IGCP Project 639
“Sea Level Change from Minutes to Millennia”

Crossing Southern Italy: a travelling meeting from Taranto to Siracusa
IGCP Project 639 - Sea Level Change from Minutes to Millennia
Crossing Southern Italy…a travelling meeting from Taranto to Siracusa

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First and Fourth of Cover: Raised wave cut platform, Aci Castello, Catania, Sicilia, Italy (Photo by Giuseppe Mastronuzzi)

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Preface

Crossing Southern Italy 2018 is the annual meeting of IGCP 639 Project: “Sea Level Change from Minutes to Millennia”. It is the occasion to summarize the results produced by previous IGCP 61, IGCP 200, IGCP 274, IGCP 437, IGCP 495 and 588.

The field guide comprises results of about 40 years of research carried out along the coast of Puglia, Basilicata, Calabria and Sicilia regions by researchers belonging to different universities and nationalities.

Most of researchers have been supported by Italian Ministry of Education, Universities and Research (MIUR) and managed by Italian Universities from Southern Italy (i.e.: Bari, Napoli, Cosenza, Catani), INGV and ENEA; several researches have been performed in the time by foreign institutions from France, USA, Japan, UK and Germany.

The field guide includes a general overview of the coastal areas of Puglia, Basilicata, Calabria and Sicilia regions, comprising the geodynamic and geomorphological contest of this area in the Mediterranean basin as well as the evolution of its landscape.

The following sections describe the seven days of the conference, two of them dedicated to the annual scientific meeting followed by five days of field trip.

The 1th day of field trip is focused on the sea level changes registered along the Apulia foreland and Bradanic foredeep; the last stop will be held in the intra-chain basin of Sibari.

The 2nd day of field trip is focused on the interaction between the Pleistocene and Holocene sea level change and the uplift of the southern Calabria.

The 3rd day of field trip is focused the interaction between the Pleistocene and Olocene sea level change and the uplift of the area surroundings the Monte Etna volcano.

The 4th day of field trip is focused on the effect of Holocene sea level change and extreme wave impact on the coastal area of the Iblean foreland.

Data and results in this field guide are not definitive. Some of them are only the starting point of future research which will surely benefit of fruitful discussions and suggestions made on the field.

We warmly thank all the contributors, which made the realization of this field guide possible.

We wish all participants a great time in Southern Italy.

The organizers of the Conference

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Why a field trip in Southern Italy?

There are not other places in the world where different geodynamic domains change so rapidly in space as in southern Italy. The Foreland/Foredeep/Chain system of the Southern Apennines can be crossed along the Ionian coasts of Southern Italy in less than 50 km from E to W, similarly in Apulia and Basilicata or even from S to N in Southeasternmost part of Sicily. The geological evidences of the complex geodynamic framework of the Mediterranean basin outcrop in this region located across the Eurasia and African tectonic plates.

Tectonic setting

The morphological and structural setting of the Mediterranean region is the result of the evolution of deep basins and arcuate fault-and-thrust belts that originated during the long-lasting convergence between the major Africa and Eurasia plates running along an east west boundary (Fig. 1).
In recent decades, several seismotectonic syntheses have been proposed for the Mediterranean basin to describe the present-time tectonics and kinematics of this region. Of the several active tectonic structures in the region, the Hellenic Arc system is the largest and the most active, causing a broad deformation in the eastern Mediterranean, and is responsible for most of the largest earthquakes. In the central part, the subduction of the Calabrian Arc and the volcanism associated with the Aeolian Islands are the most important active features that control the kinematics of this area. In this tectonic framework, geological and geodynamic processes acting on different spatial scales are causing horizontal and vertical motion of the Earth surface that involves the coastal areas. The largest present crustal movements occur at rates up to 30–40 mm a\(^{-1}\) in horizontal directions for parts of the eastern (Aegean) Mediterranean, but rarely exceed 1–2.5 mm a\(^{-1}\) in the vertical. When extrapolated back in time, their cumulative effects, assuming that they have continued at similar rates over thousands of millions of years, have produced dramatic changes of the plate margins, mountain chains and coastal environments.

The largest critical effect of these movements occurred around 6 Ma BP, with the closure of the Gibraltar Strait, blocking the entrance of Atlantic waters into the Mediterranean basin and leading to the drying out of most of the enclosed sea. This is the Messinian Salinity Crisis during which the Mediterranean Sea lost most of its water through evaporation and its level dropped more than 1300 m below the current level (Fig. 2). Thick sequences of evaporites were deposited in the hyper-saline abyssal plains and are now exposed, for example, in salt mines in Sicily. The Messinian crisis ended about 5.3 million years ago, when the marine gateway to the Atlantic was restored. Since then, the evolution of the Mediterranean coastal zone has been governed by the interaction between tectonics and climate-induced sea-level oscillations, as shown by the stratigraphic sequences in subsiding coastal plains, shallow shelves or in drowned littoral caves that provide information on Pleistocene sea-level oscillations. An example of the palaeoshoreline reconstruction for more recent periods is given by Furlani et alii (2014), who examined the area between southern Sicily and northern Africa from the LGM to the Present. Malta island was connected to the European mainland via Sicily and only between 14.5 and 13.8 Ka cal BP, when local sea level was about 100–90 m lower than today, did Malta become an island. Presently, flights of marine terraces either submerged or on emerging coastlines preserve the signatures of sea-level highstands. These have been widely used to estimate long-term vertical land movements.

These surfaces are usually assumed to have formed during successive interglacials in an uplifting environment, but in most instances it is only the Last Interglacial shoreline of about 125,000 years ago that can be securely dated (Anzidei et alii, 2014 and references therein).
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Figure 2. Schematic palaeogeographic reconstruction of the Mediterranean coastlines during the Messinian salinity crisis, about 6 Ma BP. The reconstruction does not take into account the effects of geodynamic and tectonic displacement, and continental and coastal erosion, and it is therefore only indicative. Note the closure of the Gibraltar Strait, the separation of the Black Sea from the Mediterranean Sea, the continental link between Sicily and Africa and the drying of the Adriatic Sea. In blue is the estimated Messinian extension of the Mediterranean basin (coastlines are in black). Map is based on Shuttle Radar Topographic Mission (SRTM) data for surface topography and General Bathymetric Chart of the Oceans (GEBCO) data for bathymetry. In brown are the current coastlines. No allowance has been made for the isostatic rebound of the crust in response to the water unloading.

Figure 3. Plot of the vertical velocities (coloured circles) of GPS stations located less than 20 km from coastlines, together with a smoothed horizontal velocity field (with respect to a fixed Eurasian frame) interpolated over a regular grid. RM, Riff Mountains; GC, Gulf of Cadiz; CB, Cordillera Bética; CCR, Catalan Coastal Range; PM, Pyrenees Mountains; GG, Gulf of Genoa; TU, Tuscany; LA, Lazio–Abruzzi; CA, Campania; CArc, Calabrian arc; AP, Apulia; MA, Marche–Abruzzi; FR, Friuli; IG, Ionian Greece; CI, Crete.

This convergence has been active since the Late Cretaceous and currently is of the order of few mm per year. In consequence, the region is characterized by narrow seismic belts as well as a broad zone of seismicity and deformation that characterizes the Alpine belt, indicating a complex pattern of crustal stress and strain fields across these zones. The recent earthquake distribution also identifies the boundaries of minor plates whose interiors appear to be largely aseismic. About 50 subaerial and submarine volcanoes are active or quiescent in this region (http://www.volcano.si.edu/) (Fig. 3).
19th September 2018 – From Taranto to Sibari
The Apulian Foreland and the Bradanic Trough along the Taranto Gulf

The Apulian Foreland ("Avampaese apulo" of Selli, 1962) is part of the southern Adria plate and represents the Plio-Pleistocene foreland for the Southern Apennine orogenic system (Fig. 4). It has a continental and relatively thick lithosphere in the slab sector inducing the occurrence of a slow roll back and uplift in the foreland (Southern Adriatic Sea and Apulia, Doglioni et alii, 1994; Carminati et alii, 2012). This process suddenly ends along the northern side of the Apulian foreland, where the central northern Adriatic Sea is subject to a relatively faster roll-back and subsidence (Doglioni et alii, 1996). The crustal structure of the Apulian Foreland is made up of about 6 km of Mesozoic limestones (Apulia Carbonate Platform, D’Argenio, 1974) lying on a Variscan crystalline basement. Thin Tertiary and Quaternary deposits irregularly cover the Mesozoic units (Ricchetti et alii, 1988).

The Apulian Foreland forms a large antiform, with a WNW to ESE trend (Ricchetti and Mongelli, 1980), that is obliquely oriented with respect to the southern Apennines front (Casnedi, 1988). This antiformal structure is dissected in three main blocks by the occurrence of narrow deformation zones with an approximate EW direction (Doglioni et alii, 1994; Gambini and Tozzi, 1996): Gargano, Murge and Salento. Furthermore, the Apulian antiform shows a set of down faulted blocks (Carissimo et alii, 1962) both toward the foredeep (the Bradanic trough to WSW) and the Adriatic sea (to ENE). In the Murge area (Fig. 4), main tectonic structures with Quaternary activity are WNW-ESE elongated narrow grabens or semi-grabens (Murge Alte and Murge Basse grabens of Iannone and Pieri, 1982). In the eastern margin of the Bradanic Trough, the syntectonic sedimentary units (Calcarenite di Gravina Fm. and argille subappennine) are deformed during the Pliocene lower Pleistocene subsidence phase (Tropeano et alii, 1994). Along the Ionian side of the Salento peninsula (Fig. 4), normal faults with Plio-Pleistocene activity have been recognised in the Serre Salentine area (Palmentola and Vignola, 1980; Tozzi, 1993), along the coastal sectors (Martinis, 1962; Tropeano et alii, 2004) and in the shelf (western Salento, Tramutoli et alii, 1984). Therefore, the quantitative and systematic analysis of the joint sets affecting Quaternary deposits, allowed to recognize in the Salento area three different deformational events within a general extensional regime. The oldest event, constrained to the early and middle part of the Middle Pleistocene; the penultimate was referred to the Late Middle Pleistocene; and finally, the last, characterized by lowermost dynamic indicating a sort of horizontal ‘radial’ extension, is not older than the Late Pleistocene and possibly reflects the active stress field still dominating the entire study area (Di Bucci et alii, 2011).

Figure 4. Simplified sketch of the Apulia region. Foreland, Foredeep and Chain domains of the Southern Apennines are shown.

Murge and Salento are separated by a WE fault zone that forms a high scarp in the western sectors (Soglia messapica of Pieri, 1980) and corresponds to a triangular depression toward E (Pieri et alii, 1997). This WE fault zone (Taranto-Brindisi Fault Zone or North Salento Fault Zone, Tozzi, 1993; Gambini and Tozzi, 1996) shows evidences of Quaternary activity between Matera and Mottola horst.

The Bradanic Trough (Migliorini 1937; Selli 1962; D’Argenio et alii, 1973) is the Plio-Quaternary foredeep of the Southern Apennine orogenic system (Fig. 4) and is located between the chain and the flexured Apulian Foreland (Ricchetti and Mongelli, 1980). This foreland basin forms on a wide and complex subsiding sector of the Apulian Foreland (Sella et alii, 1988) and is filled by 3-4 km of Plio-Pleistocene siliciclastic deposits.
These deposits are mainly fed by the adjacent chain sectors which is structured as an eastward thrust sheet system. Contemporary trusting within the foredeep can form wedge-top basins isolated by the external subsiding foreland basin (the foredeep s.s. of De Celles and Giles, 1996) and more or less complex geometries in the external accretionary wedge.

The buried portion of the foredeep in fill consists of turbiditic deposits (Casnedi, 1988b), while the exposed portion (about 500-600 m thick) is mainly made up of hemipelagic silts and marls.

On the basis of the vertical and lateral relationships between turbiditic and hemipelagitic deposits and the position of the accretionary wedge, Tropeano et alii (2004) distinguish three main sectors with different sedimentary evolution. The inner successions (wedge top and western foredeep) contain part of the accretionary wedge (called “allochton” in the regional literature) that is tectonically superimposed on a thick turbiditic succession; the hemipelagites deposit on the “allochton” closing the succession. The foredeep s.s. successions develops close to the external sectors of the allochton and are made up, from base to the top, of hemipelagites, turbidites and hemipelagites. Finally, the middle and outer successions deposit in the marginal foreland ramp and are made up mainly of hemipelagites.

From the end of the early Pleistocene, both the Apulian Foreland and the Bradanic Trough underwent a slow uplift. This uplift allows the study of the units deposited during the subsidence phase (that crop out in many sectors of the foreland and foredeep areas) and is interpreted as induced by buckling processes related to the subduction of the thick continental lithosphere of the Apulian Foreland (Doglioni et alii, 1994). The interaction between the regional uplift and the dramatic sea level changes due to the alternance of climatic phases imposed the shaping of a sequence of marine terraces; their deposits cover the older units both in the Apulian Foreland and Bradanic Trough (for a synthesis: Mastronuzzi and Sansò, 2003; Ferranti et alii, 2006). Georcoronal data on the Late Pleistocene sedimentary units that crop out along the Ionian side of the Apulian Foreland and the Bradanic Trough show uplift rates decreasing from the higher rates of 0.3 mm/yr at the border between Bradanic Trough and Apulian Foreland to zero in the Southernmost part of Salento (Cosentino and Gliozzi, 1988; Dai Pra and Hearty, 1988; Belluomini et alii, 2002; Mastronuzzi et alii, 2007).

Finally, it is worth to mention that the Taranto Gulf contains most of the abovementioned geodynamic domains and represent the present-day Southern Apennine Chain System. Pescatore and Senatore (1988) recognize from E to W, the prosecution of the western Salento foreland ramp, the deep foreland basin area and the active front of the southern Apennine Chain. Furthermore, as in the Miocene and Plio-Pleistocene Chain/Foredeep/Foreland system, the growth of thrust sheets (for example, Amendolara Ridge) can separate the wedge top (the Sibari basin) from the foredeep s.s. basins (Amendolara basin). More details and an updated literature review on this intriguing area is contained in Ferranti et alii (2014) and Teofilo et alii (2018).

The uplift pattern in the Taranto Gulf sector of the Calabrian Arc

Several orders of Middle-Late Pleistocene marine terraces are found along the Taranto Gulf, a major embayment of the Ionian Sea (Bruckner, 1980; Amato et alii, 1997; Westaway, 1993). The elevation of the terraces decreases steadily to the NE along the former foredeep basin of the Southern Apennines thrust belt in Basilicata region, to a minimum at the north-eastern shore of the Taranto Gulf in Apulia region. No significant jump in the regional uplift pattern is observed crossing the front of the Southern Apennines, supporting the notion that thrust displacement had vanished by the onset of the Middle Pleistocene (Patacca and Scandone, 2007). Different interpretations of the large-wavelength uplift tapering were published by Westaway and Bridgland (2007) and Caputo et alii (2010). Whereas the first authors highlighted the thermal response to sediment load in promoting crustal uplift through lower crustal flow directed beneath regions of prevailing erosion, the latter have stressed the contribution from both deep crustal and lithospheric contraction and from shallow-crustal folds. Moving to the south, in northern Calabria, the terraces elevation reaches a maximum. Ferranti et alii (2009) and Santoro et alii (2009; 2013) have shown that a regional and a local component are embedded in the deformation profile of Middle-Late Pleistocene marine terraces. The magnitude of the regional component is almost four times greater than the local component, the latter being attributed to growing folds within a transpressive displacement field. Santoro et alii (2013) mapped ten terrace orders uplifted to as much as +660 m along ~80 km of the Taranto Gulf coastline. The intermingling between the two deformation sources, regional and local, is attested by two 10 km scale undulations superimposed on a 100 km scale northeastward tilt. The undulations spatially coincide with the trace of NW-SE striking transpressional faults that affected the coastal range during the early Pleistocene. Fault numerical models predict the undulations derive from two fault-propagation folds cored by blind thrusts. The locus of predominant activity has repeatedly shifted between the two fault systems during time and slip rates on each fault have temporally changed. It is not clear if the active deformation is seismogenic or dominated by aseismic creep; however, the modeled faults are embedded in an offshore transpressional belt that may have sourced historical earthquakes (Ferranti et alii, 2014).
Sea Level changes markers

Along the Mediterranean coasts Late Pleistocene coastal deposits are arranged into widespread marine terraces (i.e. Bonifay and Mars, 1959; Selli, 1962; Paskoff and Sanlaville, 1983; De Muro and Ulzega, 1985; Hearty et alii, 1986a; Hillaire-Marcel et alii, 1996 and references therein). In southern Italy they occur as a nearly continuous and wide “flight-of-stairs” around coastal region. Generally the deposits of the lowermost ones are characterised by a fossil faunistic assemblage that in the Mediterranean geological context, is indicated as “Senegalese”; the term Senegalese fauna indicates a fossil faunistic assemblage consisting of warm species, from the Atlantic (Gignoux, 1911a, b) where “… the most famous and common is Strombus bubonius Lamarck 1822” (Gignoux, 1913). In a no-complete list, other warm water species are represented by Patella ferruginea, Conus euminus, Gemophos viverratus, Cardita calyculata senegalensis and Hyotissa hyotis.

