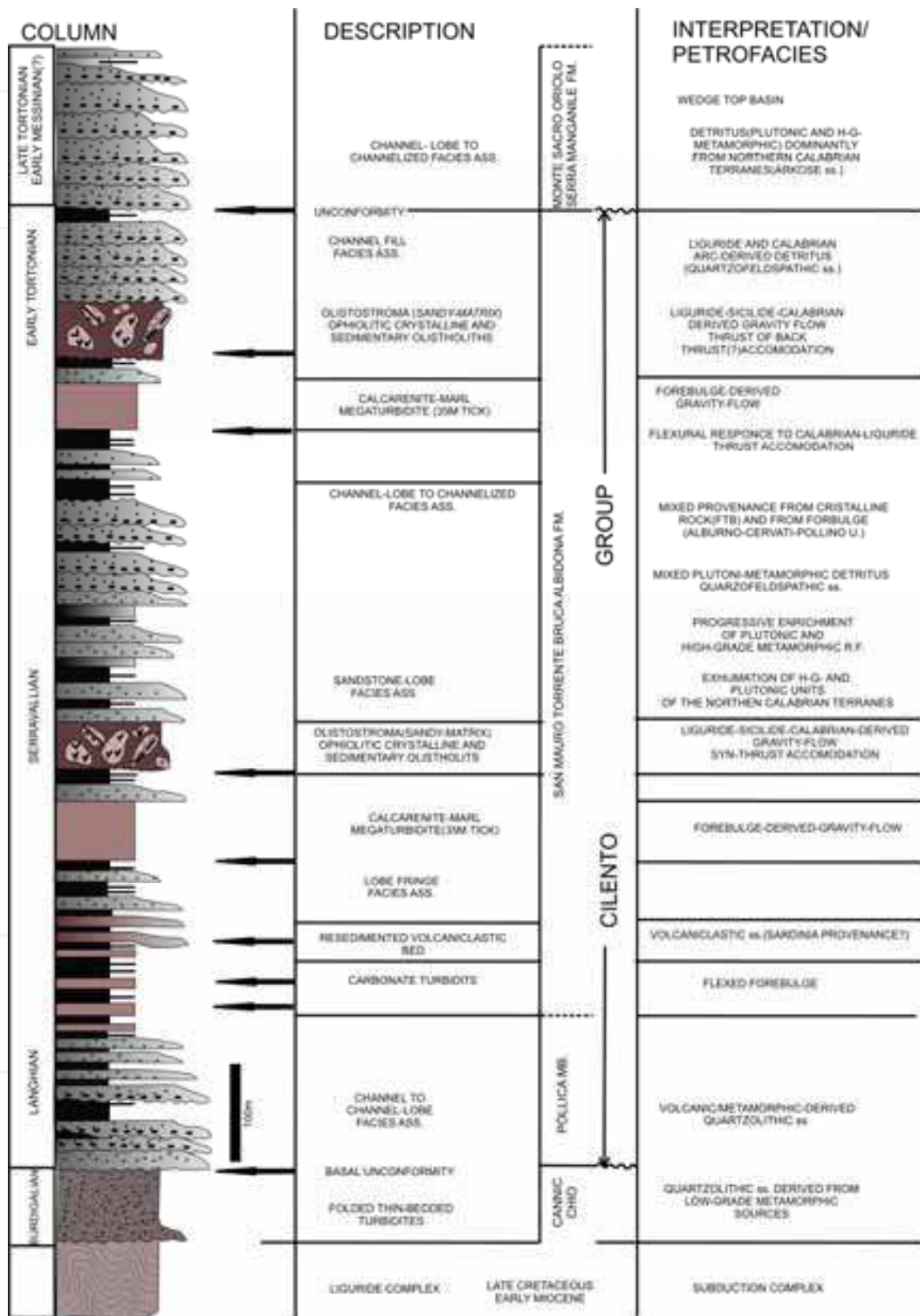


Fig. 17. Schematic columnar section of the Cilento Group. It rests unconformably on Liguride Complex and it is unconformably covered by late Tortonian-early Messinian clastic wedges (Monte Sacro, Serra Manganile and Oriolo Formations). Modified after Critelli (1999)

Forebulge sedimentation of the Monte Alpi Unit (cf. Sgroso, 1988b, 1998; Taddei and Siano, 1992) is thin (20m to about 100m in thickness) and consists of shallow-water to coastal arenite, marl and carbonate conglomerate (Fig.16). Arenite and rudite of the Monte Alpi are dominantly composed of carbonate detritus. Arenites of this forebulge sequence, are pure to impure calcilithite, composed of ancient extrabasinal carbonate grains (e.g. Zuffa, 1987) having Cretaceous to early Miocene tests, and the siliciclastic detritus includes rounded to subrounded quartz, plagioclase, radiolarian chert, fine grained quartz-siltite and quartzite, and rare serpentinite/ serpentine schist and volcanic lithic fragments. Rare quartzolithic sandstones, having abundant quartz, carbonate lithic grains and plagioclase, are interbedded with the calcilithite strata. Plutonic and metamorphic detritus is absent in these arenites.

On the Alburno-Cervati-Pollino units, locally, thin arkosic sandstone strata unconformably overlain the Miocene forebulge sequence or are directly on Cretaceous to Paleogene carbonates (cf. Patacca et al., 1992; Sgroso, 1998). These arkosic strata crop out on the Alburno-Cervati Mountains (Tempa del Prato Sandstone) and on the Pollino Mountains (Civita Sandstone) (Patacca et al., 1992; Sgroso, 1998; and include abundant plutonic and high-grade metamorphic detritus, as such as extrabasinal carbonate detritus.