The gastropod Persististrombus latus (Gmelin, 1791) (= S. bubonius Lamarck, 1822 (Nalin et alii, 2012) entered the Mediterranean Sea only during the Tyrrhenian time corresponding to the Last Interglacial Time (LIT) and to the Marine Isotope Stage 5.e (MIS 5e) (indicated also as MIS5.5 or Oxygen Isotope Stage - OIS 5.5); this term - in the form “Tirreno” - has been proposed for the first time by Issel (1914) and then by Dépèret (1918) for indicating the age of raised marine terraced deposits characterised by the presence of Senegalese fauna with or without P. latus.

Bonifay and Mars (1963) affirmed that the S. bubonius (today P. latus) is not the characteristic element of the Tyrrhenian and Senegalese deposits. In fact, it is very important to underline that it is only a fossil of facies; not all Tyrrhenian deposits (“Tirreno” in Issel, 1914) with warm fauna of Atlantic origin (sensu Gignoux, 1911a,b; 1913) or Senegalese (sensu Bonifay and Mars, 1963) must be characterised by the presence of this tropical gastropod (Fig. 5).

![Figure 5](image1.jpg)  
**Figure 5.** Main components of the Senegalese fauna recognised in the surroundings of Taranto: a - Persististrombus latus (Masseria Abateresta, Talsano, Taranto); b - Hyotissa hyotis (Il Fronte, Taranto); c - Cardita calyculata senegalensis (Il Fronte, Taranto). In d a colony of Cladocora caespitosa from Masseria Santa Teresiola (Taranto) is shown.

Usually, the Tyrrhenian is correlated to the Last Interglacial Time (LIT) MIS 5e which occurred between 132 and 116 ka roughly (Shackleton, 2000; Shackleton et alii, 2003), but it is generally extended up to 80ka (Fig. 6); this lap of time has been defined thanks to different age determinations performed by means of U/Th analysis on the coral Cladocora caespitosa and amino acid racemization analyses on mollusc shells as Glycymerts sp, Arca sp and Cerastoderma sp. associated to the before mentioned taxa (e.g.: Mastronuzzi and Sansò, 2003; Antonioli et alii, 2009).

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The Last Interglacial, being characterized by remarkable ice-sheet retreat, entails significantly warmer conditions and higher eustatic sea level than at present (i.e.: Sánchez-Goñi et alii, 2012 and references therein). Hence, the MIS 5e interval is generally regarded as a good analog of near-future Earth developments in response to the projected global warming (i.e.: Siddall and Valdes, 2011 and references therein). Dramatic oceanographic and climatic changes related to the Last Interglacial occurred over few millennia, and are recorded in several continental, marine and icecore successions of the Northern Hemisphere (i.e.: Shackleton et alii, 2002, 2003; Tzedakis et alii, 2003; Martrat et alii, 2004; Brauer et alii, 2007; Couchoud et alii, 2009; Milner et alii, 2013).

Along the Mediterranean coasts a variety of sea-level indicators, including tidal notches, marine terraces and raised beaches suggest an average sea level for the Last Interglacial 7±3m higher than modern sea level (Lambeck et alii, 2004; Ferranti et alii, 2006; Antonioli et alii, 2018). On the other hand, the present elevation of the Tyrrhenian deposits in southern Italy depends on the geodynamic history of the area (Ferranti et alii, 2006 and references therein) (Fig. 7).

For instance, Tyrrhenian deposits containing senegalese fauna have been identified at heights between 157 and 1–10 m a.s.l. Nonetheless, different attempts to calculate the uplift rates of marine deposits with Senegalese fauna have been imprecise. In fact, where there are no clear geomorphological and/or biological


Figure 7. Distribution of MIS 5e marine deposits bearing Persististrombus latus, or Senegalese fauna along the Italian coasts. A qualitative estimation of the site richness of P. latus specimens is indicated by the different dot fill. The black and pink lines indicate the present-day and the MIS 5e highstand shoreline, respectively (a); Uplift rate in mm/ya; for the colour classes see the legend. Inset shows a schematic three-dimensional representation of the same data set (b) (modified from Ferranti et alii, 2006).
evidence of former sea-levels, the individuation of past sea level is only function of the approximation that can be derived by the facies analysis. On the other hand, jointly to the sedimentological and palaeontological evidences, some geomorphological markers can permit the identification of past sea level stands occurred during the interglacial times. Notches, wave cut platforms, inner margins, alignment of sea caves, beach rocks and the bioconstructions represented by algal rims arranged in trottoir permit to individuate with high definition the position of past sea levels; unluckily, few are the possibility to attribute a significant age performing direct age determinations on them. Only the presence of speleothems permit to obtain geochronological ages. In the biggest number of cases, it is necessary to correlate coastal landforms to the nearest deposits containing Senegalese fauna or Cladocora caespitosa. Only in more recent time a complete review of the occurrence of the Tyrrhenian deposit and sea level markers all around the Italian peninsula permitted to have a precise estimation of the height of the related former sea level and uplift rates in the context of the geodynamic evolution of the Mediterranean basin (Ferranti et alii, 2006; Antonioli et alii, 2018).

Absolutely different is the degree of approximation that can be adopted to determine the sea level history during the Holocene; in fact, the possibility to use AMS age determination both on palaeontological remains, bioconcrections and speleothems, jointly to the correlation deriving from geoarchaeological data permit to detail the change in sea level position in relation to the land movement with an higher degree of confidence in the vertical space and in the time (for more detail, see, i.e.: Antonioli et alii, 2009; Aurienma and Solinas, 2009; Mastronuzzi et alii, 2017).

The Area of Taranto

The Taranto area records the main Quaternary tectonic and sedimentary phases that involve the entire Apulian foreland. Until the late Early Pleistocene, the Apulian foreland underwent a phase of extensive subsidence (Ciaranfi et alii, 1988) induced by active subduction beneath the southern Apenninic Chain (Doglioni et alii, 1994). This subsidence phase (with a rate of about 1 mm/y) is recorded by the Calcarenite di Gravina Formation, which transgressively overlies Mesozoic carbonate rocks, and the argille subappennine unit, which conformably covers the Calcarenite di Gravina Fm. The subsequent uplift phase (with a rate of about 0.1–0.3 mm/y), active at least from the beginning of the Middle Pleistocene (Ciaranfi et alii, 1988; Doglioni et alii, 1994; Ricchetti et alii, 1988), is recorded by uplifted marine, transitional and fluvial terraces. In fact, from the Middle Pleistocene onward the superimposition of regional uplift and glacialeustatic sealevel changes produced a number of steplike marine terraces that ranged in elevation between 400 m a.s.l. and present day sea level. Some of these terraces are characterised by thin bioclastic calcarenite deposits, known as “panchine” whereas others are only represented by abrasion surfaces. Marine terraces retain their original seaward slope, but show some effects of neotectonic activity, in the form of slight tilting or fracturing of the calcarenite bodies. Two terraces, with elevations between 20–30m and present sea level, are historically correlated with the Tyrrhenian high sea level stands (Gigout, 1962; Dai Pra and Stearns, 1977; Hearty and Dai Pra, 1985) (Fig. 8).

Figure 8. Geological sketch of the Taranto area (A), with cross-sections (B) of the Taranto area: A) Mesozoic carbonates; B) Gravina Calcarenite (Upper Pliocene to Lower Pleistocene); C) Blue Clay (Upper Pliocene to Lower Pleistocene); D) Middle Pleistocene (?) marine deposits; E) Upper Pleistocene (MIS 5) marine deposits; F) Holocene alluvial and beach deposits; G) Reclaimed areas (XIX–XX centuries); H) Main karstic subaerial springs; I) Main karstic underwater springs (citri); L) Stratigraphic logs in Fig. 9; M) Other outcrops with “Senegalese” fauna; N) Cores;O) Strike and dip of strata (1, sub-horizontal; 2, b15°; 3, N15°); P) Inferred normal fault; Q) Section trace (modified after Mastronuzzi, 2001 in Amorosi et alii, 2014).

Crossing Southern Italy: a travelling meeting from Taranto to Siracusa
The area of Taranto (Fig. 10) is marked by three main morphological features: i - the quasi-perfect flat surface that marks the coastal landscape around the city; ii - the apparently circular depression of the Mar Piccolo and Mar Grande inlets; iii - the deep incisions locally named “lame” or “gravine” or “valloni” that cut these surface drawing a river catchment marked by about N-S and E-W oriented water flow (Mastronuzzi and Sansò, 2002; 2003; Mastronuzzi et alii, 2013).

Figure 9. Stratigraphic correlation of Fronte Section with coeval outcrops of the Taranto area (for location, see Fig. 8). Numbers refer to the stratigraphic units of Fig. 13. The red line is the lower boundary of MIS 5e deposits. Note the widespread occurrence of two distinct Panchina layers (units 3 and 5). A - Blue Clay unit; B - tephra; C - coarse clastics; D - lacustrine deposits; E - massive calcarenite; F - cross-laminated sandstone; G - silt; H - sand; I - travertine. 1 - Persististrombus latus (= Strombus bubonius); 2 - disarticulated Cerastoderma valves; 3 - articulated Cerastoderma valves; 4 - Ostreids; 5 - Cladocora caespitosa; 6 - bivalves; 7 - gastropods; 8 - disarticulated bivalves; 9 - rhodoliths; 10 - Echinocardium cordatum ichnotraces (after Amorosi et alii, 2014).

Figure 10. The quasi-perfect flat surface on which Taranto has been built (a); the apparently circular depressions of the Mar Piccolo (on the left) and Mar Grande (on the right) inlets view from North (http://www.tuttamialacitta.it) (b); the deep incisions of the vallone Riggio (c).
In a first time, this characteristic perimeter has been referred to karstic collapses originated in the Mesozoic substrate (Verri and De Angelis D’Ossat, 1899; De Giorgi, 1922; Parenzan, 1960). According to Guerricchio (1988), the anti-clockwise rotation of the Salento area is responsible for the shaping of the depressions of the Mar Piccolo and Mar Grande; Pagliarulo and Bruno (1990) suggest that the same depressions have been generated by tectonic activity involving also the Plio-Pleistocene units. Finally, Guerricchio and Simeone (2013) interpret the Mar Piccolo depression as the result of - fluidization of the overlaying Plio-Pleistocene sedimentary cover induced by submarine karstic springs and - lateral erosion related with helicoidal vertical flows.

More recently, the morphological evolution of the Mar Grande and Mar Piccolo basins has been attributed to an initial phase of deep river erosion during the glacial emicycle that induced changes in the drainage network in the relation to the low sea level stand during the LGM (Mastronuzzi and Sansò, 1998; 2003; Mastronuzzi et alii, 2013); the following Holocene transgression is the responsible of the flooding of rivers valleys that crossing the present state of “rias” – enlarged by the retreat of the cliffs shaped on the Argille subappennine by the wave erosion – will produce a filled valleys system (Valenzano et alii, 2018).

Figure 11. A) Alternance of silt with Planorbis sp, travertines and laminated calcarenites near Sorgente Galeso; B) massive bioclastic calcarenite (a P. latus specimen can be seen on the right of the hammer) at Cimino locality; C) laminated biocalcarenites with abundant beached Cerastoderma sp at Masseria Tuglia; D) Blue Clay unit is at the base of the cliff of Punta Penne covered by silty-sands with P.latus; E) cross laminated calcarenites with ichnotraces at the Castello Aragonese.
The marine terrace deposits outcropping in the area of Taranto, at an elevation ranging from 23 to about the present sea level, have been studied by Verri and De Angelis D’Ossat (1899), Gignoux (1913), Gigout (1960, 1962), Ricchetti, (1967, 1972), Caldara and Laviano (1980), Caldara (1987), Dai Pra and Hearty (1985; 1992) and Hearty and Dai Pra (1992). It overlies in transgression the local basement of Mesozoic limestones and/or the Upper Pliocene – Middle Pleistocene units represented by the Bradanic sequences constitute by the Calcarenite di Gravina formation and by the Argille subappennine (= Blue Clay unit) and, finally, partly the older “panchine” deposits referable to OIS 7 and, probably, to OIS 9 (Dai Pra and Stears, 1977; Hearty and Dai Pra, 1992; Negri et alii, 2015). In the area of Mar Piccolo, the presence in the Blue Clay unit of Tephra layers and the fossil contents indicate an age of about 500 ka as suggested by the occurrence of *Pseudoemiliana lacunosa* calcareous nanofossil and the absence of *Gephyrocapsa omega* (Maiorano and Marino, 2004; Raffi et alii, 2006; Amorosi et alii, 2014; Negri et alii, 2015).

Around Taranto, numerous good exposures represented by lateral and vertical alternance of beach sands and algal calcarenites, dune cross laminated calcarenites, travertins and silty sands with *Planorbis* sp. indicate the highly variable facies of this deposits (Figg. 9,11), which is normal for the past articulated coastal sea bed. Mainly they are characterised by Senegalese fauna, with abundant specimens of *P. latus*, *Cardita calyculata senegalis*, *Hyotissa hyotis* and/or more or less extended colonies of *Cladocora caespitosa*. *C. caespitosa* is a bioherm-building coral that is generally associated with warm-temperate faunas and has been considered a good marker of warm or warm-temperate waters (Peirano et alii, 1996, 2004; 2009). This fossil content, along with an impressive set of relative (amino acid racemization) and radiometric (U/Th) dates (i.e.: Mastronuzzi and Sansò, 2003; Antonioli et alii, 2008; Amorosi et alii, 2014 and references therein), indicate the occurrence in this area of a tropical environment of Late Pleistocene age between 132 and 116 ka, corresponding to MIS 5e already named Tyrrhenian. In particular, all along the Mar Grande and Mar Piccolo cliffs, late Pleistocene deposits crop out with large lateral continuity: for this reason, the last part of the Pleistocene (132.000 to 80.000 years ago, corresponding to about the entire MIS 5) has been proposed to be named “Tarantian stage” and the Taranto area is suggested as the stratotype (Amorosi et alii, 2014; Negri et alii, 2015, and references therein).

The marine terrace is dissected by narrow, straight valleys that represent an important morphological element of the landscape converging in the main river Canale D’Aiedda - Leverano d’Aquino and than in the Mar Piccolo and in the Mar Grande inlets (Mastronuzzi and Sansò, 2002).
The cliff of the Il Fronte promontory (Fig. 12) guests the most important stratigraphic sequence ascribed to the MIS 5e. This area has been studied by several Authors (Dai Pra and Stearns, 1977; Hearthy and Dai Pra, 1992) who investigated the geology of the Taranto Gulf which was then proposed as a GSSP site of the LIT.

For its geological, stratigraphic, paleontological and geomorphological implication, it has been also registered as Geosite of Special Interest and Natural Monument with code CGP0432 in the census of Geosites of Puglia Region (www.geositipuglia.eu; Mastronuzzi et alii, 2015). The following descriptions are synthetically reported from Amorosi et alii (2014) and Negri et alii (2015).

Figure 12. The Il Fronte promontory; yellow numbers indicate the units of the MIS 5e deposits (cfr Fig. 13)

**The Blue Clay unit**

The lower part of the Il Fronte stratigraphic sequence at the base of the cliff is shaped on the Blue Clay unit (Fig. 13) here represented by about 5 m of massive gray clay, with upward transition to silty clays with silt intercalations. In the upper part, two distinct one-cm-thick tephra layers, separated by a thin (20 cm) mud deposit, were observed.

The lower one is white, almost entirely volcanoclastic, fine-grained and massive. The upper layer is yellowish, highly weathered, coarser and with a bioclastic and silicoclastic component. Both are interpreted to represent quite distal fall-out events.

A laminated interval above the tephra layers, about 20 cm thick (Figg. 9,13), shows features typical of a sapropel. The fossil content of the Blue Clay unit is consistent with outer shelf/upper slope muddy environments in an estimated bathymetric range of about 130–400 m (e.g., La Perna, 2003). Grain size characteristics and the mollusk content suggest an overall shallowing upward trend. As a whole, the microfossil assemblages testify to a gradual and progressive deterioration of oxygenation at the sea floor, which preceded the development of the sapropel like assemblage clearly visible in the overlying samples.

Upwards, the rapid transition to shallower environments is witnessed by the presence of sublittoral mud lover taxa and by the decreasing planktic/benthic foraminifera ratio, both suggesting sedimentation in middle to innershelf depositional environments. Calcareous nanofossils from the lowermost muds include important chronostratigraphic markers that constrain the age of this unit between about 1Ma and 480 ka (Raffi et alii, 2006).

The age of this stratigraphic interval can further be refined by the lack of G. omega, a Mediterranean marker that shows its highest occurrence around 577 ka (Maiorano and Marino, 2004). Based on these biostratigraphic data, and specifically on the occurrence of P. lacunosa and the absence of G. omega, the laminated interval above the tephra layers is tentatively correlated to sapropel S12, dated at 502 ka (Lourens, 2004).

The association of Carya, Pterocarya and Tsuga spp in very low percentages suggests an age attribution of the base of this unit to about 500 ka (Amorosi et alii, 2014; Negri et alii, 2015 and references therein).
MIS 5e deposits

MIS 5e deposits consist of five vertically stacked facies associations (units 1 to 5 in Fig. 13), which are illustrated in ascending order.

Unit 1

This unit consists of a very thin (25 cm) bioclastic, fine-sandy mud layer with common oyster shells and fining-upward bioclasts in sandy matrix. This fossil-rich horizon shows an unconformable lower boundary with the underlying Blue Clay unit, highlighted by the sparse occurrence of the mud-boring bivalve *Pholas dactylus*, articulated and in physiological position (Fig. 14). The sandymud onmud contact is clearly marked by a veneer of mollusks (abundant *Ostrea edulis* and *Bittium* spp.; common *Cerithium vulgatum*; rare *Cerastoderma glaucum*, *Cernuella* spp.). Upwards, the macrofauna includes abundant *Abra alba*, *T. distorta*, *Pitar rudis*, *Ostrea edulis* and *Antalis* spp., among others.

This highly-fossiliferous horizon is characterized by a rich mollusk assemblage representing a variety of subaerial to nearshore paleoenvironments. The presence of the diagnostic, coastal-lagoonal *P. dactylus* at the lower bounding unconformity testifies to an indurated bottom with prolonged break in sedimentation. A deepening-upward tendency within unit 1 is suggested by the overall fining-upward trend, and is supported also by the peculiar macrofaunal assemblage, suggesting admixture from different sources. In particular, mollusk shells sourced from subaerial (*Cernuella* spp) and lagoonal (*C. glaucum*) environments are overlain by nearshore to inner shelf, fully marine mollusks. Regarding the latter component, unit 1 shows high abundance of taxa characteristic of either muddy or sandy littoral bottoms (e.g., *A. alba* and *P. rudis*, respectively).

Unit 2

This unit, less than 1 m thick, is made up of highly fossiliferous, homogeneous gray mud. Diagnostic feature of this unit is the conspicuous, increasing-upward occurrence of in situ colonies of scleractinian coral *C. caespitosa*, with intracoral pelitic matrix. The ubiquitous presence of *Cladocora* colonies makes unit 2 a laterally continuous and easily identifiable stratigraphic marker across the study outcrop. The macrofauna is represented by a low diversity association, with sparse *Arca noae*, *Mimachlamys varia*, *D. lupinus*, common *A. nitida*, *Nucula* gr., *nucleus*, *T. distorta*, oysters and abundant *C. gibba*. Unit 2 is also characterized by the dominance of the euryhaline foraminiferal species *Ammonia tepida* and *Ammonia parkinsoniana*, with variable frequencies of *Cribroelphidium* species. The ostracod fauna exhibits the overwhelming dominance of the shallow-marine opportunistic species *Palmoconcha turbida*, accompanied by *Palmoconcha agilis*. An increase in: (i) species richness, and (ii) the relative abundance of...
typical marine taxa, such as *Ammonia beccarii* and *Costa edwardsii*, is recorded in the middle part of the unit. Among dinocysts remarkable is the presence of *Poly sphæridium zoharyi*, characteristic for euryhaline tropical/subtropical coastal sites from mesotrophic environments. Ten corallites of *Cladocora caespitosa* sampled at the two distinct stratigraphic levels (C1 and C2 in Fig. 13) yielded consistent uncorrected U-series ages, ranging between about 140 and 131 ka (C1 in Tab. 1) and between about 125 and 121 ka (C2 in Tab. 1), respectively, with just one exception.

### Table 1. U-series ages of *Cladocora caespitosa* corallites collected within unit 2 from two distinct stratigraphic levels (C1 and C2 in Fig. 13); lower case letters represent corallites from the same reef complex (Amorosi *et alii*, 2014).

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Location</th>
<th>U-series Age</th>
<th>226Th (nGPa)</th>
<th>232Th (nGPa)</th>
<th>226/232Th</th>
<th>238U/232Th</th>
<th>Age (kyr)</th>
<th>Age (kyr)**</th>
</tr>
</thead>
<tbody>
<tr>
<td>PW11-C1-a</td>
<td></td>
<td>3.74 ± 0.000</td>
<td>40.82 ± 0.000</td>
<td>155 ± 9</td>
<td>288 ± 1</td>
<td>0.08111 ± 0.00002</td>
<td>31.91 ± 1</td>
<td>184 ± 1</td>
</tr>
<tr>
<td>PW11-C1-b</td>
<td></td>
<td>3.66 ± 0.002</td>
<td>30.98 ± 0.017</td>
<td>194 ± 9</td>
<td>106 ± 1</td>
<td>0.0821 ± 0.0001</td>
<td>35.9 ± 1</td>
<td>190 ± 1</td>
</tr>
<tr>
<td>PW11-C1-c</td>
<td></td>
<td>3.79 ± 0.003</td>
<td>32.28 ± 0.020</td>
<td>156 ± 9</td>
<td>190 ± 1</td>
<td>0.08401 ± 0.0001</td>
<td>14.19 ± 1</td>
<td>204 ± 1</td>
</tr>
<tr>
<td>PW11-C1-d</td>
<td></td>
<td>3.71 ± 0.000</td>
<td>40.66 ± 0.035</td>
<td>194 ± 9</td>
<td>203 ± 1</td>
<td>0.0839 ± 0.0009</td>
<td>19.09 ± 1</td>
<td>190 ± 1</td>
</tr>
<tr>
<td>PW11-C1-e</td>
<td></td>
<td>3.48 ± 0.002</td>
<td>28.19 ± 0.014</td>
<td>155 ± 1</td>
<td>119 ± 3</td>
<td>0.08134 ± 0.0001</td>
<td>13.2 ± 1</td>
<td>196 ± 1</td>
</tr>
<tr>
<td>PW11-C1-f</td>
<td></td>
<td>3.73 ± 0.005</td>
<td>31.19 ± 0.021</td>
<td>136 ± 8</td>
<td>316 ± 3</td>
<td>0.08152 ± 0.0009</td>
<td>17.81 ± 1</td>
<td>205 ± 1</td>
</tr>
<tr>
<td>PW11-C2-a</td>
<td></td>
<td>3.89 ± 0.005</td>
<td>22.99 ± 0.034</td>
<td>132 ± 1</td>
<td>141 ± 2</td>
<td>0.0827 ± 0.0009</td>
<td>12.4 ± 1</td>
<td>105 ± 1</td>
</tr>
<tr>
<td>PW11-C2-b</td>
<td></td>
<td>3.96 ± 0.002</td>
<td>33.38 ± 0.010</td>
<td>184 ± 8</td>
<td>419 ± 3</td>
<td>0.0764 ± 0.0002</td>
<td>12.4 ± 1</td>
<td>105 ± 1</td>
</tr>
<tr>
<td>PW11-C2-c</td>
<td></td>
<td>4.12 ± 0.008</td>
<td>17.47 ± 0.010</td>
<td>131 ± 1</td>
<td>556 ± 4</td>
<td>0.0725 ± 0.0001</td>
<td>12.57 ± 1</td>
<td>105 ± 1</td>
</tr>
<tr>
<td>PW11-C2-d</td>
<td></td>
<td>3.89 ± 0.005</td>
<td>14.89 ± 0.011</td>
<td>126 ± 1.2</td>
<td>628 ± 7</td>
<td>0.0767 ± 0.0004</td>
<td>12.2 ± 1</td>
<td>105 ± 1</td>
</tr>
</tbody>
</table>

The presence and the abundance of in-situ growing *C. caespitosa* colonies and by the characteristic, muddy inner-shelf shallow macrofauna (Scarpone and Angeletti, 2008) suggest the establishment of stable, shallow-marine conditions. The occurrence of intracoral pelitic matrix points to a likely baffling effect exerted by the branching reef at the expense of pulses of muddy sediment from the coast. The exclusive occurrence of sublittoral mollusks may be taken as an indication that the reef was thriving at paleo depths of 10–20 m, and that muds were settling down beneath that level.

### Unit 3

This unit forms a prominent stratigraphic marker, about 2 m thick, represented by a characteristically cemented, fossil-rich, calcarenite (regionally called panchina), yielding abundant *C. caespitosa* (Fig. 15). This sandy-hash skeletal facies exhibits a transitional boundary with the underlying unit 2 (progressive upward increase in grain size), concurrently with the increase in abundance of sessile mollusks, such as *Spondylus gaederopus* and *Chama gryphoides*.

![Figure 15. Colonies of *Cladocora caespitosa* in the Units 2 and 3 at Il Fronte.](image)

Unit 3 is highly macrofossiliferous and fed by skeletal allochems, mostly mollusks, mainly bivalves and gastropods, followed in abundance by coralline algae, serpulids, and subordinate echinoids. Reworked, loose *Cladocora* corallites are common. Remarkably, the ‘panchina’ facies contains examples of the well known “Sengalese” fauna, in particular: *P. latus, Polinices lacteus, G. viverratus* and *C. ermineus* (= *C. testudinarius*).

The only clearly in situ macrofossils are represented by articulated sparse specimens of large infaunal bivalves (*Lutraria angustior* and *L. oblonga*).
Unit 4

This facies association consists of a 4.5 m-thick, reddish to greenish pelite characterised by a low-diversity mollusk fauna. Two highly diversified deep-sea meiofauna associations are recorded within unit 4. The lowermost and upper portions of this unit are mainly represented by the epifaunal-shallow infaunal species *Balimina marginata* and *Cibicidoides pachyderma*. On the other hand, the lowermiddle part of unit 4 shows scarce autochthonous ostracod fauna, mainly represented by the middle outer shelf species *Henryhowella sarsi* and *C. turbida*, and records higher abundance of the opportunistic, shallow infaunal *U. mediterranea* and *U. peregrina*, which become the dominant taxa along with *B. marginata* and *C. pachyderma*. The palynological assemblage, although almost devoid of pollen (rare grains of Pinus plus Olea, Chenopodiaceae, Brassicaceae and Rosaceae) and dinocysts, contains as a scattered occurrence *Impagidinium patulum*, a temperate to tropical oceanic species commonly well abundant under oligotrophic conditions, but also present under eutrophic conditions. Sparse cysts of *O. centrocarpum*, *Impagidinium spp.*, *Spiniferites ramosus* and *P. zoharyi* are also present.

Unit 5

The topmost part of Fronte Section is represented by another fossiliferous calcarenite about 1m thick (similar to that shown in Fig. 11C). Similar to unit 3, this calcarenite contains loose and dispersed corallites of *C. caespitosa*. The macrofauna of this unit consists of mollusks between which *Acteocina knockeri* and *Spisula subtruncata*, with subordinate coralline algae, serpulids, and echinoids. The presence of *A. knockeri*, a minute gastropod of tropical/subtropical affinity, suggests average water temperature higher than today.

Paleoenvironmental evolution during the Last Interglacial

The vertical stacking of units 1 to 5 provides the basis for reconstructing the paleoenvironmental evolution of the Taranto area during the Late Pleistocene. Although age attribution of the underlying Blue Clay unit is uncertain, the presence in this unit of *P. lacunosa* and the concomitant absence of *G. omega* enable a generic assignment to MIS 13 (Fig. 13). In contrast, the overlying succession (units 1 to 5 in Fig. 13) can readily be assigned to MIS 5e based upon four cross-checked proxies: (i) the occurrence within unit 3 of the characteristic “Sengalese” fauna, which is traditionally taken as an indication of the Tyrrhenian sensu stricto (= MIS 5e); (ii) the finding within units 2 and 4 of dinocyst *P. zoharyi*, previously reported within Sarpel S5 from several sites of the Eastern Mediterranean; (iii) a new set of ten U-series ages from unit 2 (Tab. 1), (iv) and its consistency with several tens of previously published U-series uncorrected ages from the same depositional units (Amorosi et alii, 2014; Negri et alii, 2015). The stratigraphic unconformity recorded atop Blue Clay unit, marked by the diagnostic *P. dactylus* horizon (Fig. 14), records a prolonged phase of non deposition in the study area. Above this indurated, and probably emerged surface (during MIS 6 sea level was ca. 130 m lower than today; Waelbroeck et alii, 2002; Lambeck et alii, 2004), deposition of unit 1 marks the first episode of transgression of the MIS 5e sea onto the mid-Pleistocene (Blue Clay unit) mud. Superposition of Cladocora-rich muds (unit 2) onto the basal, fossiliferous sandy muds (unit 1) reflects a rapid deepening-upward trend, with transition from brackish/nearshore to inner-shelf (b20 m deep) environments subject to fluvial discharge fluxes. A shallowing-upward tendency, paralleled by a progressive increase in grain size and the change to a shallower fauna, is reconstructed from upper unit 2 and at the transition to the overlying calcarenite body (unit 3). This prominent marker bed is interpreted to record a shortlived episode of coastal progradation, possibly induced by a phase of sea-level stillstand. The sharp boundary to the overlying muds (unit 4) marks an abrupt increase in paleobathymetry, with the establishment of a middle-outlet shelf environment (b100 m) characterized by mesotrophic conditions and limited oxygen depletion at the sea floor (as testified by the co-dominance of *C. pachyderma* and *B. marginata*). In the middle part of unit 4, the microfaunal turnover toward slightly deeper environments, with higher nutrient concentration and oxygenation (increasing percentages of *Uvigerina* species paralleled by a marked decrease of *B. sphautula*), marks the MIS 5e peak transgression. Increased reworking, concurrently with a slight shallowing-upward trend, is recorded in the uppermost three meters of unit 4. The boundary with another calcarenite body containing warm-water, inner-shelf mollusk fauna (e.g., *A. knockeri*, and *S. subtruncata* in unit 5) is inferred to reflect, again, “normal” shoreline regression (progradation) under conditions of sea-level stillstand.
STOP 2 - The Mar Piccolo and Mar Grande inlets (Lat. 40°29’55’’N – Long. 17°14’44’’E)

In the last five years, the entire area of the Mar Piccolo and part of the Mar Grande area, have been object of an important geological surveys aimed to the construction of its geological model and to the recognition of the pollution features. The entire work has been conducted by researcher of the Department of Earth and Geoenvironmental Sciences of the University of Bari thanks to an agreement between the Government Commissioner for the Remediation of the Taranto Area and the Rector of the University of Bari.