Monte Sacro, Serra Manganile, Oriolo and Nocara Conglomerate formations (Figs. 5, 17) unconformably covering Liguride and Sicilide Complexes, and the Cilento Group, represent the wedge-top depozone sequences (Fig. 10, 17).

The Castelvetero Formation (Pescatore et al., 1970) has been interpreted as the foredeep basin (Patacca et al., 1990; Critelli and Le Pera, 1995b; Fig. 10). The Castelvetero has a thick olistostrome bed in the basal portions (Fig. 19), including mountain-block carbonate olistoliths (Pescatore et al., 1970; Pescatore, 1978; Carrara and Serva, 1982), that record involvement of the Langhian to Tortonian passive margin (e.g., the Alburno-Cervati Unit) within the thrust belt. Castelvetero sandstone modes are plutoni-metamorphiclastic, with up-section increases of sedimentary detritus (Critelli and Le Pera, 1995b). Sedimentary detritus is carbonate dominant in the lower Castelvetero; up-section increases of siliciclastic detritus suggests progressive erosion of older clastic wedges. Interbedded with quartzofeldspathic turbidite sandstone, the upper Castelvetero has a thick olistostrome bed composed of clastic detritus derived from Sicilide/ Sannio and Liguride complexes, and a 1m thick volcanoclastic layer (Fig. 19). The siliciclastic olistostroma may be the signal of the syn-thrust accommodation of the Sicilide/ Sannio Complex and possibly of the Liguride Complex. The syneruptive volcanoclastic layer consists of pyroclast fragments (pumice and shards) having felsic subalkaline composition (dacite) (Critelli and Le Pera, 1995b).

Messinian sandstones of the wedge-top and foredeep basins (Patacca et al., 1990, 1992; Sgroso, 1998), the Monte Sacro, Serra Manganile, Oriolo, Nocara, Tempa del Prato, Caiazzo, San Bartolomeo, Agnone formations have homogeneous quartzofeldspathic compositions similar to the Castelvetero sandstone (Fig. 18).

*Northeastern Calabria Foreland Basin-* The Tortonian to Messinian strata of the northeastern Calabria represent the more proximal portions of the southern Apennines foreland basin system. This sequence is directly unconformably over the Paleozoic plutonic and metamorphic rocks, or over the upper Oligocene to lower Miocene turbidite strata of the Paludi Formation (Fig. 19; Roda, 1967; Roveri et al., 1992), and it represents the basin fill of a wedge-top depozone (Critelli, 1999). These strata crop out along the piedmont of the Sila Massif, from the Trionto River (Rossano-Cariati zone) to south of the Neto River (Crotone zone) (cf. Cotecchia, 1963; Roda, 1964, 1967; Ogniben, 1973; Di Nocera et al., 1974; VanDijk, 1990; Van Dijk and Okkes, 1991; Roveri et al., 1992; Barone et al., 2008).

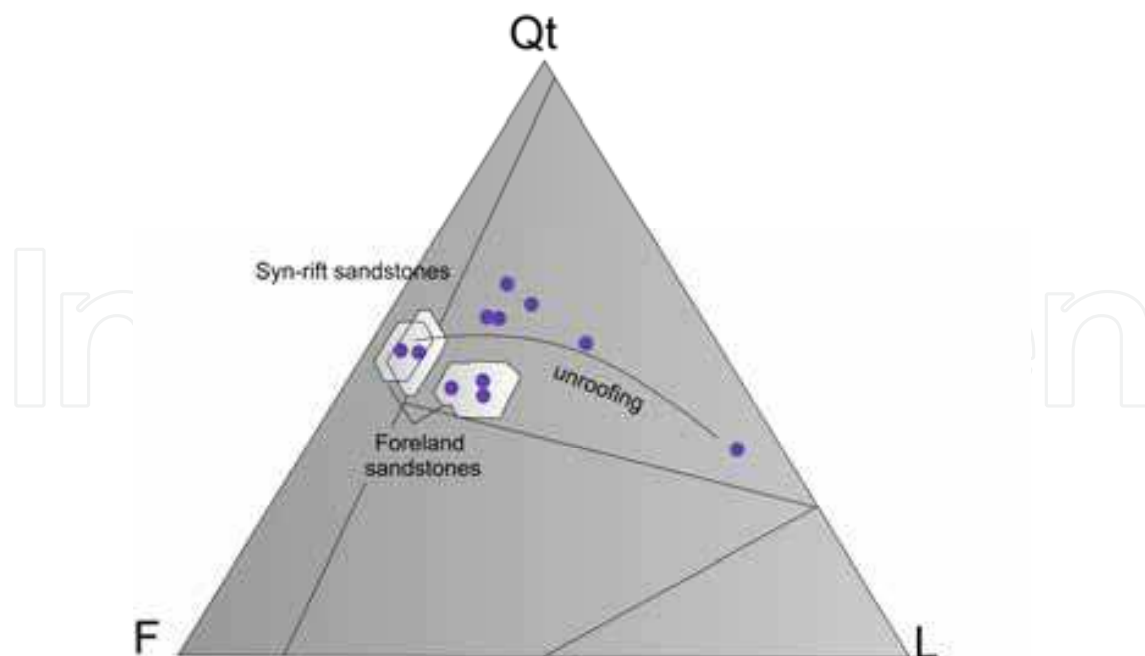


Fig. 18. QtFL plot (with superposed provenance fields of Dickinson, 1985) summarizing sandstone compositions from late Tortonian to early Messinian sedimentary assemblages deposited within the southern Apennines foreland basin system, and within the intermontane basins (northwestern Calabria) related to the backarc rifting of the Tyrrhenian Sea. Detrital modes of the Saraceno Formation, and Burdigalian to lower Tortonian foreland sandstones (only average values) are plotted as reference point of the unroofing history of the Calabrian terranes. Modified after Critelli and Le Pera (1998) and Critelli (1999)

Unconformably conglomerate and sandstone strata having rich macro-fauna (Clypeaster sandstone Formation, Cotecchia, 1963; or San Nicola dell'Alto Formation, Ogniben, 1955; Roda, 1964) represent the onset of the foreland basin system on advancing Calabrian thrust belt. These strata include diverse sedimentary facies associations, representing a depositional sequence (Roveri et al., 1992) and they are interpreted as a turbiditic system, having an overall fining and thinning upward trend, in the Croton basin (where it is over 1000m in thickness; Roveri et al., 1992).

It represents the main reservoir of dry gas (Roveri et al., 1992) In the other areas, these strata include also continental strata (alluvial fans), nearshore and shallow-water deposits (area between Bocchigliero and Campana). These strata are overlain by fine-grained turbiditic systems and, toward the thrust culminations of the Sila Massif, by shelfal deposits. These strata, correspond with the Ponda Formation (Roda, 1964) of the Croton basin, or the "Argilloso-marnosa Formation" of the Rossano basin (Ogniben, 1955) and may represent deposition during low-stand systems tract (Roveri et al., 1992).

The Rossano wedge-top depozone, during late Tortonian-early Messinian, abruptly receives huge volumes of Sicilide-derived olistostroma "Argille Scagliose Formation" (Ogniben, 1955, 1962) composed by variegated clay matrix and large blocks (olistholiths) of Cretaceous-Oligocene limestone, Miocene quartzolithic (similar to the Albanella-Colle Cappella sandstones) and quartzose sandstones (Numidian sands). These gravity flow deposits may be related to an out of sequence thrust accommodation or to a back-thrust of the Sicilide unit (Critelli, 1999). Infact, at the same time interval, within the foredeep depozone, the Castelvetero Formation has a similar olistostrome layer (Fig. 19); Critelli and Le Pera, 1995b).

On the successions of Rossano (north) and Crotona (south) Basins rests tectonically a sedimentary allochthonous succession defined named "Cariati Nappe" (Roda, 1967a; Ogniben, 1973). This succession is made up of turbiditic bodies with thinning-upward trend of Langhian-Serravallian in age, involved in backthrusting starting late Tortonian and involving the evaporitic and post evaporitic units in the Rossano Basin (Barone et al., 2008). The CN includes a Middle to Upper Miocene clastic succession unconformably covering an Oligocene to Burdigalian siliciclastic flysch. The Miocene and post Messinian emplacement of the so-called "Cariati Nappe" (CN) in the central sector of the area interrupts the lateral continuity and affects the sedimentary supply of a such configured wedge-top basin.

The Messinian sequence is characterized by evaporite deposits which record the Mediterranean salinity crisis. The evaporites consist mainly of gypsum and halite, followed by a thin mudstone interval, and thin clastic and evaporite beds (Ogniben, 1955; Roda, 1964; Romeo, 1967; Di Nocera et al., 1974). Overlying the evaporite sequence, an erosional unconformity marks the base of a Late Messinian to Pliocene depositional sequence within the Crotona Basin (Roveri et al., 1992). This depositional sequence consists of a basal conglomerate and sandstone strata with fining-upward trend (transgressive systems tract; Carvane Conglomerate Formation; Roda, 1964), overlain by basin-wide marine shales (highstand systems tract; Marne argillose dei Cavalieri Formation; Roda, 1964) (Roveri et al., 1992).

The juxtaposition of autochthonous basinal successions (Rossano and Crotona successions) and allochthonous (Cariati Nappe) would suggest the detection, during the Serravallian-Tortonian, of the sedimentary basins developed in different context; A basin on the inner set of the Arco Calabro Units which the western edge is well outcropping, and an outer external basin set on Sicilide units and Albidona formation.

Therefore, the Cariati Nappe would give the meaning of a backthrusts of Tortonian age, related to the upper-middle Miocene accretionary phases that sharing the Foreland Basin system of the intersection of southern Apennines-Calabrian terrane.

Because of its sedimentary succession, the Cariati Nappe would include many tectonostratigraphic similarities with the sedimentary successions of the Upper Ionian Calabria and Lucania, which identify the area of the Montegiordano-Nocera-Rocca Imperiale ridge (Zuppetta et alii, 1984; Mostardini & Merlini, 1986; Patacca & Scandone, 1987, 2001; Carbone & Lentini, 1990; Cinque et alii, 1993; Critelli, 1999) where the successions of the Albidona and the high portion of the Sicilidi Units, posed by the Argille Scagliose formation and Colle Cappella Sandstones, rests conglomeratic and arenaceous turbiditic successions belonging Serravallian-Tortonian of Oriolo Formation and Nocera Conglomerates Formation.

Synchronously with major tectonic events in the foreland thrust-belt, extensional tectonic activity affected the Tyrrhenian margin, just after the Tortonian compressive event; thereafter, evolution of the Tyrrhenian basin strongly influenced peripheral deformation of the Apennines foreland region (e.g., Malinverno and Ryan, 1986; Royden et al., 1987; Kastens et al., 1988; Lavecchia, 1988; Patacca et al., 1990; Sartori, 1990).

Since late Tortonian, the Calabrian terranes have provided abundant detritus to both the foreland region and intermontane basins of the backarc region (Figs. 3, 10) characterized by similar detrital provenances. Sandstones are "ideal arkose", and are identical in composition with the distal deep-marine upper Tortonian to Messinian foreland strata (e.g., Critelli and Le Pera, 1995a; Critelli et al., 1995b) (Fig. 18).

Marginal syn-rift strata of the Coastal Range of western Calabria (Amantea Basin) has Upper Tortonian to Messinian arkose, hybrid arenites and calcarenite, however, are similar to sandstone strata of the Crotona Basin (peri-ionic area) (Critelli, 1999; Barone et al., 2008).