The surveys activity has constitute in the complete survey of the sea bottom by means of Side Scan, Multibeam and Magnetometer, coupled to a survey by means of Sub Bottom Profile, Sparker, Marine Geoelectric and to a complete corer campaign that permitted to obtain 25 corers from the interface water/sediment up to the Blue Cay units, and to have more than 40 meters of Holocene sediments.

The integrated analysis of more than 250 km high resolution seismic reflection profiles together with the analysis and dating of about 1,5 km of corers performed all around the Mar Piccolo and Mar Grande area made possible a complete reconstruction of its evolution during the Holocene in the context of the still active dramatic deglaciation. Only a small part of them have been preliminarly used for the reconstructions already performed (Fig. 16). The descriptions of the surveys performed in the Mar Piccolo and Mar Grande inlets and the reconstruction of the morphoevolutive history are synthetically reported from Valenzano et alii (2018).

Figure 16. Position of the geophysical datasets; thin lines represent all acquired seismic lines, bold lines represent profiles shown in Fig. 17 ab.

Seismo-stratigraphy

Seismic analysis highlighted the presence of two key seismic horizons (Horizon A and Horizon B, Fig. 17, which together with the seabed, are the boundaries of three main seismic units (S1, S2, S3, Fig. 17).

Figure 17. a) Sparker profile (A) and interpretation (B); b) Sub-bottom profile (A) and interpretation (B); position of the profiles is displayed in Fig. 16.
Seismic Unit S1

The deeper seismic unit, named S1, represents the acoustic basement in sparker profiles, it is bounded on top by the Horizon A. Unit S1 is marked by high amplitude chaotic and discontinuous reflections and can be interpreted as a hard substratum. Horizon A is visible from about 29 m up to 82 m below sea bottom (= b.s.b.) and constitutes a key stratal surface between Seismic Unit S1 and the overlying S2, it is a rough irregular unconformity, detected only in sparker profiles, showing some local gentle depression.

Seismic Unit S2

Seismic Unit S2 is delimited at the bottom by Horizon A and at the top by Horizon B. It appears as a well layered seismic facies with continuous medium amplitude reflectors onlapping Horizon A (Fig. 17). The internal stratification is generally sub-horizontal with some very smooth folding and is visible with remarkable detail in the sparker record. Reflectors are truncated and terminate top lap under Horizon B. Parametric sub bottom profiles show a consistent decrease in signal penetration in correspondence of this unit that seemingly suggests a substratum harder than the overlying seismic unit S3. Horizon B represents an uneven surface, which deepens toward the central part of the two sub-basins, up to a maximum depth of about 40 m b.s.b. corresponding to 52 m below mean sea level (= m.s.l.) (Fig. 17). It has been picked on an unconformity between the onlapping reflectors of Seismic Unit S2 and top-lapping reflector of S3. This unconformity is visible in the whole dataset and is affected by signal blanking in its deepest portion.

Horizon B data have been interpolated to derive a DEM, that shows a complex concave up surface marked by a sinuous E-W elongated topographic low (Fig. 18). Horizon B can be interpreted as an erosional basin-wide unconformity.

Figure 18. DEM of the surface corresponding to the Horizon B. Note the presence of a sinuous area of signal loss in the deepest part of the surface. Position of cores C5 and C24 (Fig. 19) is also reported.

Seismic Unit S3

The S3 is the uppermost seismic unit shows a lenticular geometry with thickness up to 40 m. It is bounded by the Horizon B at the bottom and by the seabed on top. It is a layered seismic unit with few high amplitude sub-horizontal, slightly concave up reflectors ranging from low to high amplitude, with onlap terminations on Horizon B. It constitutes the infilling of Horizon B surface. The extremely high penetration of the seismic signal suggests that S3 is made up of soft muds. The upper part of S3 is marked by a reflector at a depth of 5-16m b.s.b. with very high amplitude and continuity, named Horizon hb6. This reflector, parallel to the sea bed, is interpreted as lower limit of more recent, coarser and less consistent sediments. In the lower part of S3, the main reflector is Horizon hb3.t, visible at an average depth of 18-19 m. Horizon hb3.t is a high amplitude, continuous reflector. It is responsible for most of the signal masking effects occurring in the central part of the basin, which are diagnostic of sediments with a high fluid content.
Sedimentary Units and datings

On the basis of lithological features, biological content and stratigraphic position, two main stratigraphic units have been distinguished in cores: ASP and H. They correspond to S2 and S3 seismic units (Tab. 2).

Table 2. Correlation of seismic units with sedimentary units.

<table>
<thead>
<tr>
<th>Seismic Unit - Horizon</th>
<th>Sedimentary Unit</th>
<th>Lithofacies</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seismic Unit S3 H</td>
<td>ASP</td>
<td>structureless very soft silts with abundant organic content and bioclasts</td>
<td></td>
</tr>
<tr>
<td>Seismic Unit S3 H</td>
<td>H</td>
<td>structureless very soft silts with abundant organic content and bioclasts</td>
<td></td>
</tr>
<tr>
<td>Seismic Unit S2</td>
<td>ASP</td>
<td>alternation of very soft silts with abundant organic content and bioclasts</td>
<td></td>
</tr>
</tbody>
</table>

Unit ASP

It consists of greenish marly silty-clay with millimetric lamination, occasionally with organic matter. Colour varies from pale olive to greenish grey with brownish (locally reddish) mottles related with iron-oxides. Oxidised sediments are located just below the unconformable surface between ASP and H units. This lithofacies is extremely stiff; pocket penetrometer values range from 7.5 Kg/cm$^2$ in the lower part and 4.5 Kg/cm$^2$ in the upper part close to the unconformable surface. Bioclastic content consists in very small shell fragments of marine fauna (<2mm) and abundant calcareous nannofossils and foraminifera (mainly Globigerina bulloides, Globorotalia inflata and Globigerinoides ruber). ASP has been observed in core C24 starting from a depth of 22 m b.s.b. and in core C5 from 42 m b.s.b. (Fig. 11). ASP unit shows the lithostratigraphic characters of argille subappennine informal unit and can be easily correlated with the Blue Clay unit outcropping at the base of the cliffs surrounding the Mar Piccolo (Figg. 11D-12).

Unit H

It consists of a succession starting from the bottom of three lithofacies (Ha, Hp, Hb in Fig. 19).

Figure 19. Sedimentary units of core C5 ad C24; a - articulated bivalves, b - disarticulated valves; c - indetermined marine mollusc fragments; d - Cerastoderma glaucum; e - Mytilus spp.; f - Ostreidae.; g - Polychaeta spp.; h - pulmonate gastropods fragments; i - pulmonate gastropods; l - Cladocora caespitosa fragments; m - carbonatic nodules; n - Iron oxides/hydroxides; o - organic matter.

The lower lithofacies, Ha abruptly truncates the underlying Unit ASP. It consists of coarse yellowish-brownish beds and firm sandy mud intercalations of pale olive colour. Coarser beds are made up of poorly sorted sand and gravel with abundant bioclastic debris containing a mélange of reworked remains of marine bivalves, marine and pulmonate gastropods, C. caespitosa fragments. Muddy levels contain some sand laminations, bioclast fragments, diagenetic carbonatic nodules, reddish mottles due to iron oxides, and occasionally dark spots of organic matter. In core C5, Ha begins with coarse poorly sorted gravel deposits, consisting in decimetric calcareous pebbles and bioclastic debris. Gradually, this basal gravel interval passes upward to alternation of...
coarse sand and silt with rare clasts of underlying ASP unit. Pocket penetrometer values measured on silt levels are around 2 Kg/cm². This lithofacies is not observed in core C24, while in C5 it has a thickness of 13 m. Ha can be interpreted as the result of deposition in subaerial settings. Fluvial incision and alluvial processes seem to be responsible for the high energy facies of this unit. Coarser grained transported material comes from the erosion of the older outcropping units: C. caespitosa fragments (from MIS 5e terraced deposits), argille subappennine clasts, limestone and calcarenitic pebbles from the Calcare di Altamura Fm and Calcarenite di Gravina Fm respectively. Ha gradually passes upward to Hp (core C5, Fig. 19). In the core C24 Hp unconformably lies on the ASP Unit, at a depth of 22 m and is marked by a sharp change in colour and stiffness of the sediments. Generally, Hp is a 4-5 m thick interval of grey clayey silts intercalated with peat beds. Grey silts are very soft, organic rich and marked by the presence of well-preserved specimens of C. glaucum. Peat beds are made up of remains of plants (mainly Phragmites australis), abundant and well-preserved ostracods and shells of pulmonate gastropods (Helix sp., Ruminia sp., Pomatias sp. and the dominant taxon Planorbis sp.), few reworked fragments of marine bivalves and gastropods, and foraminifera. Two samples of the peat coming from a depth of 26 m below seabed, have been radiocarbon dated, yielding a conventional age of about 10.5 ka (radiocarbon ages: 9188 ± 75 and 9255 ± 75). The absence of carbonate concretions and oxides in Hp lithofacies marks the transition from oxidizing to reducing conditions and is indicative of a paralic environment. Alternations of peat levels with Planorbis sp. and silts with C. glaucum represent small oscillations between low energy swamp environment to lagoonal environment. Hp gradually passes to the upper lithofacies, Hb. This consists of structure less very soft grey silts with abundant organic content and bioclasts. Hb contains well-preserved, broken shells and millimetric fragments of mainly marine gastropods and bivalves. The lower part of Hb is made up of homogenous clayey silts with abundant organic content and bioclasts. Hb contains well-preserved, broken shells and millimetric fragments of mainly marine gastropods and bivalves. The lower part of Hb is made up of homogenous clayey silts with organic matter, strongly affected both by the drilling disturbance and by intense anthropic activity, such as mussel farming, anchoring, fishing and construction in the area. Hb deposited in uninterrupted semi enclosed basin with low hydrodynamicity settings, from a restricted to the present day larger semi-enclosed basin: it represents the result of transgression and continuous vertical aggradation. A whitish tepha layer has been recognized in the lithofacies Hb, at about 18 m below the sea bottom. It is 12-15 cm thick and is made up of fine to coarse ash. Lamination is not well visible but is generally marked by some thin grey silty laminae. The tepha layer stands out as a light grey layer, with a high porosity (and high water content), very low bulk density and a very soft consistency. The closest source of volcanic materials with these compositional features is the Somma-Vesuvius area where, during last 20 ka, many high-energy Plinian and sub-Plinian events occurred (Santacroce, 1987; Cioni et alii, 2008). In detail, the tepha colour, the glass composition and the stratigraphic position are compatibles with the Pomici di Mercato event (8890±90 cal yr BP, Mele et alii, 2011).

Seismic unit correlation with well cores results

Carbonatic substratum

The acoustic basement (Seismic Unit S1) can be easily attributed to late Cretaceous Calcare di Altamura Fm and Calcarenite di Gravina Fm (Upper Pliocene-Lower Pleistocene): they are compact carbonate units that crop out in the emerged adjacent areas.

Argille subappennine

Seismic Unit S2 corresponds to the Blue Clay unit sampled at the bottom of the cores (ASP Unit in Fig. 19). High values of stiffness, recorded with pocket penetrometer, result in medium amplitude reflectors in spinner record.

Erosional surface

Core analysis results allow to correlate this surface with the abrupt contact between sedimentary Unit ASP to Unit H: in core C5, argille subappennine erosional limit is marked by the alluvial gravel of Ha lithofacies (42 m b.s.b.); in sub-bottom profiles these coarse-grained deposits locally mask the basal erosional surface.

In core C24 it coincides with the contact of Hp coastal swamp-lagoonal sediments on argille subappennine (22 m b.s.b.), that corresponds in sub-bottom profiles to the well visible unconformity between toplapping S2 reflectors and onlapping S3 reflectors (Horizon B, Fig. 17). Therefore, the depth of the erosional surface is consistent both in seismic data and core data. Furthermore, the interpolated surface (Digital Elevation Model in Fig. 18) shows clearly a sinuous and ribbon shape course, that is in agreement with the subaerial hydrographic pattern. On the other hand, the deeper part of this complex surface is covered by alluvial gravel deposits, that are visible in core C5: all data reveal a deep fluvial influence in the general morphology of this unconformity. This incision can be considered the intermediate sector of a bigger fluvial system that includes landward present day Canale D’Aiedda stream and seaward the elongated depression visible in Mar Grande (Mastronuzzi and Sansò, 1998).
Late Pleistocene - Holocene sequence

Seismic Unit S3 is associated with the whole Unit H. This unit represents the sedimentary infill of the erosional surface. It consists of lithofacies Ha subaerial/fluvial deposits, of Hp lagoonal muds and of Hb shallow marine muds. More in detail we can correlate signal attenuation areas, signal loss and reflectors hb3.t and hb6 to specific sedimentary features recognized in the cores:

- signal attenuation, that is recognizable close to the erosional surface, seems to be related with the presence of the coarser and stiffer lithofacies Ha.
- loss of signal below Reflector hb3.t perfectly matches with the depth of the tephra layer (Ht). The exceptionally high porosity of the tephra is responsible for fluid accumulation and can cause an almost complete reflection of the acoustic signal. The sinuous area of blanking in the erosional surface DEM may be due to thicker tephra deposits in the sinuous and elongated basin depocenter (Fig. 18).

The tephra layer shows unexpected thickness that might be due to redeposition, furthermore it is prone to fluid accumulation due to its porosity. Water or biogenic gas might accumulate in the bathymetrically lower part of the basin. Fluid gas charges sediments are common in the shallow stratigraphic record (Judd and Hovland, 1992) and is also verified that tephra layer in marine settings can have a high hydraulic conductivity (Harders et alii, 2010).

Morphological evolution of the Mar Piccolo

The whole data can be considered evidence of the presence of an incised valley (or paleovalley sensu Blum et alii, 2013) shaped on the Blue Clay unit. The valley is filled with a sequence (sedimentary unit H) that records a landward migration of the lithofacies related one of the most recent episode of transgression. Preliminarily we can assume that the erosional surface incision and subsequent infill (Unit H) are post MIS 5e in age for the occurrence of abundant remains of C. caespitosa found in Ha unit, indicative of the cannibalization of Tyrrhenian reefs. The age of the infill is mostly Holocene and it has been established on the basis of the following two main time constraints:

- peat level in core C5, yielding an age of about 10.3 ky;
- tephra layer identified as Pomici di Mercato eruption (8890±89 cal yr BP).

The two time-constraints have been compared with the Lambeck et alii (2004; 2011) sealevel curve available also for Taranto area (Fig. 20) that is a recognised tool for the study of paleo sea level. In detail, we consider the peat level as a good sea level marker, in fact it represents a coastal environment and in cores, is immediately preceded and followed by levels with C. glaucum a Lamellibranchia that is considered living in Mediterranean lagoon up to a depth of about -2 m (Lambeck et alii, 2004; Ferranti et alii, 2006). The dated peat level accumulated in a coastal swamp close to sea level, in accordance with the position of the peat in the graph (Fig. 20), above Lambeck et alii (2004; 2011) sea level curves.

![Figure 20. Time constraints compared with the Lambeck sea level curves for Taranto area; tephra (in pink), below Lambeck et alii (2004; 2011) curves, is compatible with deposition in a very shallow marine environment, the position of peat levels (orange and blue) above the two curves is in accordance with the deposition in a transitional environment.](image-url)

We also used Lambeck et alii (2004; 2011) sea level curves, to infer the paleowater depth during Pomici di Mercato eruption (8900 – 9000 yr BP). At this time the predicted sea level from the curve is -25 m m.s.l. (Fig. 20), while the top of the tephra layer has been found at an average depth of -32.4 m m.s.l (19 m b.s.b.); this suggests that the volcanic ashes deposited in a very shallow basin (about 6-7 m deep). According to this hypothesis the top of the tephra layer can provide a good approximation of the paleo-bathymetry at the time of deposition. The age of the lower time constraint (peat level) of Mar Piccolo paleo-valley infill, testifies that fluvial (and sub-aerial) erosion in Mar Piccolo paleo-valley system culminates during the Last Glacial Maximum, similarly to many paleo-valleys in the Mediterranean and French Atlantic coast (Amorosi et alii,
2016; Maselli and Trincardi 2013, Chaumillon et alii, 2010; Billeaud et alii, 2005). Nevertheless, the erosional unconformity mapped and reconstructed, could not be the result of just fluvial processes. As pointed out by Strong and Paola (2008), incised valleys evolve continuously throughout both sea-level fall and rise, producing erosional surfaces that are highly diachronous and amalgamated; the basal erosional unconformity forms over most of the duration of the sea level cycle and does not represent a topographic surface. Furthermore, according to Cattaneo and Steel (2003), when sedimentary record corresponding to relative sea level lowstand is not preserved, at the base of the transgression there might be a complex polygenetic surface originated from subaerial erosion during times of relative sea level lowstand (a sequence boundary) and subsequent reworking during the ensuing transgression. In the case of Mar Piccolo, although a clear fluvial origin can be detected from the erosional surface mapped, the present day its subcircular shape cannot be explained only as the result of fluvial incision. In the Mar Piccolo, despite low energy hydrodynamic conditions, wave and tide erosion strongly affect the cliffs due to their highly erodibility lithology (Fig. 11D). We believe that processes of shoreface retreat due to wave and tide ravinement might have occurred during the transgression, when Mar Piccolo fluvial incision was flooded forming a ria (sensu Evans and Pregu, 2003 and references therein). The same processes probably culminated during the slowing down of sea level rise (from 8-7 ka), causing the widening of the erosional surface (Fig. 21).

Morphological evolution of the Taranto area

The chronostratigraphic data, along with the geomorphological survey of the entire area of Taranto, permitted a reconstruction of the morphological evolution of the Taranto coastal area during the Last Interglacial/Last Glacial cycle. During MIS 5e this area was affected by a rapid, long and wide transgression that shaped a large bay which corresponds to the present day Mar Grande and Mar Piccolo. Regarding the present day flat seabed of the Mar Piccolo hides a basin-wide erosional unconformity that is buried by a thick cover of Late Pleistocene - Holocene sediments. All data show that the Mar Piccolo evolution can be interpreted as result of the erosional and depositional processes in transitional low energy settings of an incised-valley system during last sea level cycle (Valenzano et alii, 2018). Through the features of a ria, the present day circular perimeter of the two embayments of the Mar Piccolo is the result of the recent and present day continuous landward retreat of the highly erodible cliffs. Taking into account the temporal constrains obtained with datings deriving from marine terraced deposits and Holocene sequence, the evolution of the area of Taranto can be traced as follow:

- MIS 5e – maximum transgression connected to the warming of the planet and to the tropicalization of the Mediterranean Basin; sea level stands at about +7 m above the present one, shaping on the local basement, here represented by the Blue Clay unit, an incredible large and flat marine terrace characterised by deposition of coastal marine and transitional sediment locally rich in Senegalese fauna, today preserved at an elevation ranging from 23 and 4 m above present sea level;
- Post MIS 5e – beginning of fluvial incision of the marine terraced deposits ascribed to the MIS 5e and of the underlying Blue Clay unit;
  - MIS 2 – Last Glacial Maximum (LGM); deposition of alluvial high energy facies (Ha);
  - ~10ka - beginning of marine ingestion in the Mar Piccolo area and development of a ria; deposition of Hp facies in freshwater coastal swamp environment; rapid sea level rise, increase of sediment accommodation and deposition of a thick semi-enclosed low hydrodynamic succession (Hb).

- 8.9 ka BP - deposition of tephra in semi-enclosed settings;
- about 7 ka/present day – slowing-down of the sea level rise, erosion of the coastal portions, cliff retreat, widening of the erosional unconformity and contemporaneous deposition in low energy and shallow water setting (from a lagoon to the larger present-day semi-enclosed inlet).

Figure 21. Sketch of the morpho-sedimentary evolution of the Mar Piccolo basin. 1. Fluvial incision and deposition of alluvial/continental deposits of Ha lithofacies; 2. deposition of Hp paralic facies and beginning of marine incursion; 3. Deposition of Hb lithofacies during transgression; 4. basin widening and shoreface retreat during sea level still stand.

Crossing Southern Italy: a travelling meeting from Taranto to Siracusa
STOP 3 - The flight of marine terraces bordering the Gulf of Taranto (Lat. 40°13’01”N – Long. 16°40’12”E)

Introduction

This flight of marine terraces is one of the most impressive examples in the Mediterranean for the interplay between tectonic uplift and the glacio-eustatic sea level fluctuations of the Middle and Late Quaternary. Like a huge Greek-Roman theatre, the eleven-stepped staircase rises up to 392 m (T10 at Pisticci) and 604 m a.s.l. (T11 at Ponteradio), respectively (Fig. 22) were studied by Brückner in his PhD thesis (Brückner, 1980a). He continued and supplemented the diligent mapping of Boenzi et alii (1976), developed a model of the terrace genesis, and suggested a chronostratigraphy (see below).

Figure 22. The marine terraces of the Metapontino. Like a gigantic ancient theatre this staircase with 11 steps borders the Gulf of Taranto. T(0): Holocene, T1 – T11: Pleistocene; Brunhes/Matuyama boundary, 780 ka BP: between T10 and T11. The evident gap between T11 and T10 is due to the severe erosion of the Lucanian Hill Country, where the caprock, i.e. the cemented pebbles and the sands of the marine terraces, were eroded and spectacular badlands evolved (bedrock: Calabriano clays). Normally, the marine terraces are formed as depositional terraces, composed of sands and pebbles; however, in the calcareous tufa of the Murge (from Matera via Castellaneta to Statte) they are formed as erosion terraces. The insert map near the legend shows the continuation of the marine terraces towards the south (extension of the left part of the map) (Brückner, 1980a).

The tectonic setting is such that the Bradanic Trough, i.e., the southern sector of the Apennine Foredeep, had been filled since the Early-Middle Pliocene with 3-4 km thick sediments (Tropeano et alii, 2002). In the course of the general uplift of the Southern Apennines, the area became subject to uplift as well with values decreasing from SW (in proximity of the Apennines) to NE. One can consider the uppermost marine terrace T11 with the so-called Irsina Conglomerate as representing the first major regression facies.

The then following terraces reflect the general seaward shift in the shoreline summing up to ca. 25 km. This general regression trend interrupted during transgressions. The latter were not in phase with the interglacial periods and an offset of half a period was recognized. In an ideal case, a transgression starts after the marine lowstand during the glacial maximum and ends with the marine highstand during the interglacial maximum (cf. Fig. 27). Transgressions formed active cliffs by which they partly eroded the formerly deposited terrace sediments. The most landward cliff of such a cycle later turned into a terrace edge (cf. Fig. 23).
Standard profile of the marine terraces

The lowermost part of each marine sedimentary sequence is characterized by a distinct disconformity indicating the marine transgression developed on the Argille Subappennine, (i.e. the Blue Clay unit) of the Bradanic Trough (Brückner 1980a, 1983). The basal part of the marine strata mainly consists of homogeneous sand with intercalated gravel horizons (lithofacies of the transgressive basal conglomerate sensu Carobene, 2003). This terrace base continues upwards with generally coarser sediments of the main gravel layer formed in a littoral environment during sea level highstands or the beginning of the regression phase. It is usually covered by or interbedded with shallow marine to lagoonal, fluvial, and sometimes even aeolian deposits (Fig. 24).

At a given site, 8 the Petrulla outcrop in T1 (Fig. 25; location: yellow dot on T1 in Fig. 30), the sediment accumulation cycle starts with beach deposits (sands, pebbles) when the transgression reaches the site. During the following submergence, sublittoral strata are deposited (silts, sands, lenses of pebbles). The accumulation terminates with a second beach facies formed during the subsequent regression. The terrace top is build-up by lagoonal or terrestrial strata, the latter may be of alluvial or aeolian origin. The cliffs are formed during the transgression periods, the terrace surfaces during the regression periods and thereafter (Figg. 26-29, see also Brückner, 1983).

Figure 24. Generalised stratigraphy of the marine terraces of the Gulf of Taranto (Brückner, 1980a, modified). The profile shows a typical vertical sediment sequence.
Cross sections through a marine terrace

It is interesting to study the formation of the marine terraces in shore-perpendicular and shore-parallel cross sections. The former (Fig. 28) shows the composition of the marine terrace T3 with the transition from the upper terrace base (foreshore environment) to the main gravel layer (beach environment). The evolution of the terrace top was multiphased in that it shows a palaeosol (red loam) buried by younger sediments.

The shore-parallel cross section (Fig. 29) through T5 was displayed when a part of the Apulian water pipeline (Aquedotto pugliese) was constructed in the late 1970s. It shows that the main gravel layer is much thicker towards the SSW (left side of the Fiume valley) than towards the NNW (right side of the Valle valley). The reason was the strong longshore drift: After the pebbles had been delivered by the rivers, they were transported by the coastal current towards the NNE. This trend is morphologically underlined by delayed river mouths. The formation of the fluvial terraces was controlled by sea-level highstands during transgression phases.

Crossing Southern Italy: a travelling meeting from Taranto to Siracusa
Age estimates of the marine terraces

In 1980 Brückner presented the following chronology for the terrace staircase: (i) The Brunhes/Matuyama boundary is between T11 and T10; (ii) a volcanic ash layer in the cover sediments of T8 was identified by A. Rittmann, Catania, as the 500-600 ka-tephra from the Phlegrean Fields; based on 230Th/234U ages, the lowermost terrace T1 was attributed to MIS 5a. On the background of these temporal anchor points, the complete flight of terraces was wiggle-matched with the oxygen isotope record (as it was known in the 1970s), under the assumption that the transgression peaks, which created the former cliffs, *grosso modo* represented the peaks of interglacial periods. It was the first time that a correlation was suggested between a flight of marine terraces and the OIS record.

Later, more ages estimates were published: The identification of Senegalese fauna in some places, e.g. Ponte del Re (Boenzi *et alii*, 1985), provided evidence for the MIS 5e terrace. Amino acid racemization (AAR) chronology enabled the distinction of MIS 5 sub-stages at several sites (Hearty and Dai Pra, 1985; 1992). Zander *et alii* (2003, 2006) carried out OSL dating with some success; however, most of the terraces in the area were out...
of the OSL dating range (according to the then known luminescence protocols). An additional U-series age attributed T4 to MIS 7 (P. van Calsteren, in Sauer et alii, 2013). However, molluscs tend to behave as open systems, which makes U-series ages questionable. In general, correlation between sites is often difficult, because the number and elevations of the different terraces vary from SW to NE. Thus, the definite attribution of the terraces to the MIS record still bears open questions. The chronostatigraphy presented in Tab. 3 may, be slightly modified in the future. It provided a solid base for the soil chronosequence study (see below). With the ages and altitudes noted in Tab. 3, an average uplift rate of 0.5 m/ka during the Brunhes epoch can be calculated for the Metaponto-Pisticci area.


1Maximum elevation is missing because the upper terrace edge TK 9 was eroded.
2Minimum and maximum elevations are missing because the terraces edges were eroded.
3The terraces fell dry between the time of sea-level maximum and the end of the respective MIS. The mean values of these two boundary data were hence used as assumed soil ages (exceptions: age estimates of the Holocene terrace are based on OSL ages and the relative sea-level highstand during the mid-Holocene).
4Zander et alii (2003, 2006);
5Brückner (1980a);

<table>
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<th>Terrace</th>
<th>T0</th>
<th>T1</th>
<th>T2</th>
<th>T3</th>
<th>T4</th>
<th>T5</th>
<th>T6</th>
<th>T7</th>
<th>T8</th>
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<td>190-215</td>
<td>225-240</td>
<td>255-284</td>
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<td>Chronological evidence</td>
<td>OSL4</td>
<td>Mid-Holocene sea-level highstand</td>
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<td>Strombus terebra, embedded in top layer3</td>
<td>Sediment different from T8</td>
<td>T9 – T10: Brunhes, from T11 to Matuyama2</td>
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<td>Soil colours</td>
<td>T0: greyish to yellowish-brown soils</td>
<td>T1-T2: dark brown soils</td>
<td>T3-T5: reddish soils (chromic)</td>
<td>T6-T10: reddish and very red soils (chromic and rhodic)</td>
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### Soils developed on the marine terraces

The study presented in this chapter is mainly based on Sauer et alii (2010, 2013) who investigated the soils developed on the marine terraces from Metaponto (T0) to Pisticci (T10) between the valleys of Cavone and Basento (Figg. 30-31). This contributes to the understanding of the mid-Pleistocene to Holocene landscape history from a pedogenic perspective.

### Problems with soil evolution of old terraces

In the Metaponto area, the identification of progressive terrace ages based on soil development stages faces three main problems:

(a) The terrace sequence comprises a time span of about 730 ka. Several previous soil chronosequence studies have obtained logarithmic chronofunctions of most soil properties, showing high rates of soil formation during the first ~100 ka and considerably lower rates thereafter (e.g., Bockheim, 1980; Harden, 1982; Alonso et allii, 1994). This makes the use of soil development indicators for age estimations of old land surfaces difficult.

(b) Land uplift has caused erosion, so that the upper, most strongly developed part of the soils, which would exhibit the maximum Fe2+/Fe3+ ratio and minimum silt/clay ratio, is often missing and therefore not reflected in the analytical data. Erosion of the clay-illuviated soils is indicated, e.g., by the presence of clay cutans at shallow depth: In some cases, clay content maxima start already in the Ap horizon, indicating that the E horizon has been completely removed, and the Ap horizon has formed within a former Bt horizon.

(c) Where younger deposits buried the marine sediments the soil development stage in the marine sediments does not necessarily correspond to the age of the terrace.
Pedogenetic changes in connection with terrace age

REDNESS: The soils developed on the lowermost terraces T0 and T1 are weakly to moderately developed and have greyish or yellowish-brown colours, while those of T2 and T3 are more reddish. Some of the soils on the higher terraces, especially the ones on T7, are deeply developed and intensively red coloured (cf. Tab. 3). However, the trend in redness is not perfect, because the sedimentological situation differs among the sites, and the sediments forming the land surface are often younger than the respective marine terrace.

CLAY MINERALOGY: Clay mineral composition does not reflect progressive soil development, because the fresh sediments (C horizons) already contain very variable amounts of smectite, vermiculite and kaolinite, so that clay minerals cannot be used as age indicators.