## CASTELVETERE FORMATION

COLUMN	DESCRIPTION	INTERPRETATION
Lower Messinian Upper	tephra layer volcanolithic ss.	syneruptive dacite tephra
	Olistostroma (Clay matrix) similar to b) erosional surface	Sicilide Sannio-derived gravity flows Back-thrust(?) accommodation
Tortonian-Messinian Middle	Channel-lobe turbidite facies quartzofeldspathic ss.	progressive enrichment of sedimentary-derived detritus mixed carbonate's siliciclastic detritus
	Olistostroma (Clay matrix) Olistoliths: 2-sandstone (quartzarenite, quartzofeldspathic/quartzolithic) 3-Marl/Clay Strata 4-Radiolarite, Argillaceous Chert Strata 5-Silicified Carbonate and Marl erosional surface	Sicilide/Sannio-derived gravity flows Syn-thrust accommodation
Upper Tortonian Lower	quartzolithic ss. Channel-lobe turbidite facies quartzofeldspathic ss.	input of ophiolitic-derived detritus Carbonatoclastic/Siliciclastic Apron
	Channel-fill turbidite facies quartzofeldspathic ss.	Mixed Crystalline/carbonate Detritus
Mesozoic	Olistostroma (Sandy matrix) Olistolith: 1.-Mesozoic Carbonate erosional surface	Platform-derived gravity flows Inner Platform Unit involved within FTB
	basal unconformity kartic surface Carbonate basement	

Fig. 19. Schematic columnar section of the Castelvetero Formation. It rests unconformably on Mesozoic carbonate platform unit. Reconstructed stratigraphy is from personal data and from Pescatore et al. (1970), Cocco et al. (1974), and Sgrosso (1998). Modified from Critelli (1999)

### 3.5 Pliocene to Quaternary clastic units

*Southern Apennines.*- During Pliocene, loading of the lithosphere by eastward thrusting of the Southern Apennines thrust-belt over the Adriatic plate resulted in flexural warping of the Apulia platform forming a downwarp, the Bradanic foredeep, an upwarp at the Apulia western edge, the Murge-Salento forebulge (Figs. 6, 16). The Bradanic foredeep basin was

initially filled by early Pliocene pelagic and turbiditic sediments, and from middle-upper Pliocene to Pleistocene by hemipelagic clays (Argille Subappennine; Fig. 16), shallow-water calcarenite (Gravina Calcarenite Formation; Fig. 14), deltaic, coastal and alluvial clastics (e.g. Casnedi et al., 1982; Casnedi, 1988; Pieri et al., 1996). Clastics of the bradanic foredeep depozones reflect the erosion of the previous accreted clastic and carbonate thrust units of the southern Apennines fold-thrust-belt. The Sant'Arcangelo Basin was one of the wedge-top depozone of the Pliocene-Pleistocene foreland basin system (e.g. Hyppolite et al., 1994a, 1994b). Sediments of this basin show blended clasts from sedimentary thrust units, the uplifted subduction complex (Liguride Complex) and also from Calabrian arc. Most of the typical northern Calabrian provenances were trapped in piedmont areas of the Crotona and Rossano Basin, or within the Crati Basin. Because of the strike-slip movements were also active during Pliocene and Pleistocene in the southern Apennines realm, diverse fault controlled small basins were developed (e.g. Turco et al., 1990, Van Dijk et al., 2000, Tansi et al., 2007). Estimated uplift rates are almost equal to late Quaternary denudation rates. Tectonics and climate have had a strong effect on the landforms of the Calabrian mountain ranges, resulting in the higher accumulation rates.

Additional Quaternary sediment sources for the Paola and Corigliano basins are active volcanic centres bordering the Paola Basin, and submarine structural highs, such as Amendolara embankment (Romagnoli and Gabbianelli, 1990), bordering the Corigliano Basin, producing reworking intrabasinal detritus. Quaternary sedimentation of both basins is strongly influenced by glacio-eustatic changes (Chiocci, 1994; Trincardi et al., 1995).

The Corigliano trough represents the Holocene submarine wedge-top depozone of the southern Apennines and northern Calabria foreland region (Pescatore and Senatore, 1986) (Fig. 7). It is morphologically characterized by a restricted shelf area, numerous gullies and canyons, and a submarine fan, the Crati Fan, developed during Holocene and connected with the torrential-type Crati delta on the shelf (Ricci Lucchi et al., 1984; Romagnoli and Gabbianelli, 1990).

The Crati River drains both the Calabrian crustal block to the west, east and south, and the southern Apennines Mesozoic to Tertiary sedimentary terranes to the north.

The Tyrrhenian margin of northern Calabria consists of diverse small coastal drainages, draining both Calabria continental block and the southern Apennines thrust belt, supplying sediments to the deep-marine Paola Basin. The basement of the basin consists of crystalline rocks of the Calabrian terranes or the upper Tortonian to Messinian sedimentary sequences. The basal unconformity is early Pliocene in age, and sediments of this age are bathyal (Fabbri et al., 1981). The main Pliocene to early Pleistocene unconformities seem to be related to the abrupt uplift of the Calabrian Coastal Range ("Catena Costiera"; e.g. Ortolani, 1978; Fabbri et al., 1981; Barone et al., 1982; Wezel, 1985).