SOLUM THICKNESS: Over long periods of time this criterion is complicated by erosion and colluviation processes, esp. in Mediterranean landscapes which have been intensively influenced by humans over millennia. Moreover, addition of aeolian sediments to soils is a widespread phenomenon in the Mediterranean, leading to thickening of the soils and to rejuvenation in terms of soil weathering indices. Solum thickness is also a difficult criterion in cases where the terrace soils have been buried by alluvial or colluvial deposits. Anyhow, the analyses suggest that solum thickness *grosso modo* increased from T0 to T11.

SOIL EVOLUTION IN GENERAL: Soils on terraces T6 and T7 are clearly more developed in terms of solum thickness, rubefication, alteration of terrace gravel and Fe/Fe ratio than soils on terrace T4. Thus, the hypothesis of Bentivenga *et alii* (2004) who considered all of the terraces to be of the same age, though displaced by tectonics, is definitely falsified.

WEATHERING INDICES: Soil chronofunctions may add information to age estimates of terraces. Since a major process of chemical weathering of silicates is hydrolysis, i.e. exchange of Na+, K+, Ca²⁺ and Mg²⁺ in silicates for protons, a base leaching index (BLI) can be applied: \[ BLI = \frac{(Ca+Mg+K+Na)}{Al} \]. In the Metaponto area, silicate weathering and related base leaching with time can best be described by the power function: \[ BLI = 0.9724 \times \text{terrace age}^{0.184} \] (\( R^2 = 0.95 \)). Another weathering index is pedogenic iron (Fe₉/Fe₄). Based on these two
soil chronofunctions (the fit of the T6 and T7 soil data with the BLI curve and the Fe$_{\text{d}}$/Fe$_{\text{a}}$ ratios), the terraces T6 and T7 can be attributed to MIS 11 and 13, respectively (cf. Tab. 3).

POLYGENETIC SOIL EVOLUTION: Studying the soil chronosequence, on the one hand, soil properties indicating progressive soil development with time were identified. On the other hand, it is obvious that soil development was not uniform but underwent a number of glacial/interglacial cycles, whereas all soils (except for those on the Holocene terrace) are polygenetic. The soils did not only experience different kinds and rates of progressive pedogenesis, but also “regressive” processes: During the studied time span of 730,000 years, significant geomorphological processes such as erosion, colluviation, in places also loess accumulation, played an important role.

**Policoro – A short history of the ancient sites of Siris, Polieion and Heracleia**

On the Castle Hill of Policoro, in the valley of the Varatizzo and on the southern plateau, towards the Sinni valley, the archaeologists discovered the traces of several ancient cities, one on top of the other. The most ancient remains are a mixture of Greek (Aegean) and local (Italic) ceramics, going back to the beginning of the 7th century BC. Called the Chonian Siris by the Greek geographer Strabo, this was a Greco-Indigenous centre of production and exchange. It took advantage of the fertility of the soils, and of the connectivity between the coast and some mountainous indigenous sites (like Anglona) through the valleys of the Sinni (ancient Siris, the river which gave the name to the whole region), Agri (ancient Akryis) and Cavone (ancient Acalandrus).

Shortly before the mid 7th century BC, an “Ionian” fortification was built, together with sanctuaries and an urban frame: this corresponds to the foundation of Polieion, a colony of Colophon. The name comes from the goddess Athena Polias (“of the city”). After a period of prosperity, the neighbour cities of Metaponto, Sybaris and Croton destroyed Siris-Polieion.

Survivors and immigrants (maybe from Sybaris, destroyed in 510 BC, and from Metaponto) continued to live on the territory of Siris. Thus, the third city from which archaeologists discovered traces of urban habitation and of a temple dates back to 510-480 BC; the coins representing the river-god Siris with the inscription SIRINOS probably correspond to this short-time foundation.

However, already in 480 BC, before the naval battle of Salamis (won by the Athenians against the Persians), the Athenian chief Themistocles referred to Siris as a land which should belong to the Athenians. The information is transmitted by Herodotus, the “Father of History”, who settled in Thurioi, a Panhellenic foundation under the direction of Athens on the ancient site of Sybaris. Indeed, ca. 435-433 BC, after a war for influence over the region, Thurioi and Taras planned to re-found Siris. Finally, Taras alone accomplished the project in 433/432 BC; the city was called Heracleia, after the name of the Dorian hero Heracles.

Centre of the Italic League, conquered by the Lucanians in ca. 338 BC, Heracleia was the theatre of some of the most terrible wars in the ancient history of Italy: during the invasions of Alexander the Molossian, Pyrrhus of Epirus, Hannibal of Carthage, the revolt of Spartacus. After the Social War, people in Italy got Roman citizenship, and the territory of Heracleia benefited from the Pax Romana until the end of Antiquity.

The ancient authors do not agree upon the identification of these cities. In particular, Strabo claims that Heracleia occupied a different site from that of Siris, which would have been Heracleia’s harbour (Greek epineion) close to the Siris river. The archaeological evidence on the Castle hill, however, contradicts him: this hill has been the centre of all the ancient cities of this region. Yet, we still have to find the harbour, in order to understand how Strabo could have made this mistake. Moreover, the study of the littoral sandbars separating the hill from the mouth of the Sinni can help us to reconstruct the rural territory (Greek chorai) of the city, renowned for its fertility and land division (known from the bronze inscription of the “Heracleian Tables”).

Finally, in Medieval times the settlement was called Policoro, meaning “big village”.

Recently, research has been carried out in the framework of ongoing archaeological and geoastronomical projects under the responsibility of Stéphane Verger, Francesca Sogliani, Rossella Pace, Anca Dan, Helmut Brückner & collaborators (Chora: [http://www.archeo.ens.fr/spip.php?article1066]; Legecartas: [http://www.archeo.ens.fr/spip.php?article1685]).

**Crossing Southern Italy: a travelling meeting from Taranto to Siracusa**

40
STOP 4 - The enigmatic tidal notch of Roseto Capo Spulico (Lat. 39°58’52”N – Long. 16°37’02”E)

Along the Ionian Sea coast of Northern Calabria, uplift of Middle Pleistocene and younger marine terraces occurred at average rates of ~1 mm/yr (Cucci and Cinti, 1998; Santoro et alii, 2009), and is attributed to a combination of regional, deep-seated deformation and upper-crustal, active transpression (Ferranti et alii, 2009; Santoro et alii, 2013). The extent of uplift is imaged by the pattern of the Tyrrenian or MIS 5.5 (Ferranti et alii, 2006) terrace, which reveals that this region lies at the northern border of the Calabrian Arc uplifting province.

Although the Tyrrenian terrace provides estimate of a 1 mm/yr rate close to the long-term average, detailed mapping of the last 400 ka terrace flight in the region reveals that uplift was not constant, and during the last 50 ka it was 1.7mm/yr on average (Santoro et alii, 2009) Fig. 32.

Figure 32. Uplift rate history for the northeastern Calabria coast during the last 400 ka based on the established chronology of uplifted paleo shorelines (Santoro et alii,2009). Uplifts are non steady-state, and the rate during the last 50 ka was 1.7 mm/yr on average.

The Holocene notch at Roseto Ferranti and Antonioli (2009) reported the first evidence for Holocene uplift at this coast, represented by a wide notch carved in a carbonate intercalation within argillite basinal rocks (Sicilide Unit) at Roseto Capo Spulico, a promontory projecting in the Ionian Sea along the Gulf of Taranto shoreline. The mushroom shaped limestone block that hosts the notch is found few m off the present coastline just underneath the Roseto castle. The notch has a well preserved roof and is bored by a band of holes up to 2.25 m a.s.l., an elevation broadly coinciding with the maximum notch concavity. Underneath this concavity, a minor horizontal smoothed notch is found at 1.05m. No material for dating was retrieved in the holes. However, a broad constrain on the notch age was indirectly placed by Ferranti and Antonioli (2009) by considering the uplift pattern established at this coast by Santoro et alii (2009), who estimate a rate of 1.1-2 mm/yr since the last 125 ka, with a recent (40 ka) value close to the upper bound (1.7 mm/yr on average). Based on the predicted Lambeck sea-level rise curve for this coast and the inferred uplift rate of 1.7 mm/yr estimated at this coast, Ferranti and Antonioli (2009) argued that the notch has a ~2.5 ka best fitting age (total uplift of 4.2m). Later on, Santoro et alii (2013) provided an updated estimate of the long-term uplift rate, and attributed the lowermost terrace at 8 m elevation to the mid Holocene (6 ka), with an inferred uplift rate of 2.2 mm/yr. By using the refined uplift rate estimate, the notch would have a younger age of ~1.7 ka. The present uplift of the notch can be attributed to a combination of the regional uplift and of activity of the transpressional faults documented in the area (Ferranti et alii, 2009; Santoro et alii, 2013).

Figure 33. The mushroom shaped limestone block in the 2006 (a-b) and in 2018 (c).
STOP 5 - The archaeological site of Sybaris (Calabria, Southern Italy) and the sea level change (Lat. 39°43'05"N – Long. 16°29'28"E)

The archaeological site of Sybaris is an important case history of the interplay among geology, tectonics, relative sea level changes (RSL), geotechnics and archaeology given the peculiar geological set up of the area, where important subsidence processes have involved the archaeological sites, initially lying on and later submerged by alluvial deposits. Processes like the eustatic rise, regional and local tectonic displacements, sedimentary accumulation and related consolidation, fluid withdrawal, due to human activity, closely interact and it is hard in discerning their relative contribution to changes in shoreline position. The archaic site of Sybaris, once a flourishing Greek colony, lies in the North of Calabria, within the alluvial plain of the same name. The archaeological finds, concentrated in some areas, Parco del Cavallo, Stomi and Casa Bianca, along the actual course of the Crati River, show three superimposed levels of continuous habitation since VIII-VII century BC: the Greek town of Sybaris (720-510 BC), the Hellenistic Thurium (444-285 BC) and the Roman Copia (193 BC), currently at a depth from 7 to 3.5 m below ground level (Fig. 34).

In order to investigate the geological and geotechnical processes generating the burial of the archaeological site, researches have been carried in the last 20 years at the Research Institute for Geo Hydrological Protection (IRPI-Bari) the Technical University of Bari and the City University of London within the Strategic Project for Cultural Heritage (Italian CNR), by means of more than 20 boreholes drilled with a variable depth from 10 to 120 meters from the actual ground level. (Cotecchia et alii, 1994; Pagliarulo et alii, 1995; Coop and Cotecchia, 1997; Cherubini et alii, 2000; Pagliarulo and Cotecchia F, 2000; Pagliarulo, 2006).

Figure 34. The location of the archaeological areas, the geomorphological sketch, the Roman Theatre, the three superimposed old towns and the “Sigillo di Frequenzazione” where the three superimposed towns coexist.

The thickness of the alluvial sediments filling up the plain increases from the North to the Southwest, (from 103 to 478 m respectively to the South of the Crati River). They consist of sands, silty sands, clays, sandy clays and gravels in heteropic facies, also interbedding peat levels at places, suggesting the presence of swamps along the river floodplain, at time of deposition, with a different distribution of water (Fig. 35). Numerous samples were subjected to $^{14}$C dating (Tab. 4). Deep borehole analysis supplied a longer (Latest Pleistocene-Holocene) subsidence record, and led to the conclusion that sinking of Sybaris resulted from the interplay of tectonics, eustasy and sediment constipation). Average subsidence at 11 ka BP occurred at ~5.4 mm/yr, but decreased upward to ~1.6 mm/yr (Cherubini et alii, 2000; Pagliarulo, 2006). This place, however represents an anomaly in the regional tectonic frame that is characterized by strong uplift in northeastern Calabria around Sybaris since the Middle Pleistocene (Ferranti et alii, 2009; 2010).

Large uplift is well testified by terrace flights surrounding both the plain and the range flanks, although at locally variable rates. While Pollino and Sila mountains are structural culminations, the Sibari plain represents a local structural depression in the general uplift contest (Santoro et alii, 2009). Moreover, integration of marine and on-land data broadly locates the trace of major anticline and syncline showing that the Sibari off shore basin represents a deep trough between two structural highs to the North and the South. Lithofacies analysis of the well logs showed that during the Early Holocene, whereas the Parco del Cavallo site was characterized by a sheltered
littoral environment where compressible sediments were deposited, an alluvial environment existed to the NW around Stombi (Pagliarulo et alii, 1995).

Figure 35. Log of the deep borehole showing position and age of dated samples. Comparison of observed data and the predicted sea level for Sybaris. RLS history for Latest Pleistocene-Holocene from boreholes S1, S15, S16, S18. Data points are in Table 1. There is difference in vertical displacement recorded by boreholes located away from (S1) or close to (S15, S18) the ancient town and the Crati River (S16) course. Borehole S1 (placed ~2 km W of the Stombi archeological site) shows little shifting from the predicted sea-level rise curve from ~11.3 to 5.4 ky BP and between depths of 60 to 5 m b.s.l. Borehole S16, drilled south of the archeological park close to the northern bank of the Crati river, shows an history of subsidence at quite high rates (Ferranti et alii, 2011).

Table 4. Vertical displacement computation for paleo sea level in boreholes at Sybaris (Ferranti et alii, 2011).

Later in the Holocene, alluvial conditions were established all over the area; an alluvial depositional surface forms the last archeologically sterile level before settlement. Well log analysis and laboratory tests identified this laterally discontinuous, highly compressible, mainly clayey layer at 35-60 m depth, which is responsible for the marked geotechnical subsidence at Parco del Cavallo (Coop and Cotecchia, 1997). Given that in recent times the rate of eustatic rise and regional uplift approximately balance, sediment consolidation has been inquired as the main cause of subsidence.
The radiocarbon datings coming from borehole S1 have been included as sea-level markers in Lambeck et alii, model (2004) that studied the Holocene sea level changes along the Italian coastline, contributing to reconstruct the specific RSL curve for the Calabria region. It suggests that the site lies in apparent stability, so any tectonic uplift must be almost completely balanced by subsidence. The Holocene data, tectonically corrected by using the displaced elevation of the MIS 5e (125 ky BP) terrace drawn from Bordoni and Valensise (1998) fitted the predicted curve at the lower bound. The RLS history for each boreholes have been referred to the predicted sea level curve (Lambeck et alii, 2011). For samples older than 14 Ky the predicted sea level values were taken from the global deep sea isotopic model of Waelbroeck et alii (2002) (Ferranti et alii, 2011; 2012). A relevant pattern of changes in rate of vertical motion during time emerges from borehole log analysis. It is possible to summarize that large subsidence occurred during Early-Middle Holocene (5-6 mm/yr) and was concentrated beneath the ancient town, then it slowed progressively during Late Holocene until a rate of 1.5 mm/yr. The slowing down trend continued during and after historical occupation at about 0.8 mm/yr. Field observations documented the large Holocene geomorphological changes occurred in the plain. The more relevant processes involved recurrent, even historical catchment and separations between the Coscile and Crati river courses (Fig. 34). The ancient town of Sybaris was located between the two rivers Crathis and Sybaris (now called Coscile) and the two courses flowed separately and bordered the settlement North and South. Today, they are joined west of it. A high coastal progradation rate (1 m/yr) has occurred in the plain since Greek (2.4 ka BP) and possibly Neolithic (7 ka BP) times, mainly deriving by the large fluvial discharge outpacing the slow eustatic rise. An archaeoseisimological field survey carried on by Cinti et alii (2015) shows oriented fractures, tilting, warping, and clockwise horizontal rotation of walls and patched up/dismantled walls affecting the residential buildings at Parco del Cavallo and Casa Bianca, most probably related to different seismic events occurred between the II-VII century AD. The effects of the earthquakes observed in the archaeological areas are generally not removed, but instead, become part of the archaeological stratigraphy. The multidisciplinary research has been integrated and completed with other determinations on samples taken at Casa Bianca (Ferranti et alii, 2011; 2012) (Tab. 5).

Table 5. Vertical displacement computation for palaeo sea level at Casa Bianca (Ferranti et alii, 2011).

<table>
<thead>
<tr>
<th>No</th>
<th>Site</th>
<th>Material – marker</th>
<th>Coordinate</th>
<th>Dig (m)</th>
<th>13C Age (ky BP)</th>
<th>Reference</th>
<th>Sea-level correction</th>
<th>Predicted Vertical displacement</th>
<th>Displacement 2011</th>
<th>Rate (mm/yr)</th>
<th>Displacement clas</th>
</tr>
</thead>
<tbody>
<tr>
<td>25</td>
<td>Coral</td>
<td>S-Y</td>
<td>1415±23</td>
<td>This paper</td>
<td>1.14</td>
<td>-0.46</td>
<td>0.68</td>
<td>1.14</td>
<td>+0.12</td>
<td>-0.32</td>
<td>+0.05</td>
</tr>
<tr>
<td>26</td>
<td>Black</td>
<td>S-Y</td>
<td>1414±16</td>
<td>This paper</td>
<td>1.14</td>
<td>-0.46</td>
<td>1.31</td>
<td>1.14</td>
<td>+0.15</td>
<td>-0.77</td>
<td>+0.05</td>
</tr>
<tr>
<td>27</td>
<td>Shell (Sea-</td>
<td>S-Y</td>
<td>1415±23</td>
<td>This paper</td>
<td>1.14</td>
<td>-0.46</td>
<td>1.31</td>
<td>1.14</td>
<td>+0.15</td>
<td>-0.77</td>
<td>+0.05</td>
</tr>
<tr>
<td>28</td>
<td>Gloss</td>
<td>S-Y</td>
<td>1414±16</td>
<td>This paper</td>
<td>1.14</td>
<td>-0.46</td>
<td>1.31</td>
<td>1.14</td>
<td>+0.15</td>
<td>-0.77</td>
<td>+0.05</td>
</tr>
</tbody>
</table>

This area is related to the Hellenistic town of Thurium, now located at approximately 2.5 Km from the sea, was originally on the coastline, it was the harbour, the towpath, called Porta Marina (= Sea Gate) (Fig. 36).

Figure 36. The archaeological site of Casa Bianca that was the harbour of the Hellenistic town Thurium.
The data coming from Casa Bianca yields evidence of broadly consistent subsidence rates which covers the whole historical lifespan of the settlement. The Long Wall confined the settlement to the North and was built on a previous coarser fortification possibly dating back the foundation of Copia (194 BC). (Fig. 37). The lower part cuts obliquely across the paved towpath. The stone basement dated back to the IV century AD, hosted the boats when they were not in use. Sea Gate was the limit with the coastline or with a backshore lagoon, joining the Long Wall with a circular tower (Fondazione Lerici, 1967). The vertical variability of hydraulic conductivity is very great due to the different grain size distribution of sediments. The shallow water table is at a depth of 0.5 m b.g.l. (below ground level) while the piezometric surface relating to the deep strata is at 2 m b.g.l. A well point system drained locally the groundwater in some of the archaeological digs to allow the touristic access. The excessive and uncontrolled withdrawal of the underground waters, since 1945, is the cause of an increased recent subsidence in the plain (Cafaro et alii, 2013).

Figure 37. The field context of the data samples at Casa Bianca. Samples 25,27,28 are related to the Long Wall that confined the archaeological site to the North, where a marshy environment was progressively developing perhaps due to the ancient Coscile river. Sample 26 is a brown palaeosoil that, West of the towpath and the Macellum directly covers the mosaic adorned ground of a building.
20th September 2018 – From Sibari to Reggio Calabria
Regional tectonic framework of the Calabrian Arc

The Calabrian Arc, including Calabria and north-eastern Sicily, represents the emerging part of a forearc emplaced above the subducted Ionian slab, which dips steeply to the NW underneath the Tyrrhenian back-arc basin (Fig. 38) (Chiarabba et alii, 2005). Although formed during subduction, the forearc has been stretched by Pliocene-Quaternary extension, and today a belt of active extensional faults are running along the Tyrrhenian Sea margin and the chain axis. The extension direction, determined by fault slip analysis (Tortorici et alii, 1995; Faccenna et alii, 2011), focal mechanisms of crustal earthquakes and global positioning system (GPS) geodetic velocities (D’Agostino and Selvaggi, 2004; Palano et alii, 2012, Serpelloni et alii, 2014), is NW-SE. Residual geodetic velocities across the normal fault belt point to extension rates of up to ~3 mm/yr (Serpelloni et alii, 2010; Devoti et alii, 2011).

Conversely, beneath the Ionian Sea, the recent and possibly active deformation pattern is related to the Ionian accretionary complex (Fig. 38) (Polonia et alii, 2011; 2012). Active contraction at the front and along thrusts splaying within the accretionary wedge from the basal detachment is suggested by analysis of seismic reflection profiles (Polonia et alii, 2011). In addition, the wedge is segmented across strike by crustal transfer tectonic systems, which represent the shallow expression of shears in the subducted plate (Fig. 38) (Polonia et alii, 2016). GPS velocities and dynamic modelling suggest up to ~5 mm/yr contraction across the Calabria forearc (D’Agostino and Selvaggi, 2004; D’Agostino et alii, 2011; Carafa et alii, 2015), which could stem from current subduction (Devoti et alii, 2008 and reference therein).

During Quaternary, the Calabrian Arc experienced vigorous uplift, as documented by widespread flights of marine terraces raised at up to several hundred meters elevation (Westaway, 1993; Miyauuchi et alii, 1994; Antonioli et alii, 2006). Uplift estimates cumulate the effects of both regional and local processes, the latter related to faulting (Westaway, 1993; Ferrari et alii, 2007; 2010; Roberts et alii, 2013) and possibly folding. Much of the regional uplift is associated to deep processes stemming from the Ionian subduction (Faccenna et alii, 2011; Roberts et alii, 2013), as suggested by the spatial coincidence between the locus of largest surface uplift and the extent of the slab (Fig. 38). Uplift is interpreted as the response to asthenospheretic wedging into the gap resulting from slab detachment (e.g. Westaway, 1993; Wortel and Spakman, 2000) or from crustal delamination (Gvirtzman and Nur, 2001), or as due to underplating beneath the accretionary wedge or finally as the viscoelastic response to enhanced erosional flux from land to sea following the onset of glacial-interglacial
cycles (Westaway and Bridgland, 2007). A smaller fraction of Quaternary uplift has been related to footwall uplift along extensional faults (e.g. Westaway, 1993; Tortorici et alii, 2003). This occurrence concerns large planar normal faults that rotate about a horizontal axis while they move, causing a tilt that is observable in the geological record (Jackson and McKenzie, 1988). Conversely, hanging-wall subsidence counteracts the effects of regional uplift, which in the long-term prevails (Valensise and Pantosti, 1992; Roberts et alii, 2013). The pattern of uplift of the Calabrian Arc is well depicted by using as a prominent benchmark the Last Interglacial Time (LIT) marine terrace (Ferranti et alii, 2006, and references therein). The LIT shoreline coincides with the Marine Isotope Sub-stage (MIS) 5.5, which occurred between 132 and 116 kyr (Shackleton et alii, 2003). On the Tyrrhenian Sea side of Calabria (Fig. 39), the LIT marker elevation provides average uplift rates which peak abruptly at Capo Vaticano (~0.6-1.2 mm/yr, and up to 2 mm/yr according to Tortorici et alii, 2003) and again to the south in the Messina Straits (~1.0-1.4 mm/yr) between Calabria and Sicily. The uplift pattern in the Strait is asymmetric, with larger motion on the eastern (Calabria) side. Moving southward, larger uplift prevails on the western side (Sicily), an occurrence that may reflect a contribution from the Etna volcano. Further south, along the eastern Sicily coast, uplift rates decrease steadily and attains a quasi-eustatic elevation at the southern tip of the island. Similarly, along the north-eastern shore of Sicily, terraces elevation documents that the large uplift province of the Calabrian arc grades westward to quasi-eustatic values.

Figure 39. Vertical displacement rates (mm/yr) computed from the elevation of Late Pleistocene markers plotted with main structures on a DEM of Calabria and eastern Sicily (from Ferranti et alii, 2010).

On the Ionian Sea side of Calabria, uplift rates recorded by the LIT terrace are slightly weaker (0.6-0.8 mm/yr) along a 300 km stretch of coast till the border with the Basilicata region (Fig. 39). Along this coast, major faults are not mapped and large earthquakes are not recorded, and so vertical displacement might chiefly embed the regional component. Along the coasts of Sila and Pollino mountain ranges in northern Calabria, however, the LIG terrace indicates again uplift rates of magnitude similar to southern Calabria (Fig. 39). Unlike the Tyrrhenian side of Calabria, however, uplift was attributed to the combination of regional uplift and growth of crustal transpressional folds, based on coastal morphotectonic and offshore geophysical data (Ferranti et alii, 2009; Caputo et alii, 2010; Santoro et alii, 2013; Ferranti et alii, 2014).

Holocene uplift rates were established in the southern part of the Calabrian Arc, where they are nearly 50% higher than late Pleistocene rates (Pirazzoli et alii, 1997; De Guidi et alii, 2003; Antonioli et alii, 2006; 2009; Ferranti et alii, 2010), but they depict a pattern similar to that illustrated by the LIG marker. Higher Holocene rates have been attributed to clustered fault activity which equals regional displacements (Ferranti et alii, 2007), and several paleoseismic events are provided by analysis of displaced Holocene strandlines (Ferranti et alii, 2017).
In the regional geodynamic framework above sketched, the southern Tyrrenian Sea (Fig. 40) belongs to a segment of the Central Mediterranean plate boundary zone. During the Neogene and Quaternary the evolution of this zone underwent to a slow relative Africa–Eurasia plate convergence and fast subduction and roll-back of the Ionian lithosphere beneath the Calabrian Arc, accompanied by fast back-arc extension in the Tyrrenian Sea. Today, the Ionian slab is SW–NE striking and NW-dipping, along a narrow and steep Benioff plane down to about 500 km, as imaged by seismicity and seismic tomography. Although the subduction of the Ionian lithosphere was characterized by vigorous trench retreat and back-arc extension at a rate of 50–70 mm/yr during the Neogene and Quaternary, a large spectrum of geophysical and geological data document a major tectonic reorganization in the central Mediterranean around 0.8–0.5 Ma, when rapid trench migration and consequent Tyrrenian back-arc extension essentially stopped. Proposed models for the present-day geodynamics of the Mediterranean include plate force interaction, subduction slab dynamics, large-scale mantle flow and small-scale density anomalies in the mantle.

Figure 40. Tectonic map of the Sicily and Calabria area.
- Earthquake focal mechanisms: Harvard-CMT, INGV-RCMT and ETH-RMT catalogues;
- Red stars: historical earthquakes with M≥6.5 from the CPTI04 catalogue (http://emidius.mi.ingv.it/CPTI04);
- White arrows: competing kinematic boundary conditions for the Ionian Sea, after (D’Agostino et alii, 2008);


The NE Sicily and Calabria region is characterized by some of the highest rates of seismic moment release along the Nubia–Eurasia plate boundary. Focal solutions of shallow earthquakes offshore northern Sicily outline a narrow E–W compressive belt, consistent also with geologic data. East of the central Aeolian area extensional to strike–slip mechanisms, occur along a NNW–SSE seismic lineament that runs from the central Aeolian Islands to Mount Etna and continues S-ward along the eastern Sicily escarpment. Shallow extensional seismicity also characterizes the Mt. Etna area, and a diffuse belt running ESE–WNW in northern Sicily. The transition zone between these two domains approximately corresponds to the western lateral termination of the Ionian slab at depth of 100–200 km, and to the Vulcano–Tindari–Giardini Fault system at shallow depths. The active tectonics inferred from focal mechanisms in Calabria is less well defined. However, extensional tectonics and regional uplift characterized the onshore Calabrian Arc during the last 0.8 Ma, and strong historical normal-
faulting earthquakes indicate that extension is the dominant style of faulting along its Tyrrhenian side. GPS and seismological data consistently locate the Messina Straits at the boundary between two different kinematic and tectonic domains, depicting NE-Sicily and Calabria as a key region for understanding of active geodynamic processes in the central Mediterranean. Although most of GPS data show that the Tyrrhenian basin is not actively spreading, or at least at a fraction of the past rates, debate concerning the present-day Ionian–Calabria convergence rate and direction, and consequently the way this convergence is eventually accommodated along the plate boundary. The motion of Sicily with respect to Nubia has been interpreted either as evidence of an independent Sicilian microplate, or, together with the observed GPS velocities on Apulia, as the expression of a larger Ionian–Apulian plate, which would include the Hyblean plateau in SE Sicily (Figg. 41–42). The presence of an Ionian–Apulian plate has been invoked to explain the observed kinematics along the Calabrian Arc, that Goes et alii (2004) interpreted in the framework of recent plate reorganization, causing the Calabrian and Ionian domains to be driven E-ward by the Hellenic slab pull. In this complex crustal dynamics, the Aeolian Islands are subsiding (Figg. 40–43) while Calabria is uplifting (Fig. 40) (Serpelloni et alii, 2014).

**Figure 41.** GPS velocity field with respect to the fixed Eurasian frame. Error ellipses are at 95% confidence level.

**Figure 42.** Geodetic strain rate field computed over a regular 0.2°×0.2° grid. Red and blue arrows show extensional and compressional strain rates, respectively. Grey crosses display 1σ uncertainties. Grey arrows show GPS velocities with respect to the Nubian fixed frame.

**Figure 43.** Geodetic vertical velocity in the Aeolian Islands from the CGPS network.

Between Vibo Valentia Marina and Briatico, the tectonic unit of Capo Vaticano is exposed. This is a structural NE-SW trending high, bordered toward the SE by the Mesima basin, two main NW-SE trending antithetic faults (Mileto fault), in SSW by the Coccorino and Nicotera faults with WNW-ESE trends, and toward N-NW by a fault system, down-lifted northward. Tortorici et alii (2003) found in the area of Briatico a typical morphological aspect of a bridge zone, placed between two fault planes. The current elevation of the Pleistocene marine terraces allowed estimation of the vertical tectonic rates in this area. These outcrops, which are approximately homogeneously distributed, show significant height differences. Ferranti et alii (2006) speculated that the large uplift values recorded by the Calabria-Peloritano arc can be related to the general uplift of the lithosphere above a subducting plate.
In 1994, an impressive geomorphological work done both in field and on the computer by Dai Pra, Sylos Labini and T. Miayuchi, a Japanese researcher who did his doctoral thesis in Italy was published (Miayuchi et alii, 1994). After 25 years this work still remains one of the most detailed and well done geomorphological map for the SW Calabria coast (see the map shown in Fig. 45): Terrace X, aged Last Interglacial Time since often contains Senegalese fauna or *Persistrombus latus* shows its inner margin at an altitude varying between 48 m (Vibo Valentia, north Calabria) and 143 m (Nocella, southern Calabria) suggesting an uplift-rate ranging between 1.14 mm/yr and 0.38 mm/yr. Terraces I and II correlate to the Lower Pleistocene (about 1.1 Ma) stay between 630 and 1.350 meters suggesting an uplift rate that vary between 0.5 and 1.12 mm/yr.

**Figure 44.** The MIS 5.5 terrace (green line with elevation in meters) and the main faults at Capo Vaticano promontory (from Miyauchi *et alii*, 1994). Numbered blue squares are 1) the location of the Scoglio Galera fish tank and 2) the pier at the mouth of Trainiti creek.

**Figure 45.** The marine terraces of the Tyrrhenian coast of the Calabrian Arc in the area of Capo Vaticano by Miyauchi *et alii* (1994).