Modern beach and fluvial sands of the Tyrrhenian margin of northern Calabria have three distinct petrofacies from north to the south, namely, (1) the calcilithic Lao petrofacies (at the northern end of the Paola Basin drainage area), having a provenance from the south-western flank of the southern Apennines slope, including dominantly Mesozoic carbonate rocks, (2) the quartzolithic Coastal Ranges petrofacies (in the central portion of the Paola Basin drainage area), having a provenance from dominantly metamorphic terranes (dominantly phyllite and schist, and gneiss) of the Coastal Ranges, and (3) the quartzofeldspathic Santa Eufemia Gulf petrofacies (at the southern end of the Paola Basin drainage area), having a provenance from metamorphic (dominantly gneiss, and phyllite and schist) and plutonic terranes of the Sila and Serre Mountains and from sedimentary terranes of the Catanzaro



Graben. They represent the actualistic petrofacies of the mainland areas of the deep-marine Paola Basin (e.g. Le Pera and Critelli, 1997; Critelli and Le Pera, 2003).

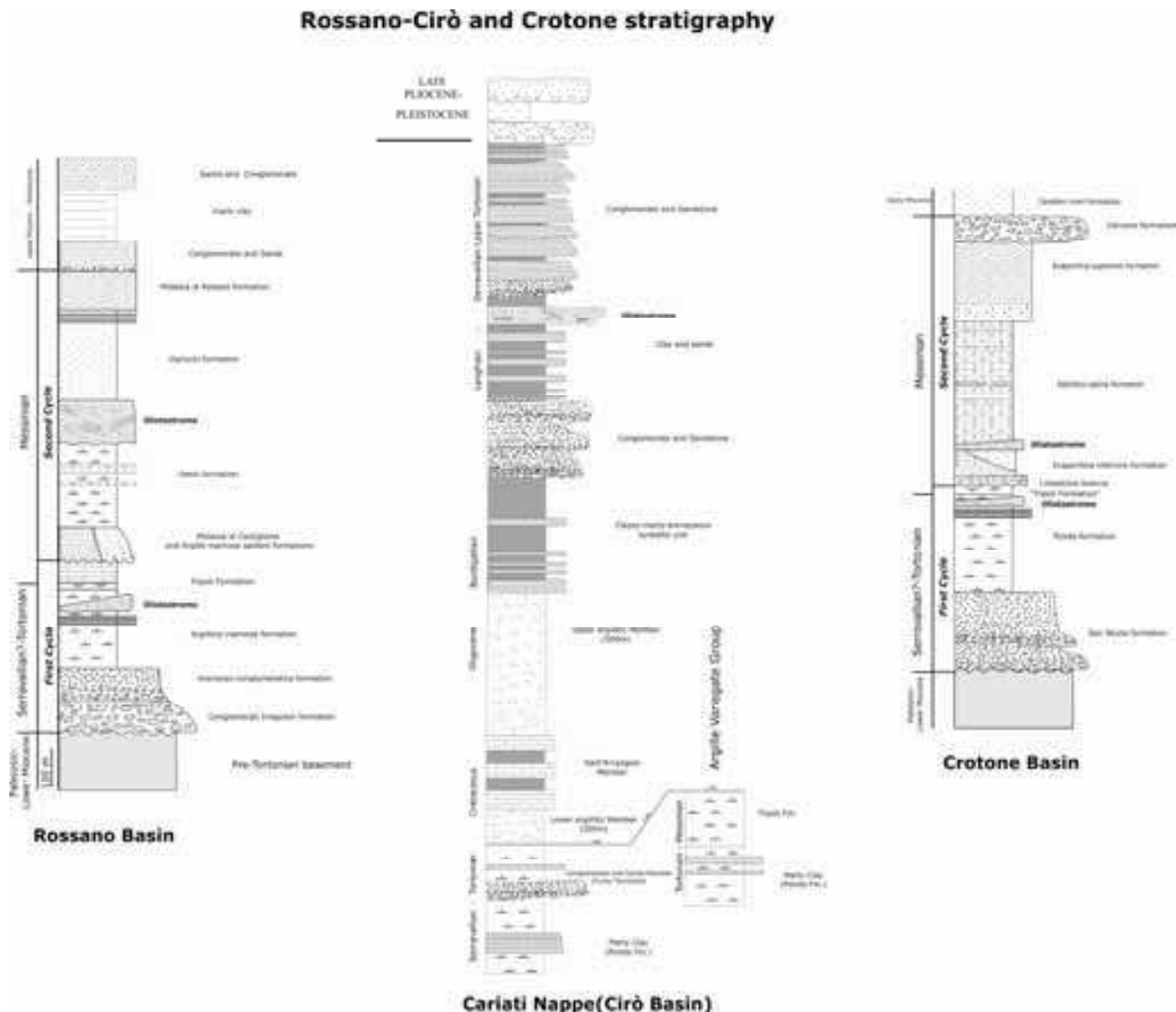


Fig. 20. Schematic columnar sections of the Crotona-Cirò and Rossano basins. The sedimentary successions rest unconformably on Paleozoic plutonic and metamorphic rocks of the northern Calabrian Arc (Sila Unit), or on Oligocene to lower Miocene Paludi Formation. These sequences represent the more proximal late Tortonian to Pliocene strata (wedge-top depozone) of the southern Apennines foreland basin system. Reconstructed stratigraphy is from personal data and from Ogniben (1962), Roda (1964, 1967), Romeo (1967), and Roveri *et al.* (1992). Modified after Critelli (1999)

Late Quaternary turbidite sands of the Paola Basin have distinct petrofacies (Fig. 21), that are: (a) a quartzolithic petrofacies, including also calcilitic turbidite sands, and (b) a volcanic-rich petrofacies, including distinctive syneruptive volcanoclastic sands (Critelli, 1999).

The quartzolithic sand petrofacies widely occurs into the Paola Basin and it is strictly related to the composition of the Coastal Range littoral province. At the northern end of the Paola Basin, distinct sedimentoclastic (calcilitic) turbidite sands, reflects a provenance from the Lao littoral province (Le Pera and Critelli, 1997; Le Pera, 1998; Critelli and Le Pera, 2003).

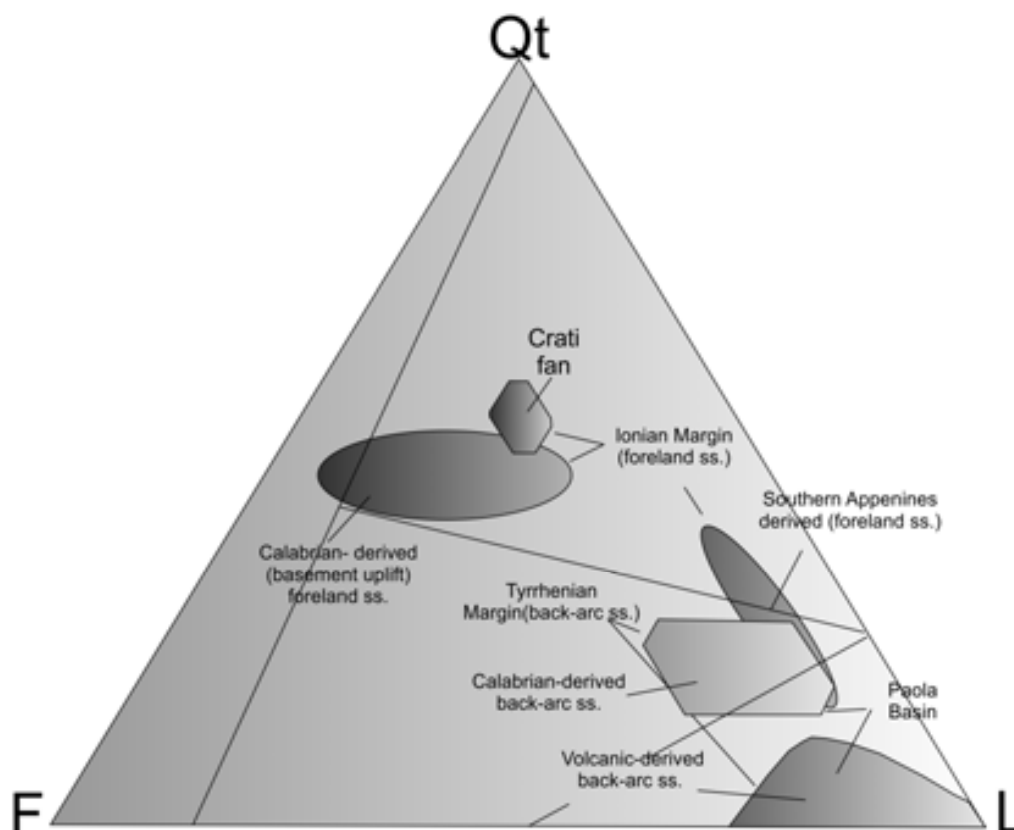


Fig. 21. QtFL plot (with superposed provenance fields of Dickinson, 1985) summarizing sand compositions for upper Pleistocene to modern sands of the Crati Fan, the Paola Basin (turbidite sands), and beach and fluvial sands of both Ionian and Tyrrhenian margins of northern Calabria. Sands from Ionian margin include the Crati Fan turbidites, the Crati River and Delta, the drainages derived from the Sila Massif and from the southern Apennines thrust belt. Sands from Tyrrhenian margin include Paola Basin turbidites, tephra layers interbedded with turbidites, coastal and fluvial systems of the Coastal Range. Data from Le Pera (1998) and Critelli and Le Pera (2003). Modified after Critelli (1999)

The volcanic-rich sand petrofacies also well represented within the Late Quaternary stratigraphic sequence of the Paola Basin. This petrofacies includes two main syneruptive volcanoclastic turbidites, one is located close to a datum plane at 20,000 y (calcalkaline volcanic provenance), and the upper one (alkaline volcanic provenance) is at the top of the basin-fill.

### 3.5 The climax of accretionary processes and evolution of Foreland basin system

The unroofing history of the Calabrian terranes, started during final closure of the Liguride basin, abruptly increased during accretionary processes over the Adria margin, occurred during early Miocene (Fig. 8). Increasing detrital feldspars and metamorphic detritus in the early Miocene sandstones (lower Cilento Group; upper Sicilide Complex) suggest dissection of the frontal terranes of the northern Calabrian arc. Local huge arrivals of volcanoclastic detritus testifies the climax of activity of the calcalkaline volcanic arc of Sardinia (Fig. 8). In middle Miocene (Serravallian to Tortonian) sandstone detrital modes recorded a major change from lower Cilento Group to upper Cilento Group and Gorgoglione Formation (Fig.

14), with marked increase in detrital feldspars, medium- to high-grade metamorphic and plutonic rock fragments (Critelli and Le Pera, 1994). This compositional change, related to rapid northeastern movement of the Calabrian terranes and thrust accommodation of the high-grade Hercynian metamorphic rocks (Fig. 12), reflects rapid rise of Calabria and sharp increase in denudation rates, as documented also by fission-tracks (e.g. Thomson, 1998).

At this stage, the Adria (Alburno-Cervati-Pollino-Monti della Maddalena units) forebulge was involved in tectonic deformation and assembled within the orogenic belt (Fig. 10); the Liguride subduction complex, that was part of the deep duplex system, locally emerges producing abundant detritus to the foreland (Piaggine Sandstone and Olistostroma beds of the upper Cilento Group; Critelli and Le Pera, 1995a, 1998) (Figs. 15, 17).

In upper Miocene (late Tortonian to Messinian) sandstone detrital modes recorded an other major change (increasing feldspars, high-grade metamorphic and plutonic fragments), and the composition shift toward "ideal arkose" (Fig. 17; Critelli and Le Pera, 1995a). This time is also marked by the onset of the Tyrrhenian rifting on the back of the orogenic belt, causing an increased eastern displacement of the thrust system (e.g., Cello et al., 1981, 1989; Carbone and Lentini, 1990; Patacca et al., 1990; Lentini et al., 1994; Sgrosso, 1998). This other compositional change of the foreland sandstones reflects an increasing of the uplift rates, and deep erosion levels into mid-crustal rocks along the core of the Calabrian thrust belt.

### 3.6 Discontinuous migration of flexural features

The syntectonic Miocene stratigraphic succession indicates episodic, eastern migration of the forebulge. The position of the flexural forebulge did not progress continuously eastward through time but appeared to stall at its initial position from Langhian to Tortonian during deposition of the Cilento Group, Gorgoglione Formation and "Cariati Nappe" succession (Fig. 9). Only during the upper Tortonian to Messinian the forebulge moved rapidly eastward during deposition of Castelvete and San Bartolomeo formations (Fig. 10). Possible reasons for discontinuous migration of flexural features that may apply to the southern Apennines foreland are episodic migration of the thrust load and inhomogeneities within the lithosphere (e.g. Boyer and Elliot, 1982; Waschbusch and Royden, 1992; Giles and Dickinson, 1995; Critelli, 1999).

The buildup and migration of an accretionary prism includes progressive cratonward outstepping of the thrust front incorporating new material into the thrust load as it migrates (e.g. Giles and Dickinson, 1995). Thrust loads may also build up by almost vertical stacking of thrust sheets along ramps within the hinterland, producing the critical taper needed for the thrust system to migrate (e.g. Boyer and Elliot, 1982). In this case, flexural features would not continuously migrate cratonward because the thrust load itself is not continuously migrating cratonward. Waschbusch and Royden (1992) suggest that discontinuous migration of flexural features may also be caused by inhomogeneities within the lithosphere that fix the position of the forebulge to a weak segment of the foreland lithosphere.

Flexed features of the southern Apennines foreland basin system change discontinuously during the last 23 my. The forebulge did not migrate from early Miocene to the Tortonian and it may have been fixed to a weak zone corresponding to the former miogeoclinal shelf margin of the Alburno-Cervati-Pollino inner platform domain (Figs. 9,16; Patacca et al., 1992). The Calabrian allochthon terranes and associated subduction zone progressively migrated toward the fixed forebulge until stresses reached a threshold during late

Tortonian-early Messinian (Fig. 10), the time of deposition of the Castelvetero Formation (Critelli and Le Pera, 1995b).

At this time the forebulge migrated rapidly eastward causing deformation of the former back-bulge basin strata (Lagonegro basin). The inferred new forebulge may be the inner Apulia platform unit (or Monte Alpi Unit; Figs. 10, 16). The Monte Alpi Unit has an unconformable lower Messinian carbonatoclastic sequence (Figs. 16), that is transitional to shallow-water (cf. Sgrosso, 1988a, 1988b, 1998; Taddei and Siano, 1992), representing remnants of deposition on the forebulge. The last eastward forebulge migration occurred during Pliocene, the new forebulge is the Apulia platform (Fig. 16), and the foredeep is the Bradanic trough (Fig. 6; e.g. Ricchetti, 1980; Ricchetti and Mongelli, 1980; Casnedi et al., 1982; Critelli, 1999).

#### 4. Conclusions

In foreland settings, subsidence and uplift are profoundly affected by lithospheric flexure. Foreland basin subsidence is primarily controlled by downflexing of the lithosphere in response to thrust accommodation and loading. The interrelationships between lithospheric flexure, single thrust accommodation within the accretionary wedge and flexural subsidence experiences geometrically complex entities within the foreland region (e.g. Critelli, 1999).

This chapter has examined clastic sediments and interpreted many stratigraphic sequences that were deposited in the southern Apennines foreland basin system during the complex orogenic history of the western Mediterranean, suggesting that interplay of lithospheric flexure and thrust accommodation were important factors in controlling accommodation trends. From late Paleogene to the present the siliciclastic sedimentary sequences of southern Italy filled basins that are directly related to this convergent setting, causing consumption of the oceanic lithosphere, and subsequent accretion of the Calabrian allochthonous terranes over the Adria-Africa plate generating post-Oligocene foreland basin systems.

Earliest onset of continental accretion on Adria margin occurred during late Burdigalian-early Langhian; foreland clastic strata of the upper Sicilide Complex were derived from dominantly metasedimentary and related sedimentary covers rocks of the frontal Calabrian terranes (e.g. Critelli, 1999). Sudden influx of neovolcanic detritus suggests continuing provenance from the active volcanic arc that was possibly hundred kilometers distant (Fig. 8).

Thick foreland clastic sequences formed during Langhian (Fig. 9) over accreted Liguride and Sicilide Complexes.

Langhian to Serravallian detrital modes of the Cilento Group (Fig. 17) have abrupt changes from quartzolithic (phyllite and schist source rocks) sandstone to quartzofeldspathic (plutonic and gneissic source rocks) sandstone. This change in detrital modes was accompanied by interbedded carbonatoclastic detritus derived from abrupt flexure of the Alburno-Cervati-Pollino forebulge (Figs. 17). These major changes in foreland clastic deposition occurred *c.* 16-15 Ma (Critelli and Le Pera, 1994), the time of exhumation of the northern Calabria crystalline rocks (Thomson, 1994, 1998). Signals of sandstone detrital modes and fission tracks on source rocks indicate increasing uplift between 15 and 10 Ma, after which deep dissection affected the northern Calabrian Arc.

During Serravallian to lower Tortonian (15 to about 10 Ma), abundant ophiolitic and pelagic detritus within the foreland basin (upper Cilento Group and Piaggine Sandstone) record emersion and erosion of the Liguride and Sicilide complexes. At *c.* 11 Ma, volcanism of western Sardinia abruptly terminated, rendering Sardinia into a remnant arc.

During Late Tortonian to Early Messinian, abrupt changes in the southern Apennines foreland region occurred. The former forebulge (e.g., Alburno-Cervati-Pollino units) was involved in tectonic deformation and incorporated within the thrust belt, and an eastward shifting of the foredeep depozone occurred. Foredeep and wedge-top sandstone strata shifted in composition toward continental block or arkose, suggesting major uplift of the Calabrian thrust belt. Coeval thrust accommodation of Sicilide-Sannio units, recorded as large gravitative deposits within foredeep and wedge-top depozones (Figs. 19, 20) and recycling of older sedimentary sequences are important signals of accretionary processes within the thrust-belt. Signals of post-8 Ma volcanism are recorded in the foreland sequences, but sources are unknown.

Since Late Tortonian (10-8 Ma; Fig. 10), backarc rifting has produced the Tyrrhenian Sea. At this time, the northern Calabrian Arc has been the western border of the northern Ionian foreland region and the eastern margin of the Tyrrhenian backarc basin (Fig. 10). Intermontane and syn-rift basins of the western Calabria and proximal and distal foreland basins of eastern Calabria and southern Apennines have identical sand composition, plotting within ideal arkose or continental block provenance field (e.g., Dickinson, 1985, 1988) (Fig. 18). The maximum rate of foreland thrust advancement ( $8 \text{ cm yr}^{-1}$ ) occurred in late Tortonian to Messinian (Patacca et al., 1990); major changes in uplift rate in the northern Calabrian Arc correspond with the abrupt change in sandstone detrital modes (Critelli and Le Pera, 1995a; Critelli, 1999; Barone et al. 2006).

Upper Tortonian to Messinian nonmarine to shallow-marine and deep-marine successions, cropping out on the western and eastern Calabrian Arc, representing synrift clastic wedges related to backarc rifting in the peri-Tyrrhenian area (western sequences), or foreland clastic wedges in the peri-Ionian area (eastern sequences; Barone et al., 2008).

Pliocene and Quaternary of the northern Calabrian Arc, represented by foredeep and related wedge-top basins on the eastern side (Gulf of Taranto and Corigliano Basin), and a slope basin on the western side, the eastern Tyrrhenian margin (Paola Basin). These receive detritus primarily from deep erosion of northern Calabria. The modern deep-marine basins of offshore northern Calabria have many similarities to the middle to upper Miocene clastic sequences in both foreland and backarc regions of the southern Italy. The type of sedimentary provenance of the southern Italy foreland basin system, providing an example of the changing nature of the orogenic belt through time, may contribute to general understanding and application of other major orogens.

Quaternary erosional rates for some areas of Calabria are over  $200 \text{ mm/ Ka}$  with maximum values exceeding  $800 \text{ mm/ Ka}$  (Ibbeken and Schleyer, 1991). The enormous sediment production of Calabria crosses the river-mouth areas and the beaches, and is spread over shelf and slope or is transferred via submarine canyons to lower bathyal plains of the Ionian seas. The Corigliano Basin, in the Ionian Sea, record the sedimentary processes acting on the northern Calabrian Arc terranes and can be considered as a modern analogue of / thrust-belt/ foreland transect. During Quaternary, in the southern Apennines thrust-belt, other major thrust accommodations, as such as normal faulting, sinistral strike-slip movement, block rotation and strong uplift occur defining the morphotectonic zones of the orogenic belt. The northern Ionian Sea and the Bradanic River basin represent the submarine and subaerial foredeep depozone, respectively. The Corigliano-Amendolara submarine basins, and the Sant'Arcangelo basin represent the wedge-top depozones,. The forebulge is finally located to the western Apulia platform, while the back-bulge depozone is located to the southern Adriatic Sea (Critelli, 1999).

The flexed foreland lithosphere and its forebulge has played an important role to the development of the foreland basin system. The initially formed forebulge is interpreted to have been in the Alburno-Cervati-Pollino domain, and the final position of the forebulge is the present Apulia platform domain (Figs. 6, 7, 16). Thus the forebulge migration distance was over 150 km. Forebulges contributed to the sediment supply within the foreland basin, even if it is minor with respect to the orogenic provenance. Forebulges have produced instantaneous huge volumes of single carbonatoclastic megabeds testifying major forebulge instability (Critelli, 1999).

An estimation of detrital supply from Calabria during the last 25 my suggests that at least 5 to 8 km of crust has been removed from the Calabrian orogenic belt and deposited in the marine basins (Critelli and Le Pera, 1998; Critelli, 1999). Erosional budgets and accumulation rates document the immense volume of detrital sediments transferred from deeply weathered crystalline rocks of the Calabrian Arc to marine basins.

The clastic compositions of the southern Apennines foreland basin strata reflect the changing nature of the thrust belt through time, recording the history of accretion of the Calabria microplate over the Adria margin. The type of sedimentary provenance analysis, providing an example of the close relations between clastic compositions and growing orogen in southern Italy orogenic system, may contribute and have general application to other major orogens.

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Ocean closure involves a variety of converging tectonic processes that reshape shrinking basins, their adjacent margins and the entire earth underneath. Following continental breakup, margin formation and sediment accumulation, tectonics normally relaxes and the margins become passive for millions of years. However, when final convergence is at the gate, the passive days of any ocean and its margins are over or soon will be. The fate of the Mediterranean and Persian Gulf is seemingly known beforehand, as they are nestled in the midst of Africa-Arabia plate convergence with Eurasia. Over millions of years through the Cenozoic era they progressively shriveled, leaving only a glimpse of the Tethys Ocean. Eventually, the basins will adhere to the Alpine-Himalaya orogen and dissipate. This book focuses on a unique stage in the ocean closure process, when significant convergence already induced major deformations, yet the inter-plate basins and margins still record the geological history.

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