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As for the coastal deposits uplifted in late Holocene, different Authors calculated the uplift rates according to the model of Lambeck et alii, 2004, 2011, in the Calabrian sites of Scilla, Palmi, Joppolo and Capo Vaticano (Pirazzoli et alii, 1997; Antonioli et alii, 2006, Ferranti et alii, 2008, 2009, Spampinato et alii, 2016) and in the Sicilian sites of Taormina, Sant’Alessio, Capo Peloro and Capo Milazzo in Sicily (Antonioli et alii, 2003, Antonioli et alii, 2006, Scicchitano et alii, 2011); they find values that, when compared to those of MIS 5e vary between 1.4 and 2.2 mm/yr seem indicate a double uplift rates respect Mis 5e. Miyauchi et alii (1994) observed along the Tyrrhenian coast of southern Calabria twelve orders of Pleistocene terraces up to 1350m above sea level. Among the areas where uplift is greater (Monte Poro, Le Serre and Aspromonte), the zone of Capo Vaticano promontory shows a large differential uplift (Fig. 45). Here, the Tyrrhenian terrace of MIS 5e is at about 50 m above sea level near Vibo Valentia. At Capo Vaticano, it is up to 120m above sea level, a difference of about 70m, in contrast with the recent ages published by Tortorici et alii (2003) who found an elevation even up to 285 m in this area. Fig. 46 shows the seismicity of the Calabrian region according to the database of the Seismic Network of Calabria University. In black are plotted the events with ML >2.5 and focal depth to 50 km in the period 1981-2011. Red void squares indicate macroseismic epicentres of historical earthquakes in Calabria with ISO VIII MCS from CPTI catalogue (CPTI Working Group, 2004). The inset shows a compilation of fault plane solutions for earthquakes shallower than 50 km in the Calabrian area and ML > 2.8, obtained by the combination of: a) the EMMA database by Vannucci et alii (2004); b) the Global and Italian CMT Catalogs http://www.globalcmt.org/; Pondrelli et alii, 2006); c) original computations performed in the framework of the SyNaRMs Interreg IIIB Project (Guerra, 2007); d) specific papers published more recently.

In historical times, the city of Vibo Valentia was destroyed by the catastrophic earthquake (M 7.0, ISO XI MCS) of September 8th, 1905. The location of the source of this shock, and consequently the identification of the major tectonic structure from which it was generated, is still debated since Mercalli (1906) and Baratta (1906). Several Authors have studied more recently or are presently studying the problem, by using different approaches, but a sound solution has not yet been reached. The Capo Vaticano promontory and its surrounding sea are characterized by a very low level of surface seismicity. On the basis of the knowledge of the historical seismicity and of the instrumental data for the last decades, the existence of coseismic differential displacements between different domains inside the Monte Poro complex in historical times can be excluded.
STOP 7 - Archaeological evidence of relative sea level changes: the Trainiti pier and the Scoglio Galera fish tank (Lat. 38°43’28”N – Long. 16°00’01”E)

Between Vibo Valentia marina, at the mouth of the Trainiti creek and the village of Briatico, two archaeological structures of Roman age are located (Fig. 47). The first is represented by a breakwater built at the mouth of Trainiti creek near Porto Salvo village. The second is a fish tank, excavated on the Scoglio Galera, a small rocky islet near the small village of Santa Irene (Briatico) (Fig. 42). With the goal to estimate the vertical motion of the around the fish tank of Scoglio Galera, to get an overview of the morphology of the seafloor, to help in interpretation.

7.1 The Trainiti pier

This archaeological site consists of a submerged structure, approximately rectangular in shape, about 320 m long and 40m wide (Fig. 47 ab). It extends NW-SE, oblique with respect to the coastline to which is connected. Another smaller wing was previously found by Mariottini (2001), but it was not found during these investigations. Both wings formed a typical harbor entrance named fauces, by their position suitable to protect an inner harbor. These structures can be interpreted as offshore breakwaters pertaining to a large harbor, likely belonging to the ancient port of Hipponion/Valentia, which presently is completely buried by coastal sediments (i.e: Schmiedt et alii, 1966 and reference therein). The structure was built with large blocks of concrete that contain several fragments of amphorae dated to Roman age (2 ka BP).

The constructional features of the main wing show at least three homogenous building sectors of: i) strongly cemented material containing abundant fragments of archaeological remains, ii) dissolved material, with poor archaeological remains and finally, iii) breakwater material, with elements partially built but without archaeological material. The surface of the structure is rough, and its base follows the irregular topography of the seafloor, which is tilted toward offshore. It begins from the shore at an elevation of about 1 m and extends offshore up to about 9 m at its end. The surface of this structure was likely covered by pavements, now destroyed. Similar structures are found in other areas of the Mediterranean settled since Roman or pre-roman times, such as at Tharros, in Sardinia. Previous geomorphological and paleo-geographical studies determined that this area suffered from significant changes of the coastline, which have caused the modification of the coastal plain that evolved toward a coastal environment with marshes and lagoons, separated by the sea from sand dunes. Hence, this transitional zone underwent continuous flooding and silting caused by the Trainiti and S. Anna creeks. The proggradation of the coastline and its timing is confirmed by Cucarzi et alii (1995) from the migration toward north of several phases of human settlements.

The old topographic maps of the Italian Istituto Geografico Militare (IGMI) show some ruins that could match the ancient harbor structures as reported in XVIII century by Priest Fiore. The description made by Schmiedt (1966) is the same as left by Fiore in 1680, when the harbour was “destroyed under the order by the Roman pope to give to the barbarians a poorhouse” and that this harbour was “built with cut stones from the ancient inhabitants of Hipponion, in a shape similar to a bended arm”, in a time when most of its parts could still be seen. Schmiedt (1966) cites previous descriptions “Even today in the low and calm tide can be observed large remains of a big construction composed by very large blocks with arches and pillars of concrete and even the moorings to tie ships” and “the Harbour of Hipponium was placed in the bay in front of the castle of Bivona, at that time partly in a lagoon and connected with sea; near the shore of the lagoon there were large squared pillars built with bricks, distributed in regular intervals, that were outcropping from the sand and likely holding arcades circling the whole harbor”. Schmiedt himself (1966) observes that "on the shore, in the same zone indicated by the aerial photographs, are present structures"
built with bricks that seem to be ancient; it is not an hazard to collocate the ancient basin of the harbour of Hipponium in the bay where a lagoon was existent in the previous century”. All these information support the hypothesis that during the Roman time the current coastal plain was the site of the ancient port. The submerged structures, in particular those placed at the mouth of Trainiti creek, are likely part of the ancient Harbour which had two entrances: the first at the mouth of Trainiti creek, and the second in the eastern side of the lagoon. The underwater structure can be interpreted as a system to facilitate the outflow of the debris transported by creeks that exit in the basin, preventing its progressive infilling.

7.2 The Scoglio Galera fish tank

Scoglio Galera is located near the village of Briatico, about 100 m from the coastline. It is about 120 m long and about 40 m wide, extending E-W. This small islet was known from historical tradition to be used by Arabs to jail the Christians, sinking them in the pools. Recently, marine archaeologists classified this site as a fish tank and a fish processing plant (Fig. 48). The islet was excavated and cut at its surface, thanks to the softness of the biomarlstone and limestone of Miocene age.

Along the NNW side of the islet, which is the most exposed to the sea, the remnant of a wall and traces of the formworks used for the concrete, are still present. The fish tank consists of four nearby pools, E-W aligned, that follow the natural morphology of the islet (Fig. 49). The pools have a total length of about 28 m and a constant width of 2.5 m. The two main pools are subdivided into minor pools and their inner walls show some holes at 1 m above sea level, that were likely used to host horizontal wooden beams and a roof. The pools are crossed by two main channels, A and B, which link the inner basin with the open sea. The latter is protected and suitable for the moorings of ships. In particular, channel B crosses the islets by a tunnel. Two additional minor channels, C and D, connect the pools 2 and 3 with the inner basin. The pools are all connected by channels and separated by partition sects. All the channels show the signs of the grooves used to operate the sluice gates, similar to those found in the fish tanks along the Tyrrhenian coasts of central Italy and other localities. These were used to provide an effective water exchange in the basin but without letting the fish escape. The inner side of the basin shows i) the crepidini, narrow sidewalks used to walk around the pools without getting wet, ii) the surface of a dock, and iii) ten bollards of different size, all rock cut in the islet.

![Figure 48. Sketch of the channel sluice gate with sliding posts, threshold and lowest level crepido as viewed from within the fish tank. The threshold defines the lower limit estimate and a level 20 cm below the lowest footwalk defines the upper limit estimate. The top of the sluice gates coincides with the elevation of the lowest level footwalks and corresponds to a position above the highest tide (from Lambeck et alii, 2004b).](image-url)

The inner side of the pools show a present day notch about 40 cm high and 30-60 cm deep. Its lower part shows an organic platform typical of environments at high hydrodynamics, which is in agreement with the amplitude of the local tides (Fig. 49 cde). The underwater part of the islet, particularly inside the mooring basin, shows a selective erosion process, acting in coincidence with the different level of local stratigraphy, separated each other by 40 cm. The surface of the islet shows an additional squared small pool of about 1.5 x 1.5 m in width, crossed by a channel without sluice gates.

The geology of Scoglio Galera belongs to the outer limits of the youngest marine terrace of Upper Pleistocene that is exposed between 0 and 30 m above sea level. It consists of littoral deposits formed by alternating clastic facies with variable size of strongly cemented conglomerates, sands and bio-limestone. The thickness of the strata ranges from 20 to 250 cm and those aligned N 260°-290° are tilted 10°-15° toward NNE. The thickness of the bio-limestones is about 200 cm in the islet. As the strata are all tilted NNE, the islet has an asymmetric morphology. Its southern part is indented and with a vertical coastline up to 3 m high, while the northern part shows a linear trend and is gently tilted NNE.

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Along the southern border, inside the pools and in the mooring basin, a present day erosion notch 30-60 cm wide and 50-100 cm high, is present. The northern border displays an abrasion platform up to 6 m wide, characterized by *Dendropoma petreum*. These gastropods usually live in the lower intertidal environment.

The northern border shows also relics of a marine terrace 60-80 cm above sea level, at the same level of the walking surface (*crepidine*) that runs along the mooring basin. The *crepidine* was partially excavated on the surface of the marine terrace. The cross section shows the cliff running along the northern coast of the islet whose bottom ends in the seafloor at a depth of 6-7 m. The elevation of the intertidal abrasion shows a level of 7-8 cm of *D. petreum* which is in agreement with the current sea level. This platform includes several squared holes, excavated in the islet to support the wooden poles that made up the reinforcements of high concrete walls, built to protect the islet from the waves. The platform ends with a small scarp 80 cm high, over which were developed a small terrace 3.5 m wide. The latter ends with an additional small scarp, and its surface shows other squared holes where parts of the walls are still remaining. Although this may appear as natural terrace, it appears to have been excavated to adapt the surface of the islet to the wall foundations. Along the opposite side of the islet the seafloor is up to 4-5 m deep and shows an erosion-smoothed marine notch. The present day erosion notch has a size a few cm larger than the tidal range.

![Figure 49](image_url)

**Figure 49.** a) Aerial view of the Scoglio Galera and b) its topographic map (modified from Mariottini, 2001) c) the pools of the fish tanks and d) their partition sects and the erosion notch. Below the sects run the channels that link the nearby pools. e) Particular of the Scoglio Galera: Black dots are the bollards excavated in the rock; pools are numbered 1-5; the main (A,B) and secondary (C,D,E,F) channels, are also reported in the map. f) An underwater view of the islet from inside the harbor basin. These notches are caused by selective erosion processes occurring along the different levels of the geological unit. g) The abrasion platform with the *Dendropoma petreum* which is drilled with several squared holes used to host wooden poles. A relic of wood allowed us to date by 14C AMS this site at 1806±60 years BP. h) One of the holes for the wooden poles of 20x20 cm size. k) One of the channels used for water exchange in the pools. Their functional elevations still correspond to those at the time of the construction of the fish tank. The openings of the channels were originally closed by fixed gates to avoid the escape of fish.
Discussion and conclusions

The sea level curve estimated by Lambeck et alii (2011) predicts a sea level change for this location of about -113 cm since the last 1806±50 years, at a mean rate of -0.63 mm/yr (i.e.: Fig. 39). Observations have estimated that the relative sea level has changed only 7 cm since the construction of the archaeological site. Therefore, the recent tectonic uplift inferred from the fish tank exceeds by 0.18 mm/yr the previous estimates of Miyauchi et alii (1994) obtained from the elevation of the MIS 5e marine terrace, which infers an uplift rate of 0.47 mm/yr at Briatico. Bianca et alii (2011) found the MIS 5e terrace at 216 m of elevation, in contrast to the value of about 60m reported in Miyauchi et alii (1994), thus inferring a long term uplift rate up to 1.74 mm/yr. As this value is in large contrast with the previous measurements of Miyauchi et alii (1994), who dated this terrace through the P. latus, the latter is preferred for comparison here. Besides the long term geological data, recent geodetic observations are available in Calabria which show the present day uplift of this region. The tide gauge of Reggio Calabria has recorded a sea level trend of 0.28 mm/yr during the time span 1999-2007 (Fig. 50). Although the duration of the sea level recordings for this station is still too short to provide a robust estimation, this value can be considered a preliminary indication that should correspond to a deficit of about 0.8 mm/yr in the sea level increase with respect to tectonically stable coastlines of Italy. This result is in agreement with Braitenberg et alii (2011). Moreover, the available GPS data show a variable uplifting trend for southern Calabria (Devoti et alii, 2010). Therefore, the rate of crustal uplift inferred from the marine archaeological site of Scoglio Galera provides the evidence of a continuous uplift, whose trend is in agreement with the recent instrumental observations. The archaeological data fills a gap between geological and instrumental data, providing new estimations on the tectonic trend of this region during historical times.

![Figure 50](image_url)

**Figure 50.** Plot of the tide gauge recordings at Reggio Calabria station from 1999 to 2007. Data show a sea level trend at 0.28±0.1 mm/yr (linear fit shown by the black line). The maximum tidal range for this location is 0.63 m, in agreement with the height of the erosion notch at Scoglio Galera.

The Calabrian Arc is a very active region dominated by large horizontal and vertical land movements. In southern Calabria, the timing of the relative sea level changes at Briatico, has been inferred from the elevation of the Quaternary marine terrace and the coastal archaeological sites of roman age. As the pier located at the mouth of Trainiti creek does not provide any precise data, the only available indicator along this coast is the Scoglio Galera. Analysis used: i) functional elevations of channels used for water exchange within the pools; ii) functional elevation of crepí dine and bollards; iii) elevation of the erosion notch in the pools, correlated with the Dendropoma platform placed along the northern side of the islet, and iv) the elevation of the marine terrace of the upper Holocene, excavated at 80 cm above sea level along the northern side of the islet.

As the age of the fish tank is 1806±50 cal BP, the present day erosion notch in the inner side of the pools formed only after the fish tank was excavated. This geomorphological indicator, strictly related to the recent sea level, developed during this time, reaching a size related to the mean tidal range for this location, as estimated from tidal data (Fig. 47). Excluding any uplifting coseismic displacement, these observations imply that the fish tank, since the time of its construction, underwent continuous uplift with the same value of the subsidence signal caused by the glaciohydroisostasy.

The erosion notches formed in this tectonic environment and developed under a constant tidal range amplitude, since the relative sea level remained steadily at the same relative level while it was continuously rising together with the Earth’s crust. The pools do not show any morphological evidence of submerged erosion notches, supporting the hypothesis that relative sea level has not changed since their construction.
Assuming a constant rate of uplift since the late Quaternary, including the last 1806 years, then the current elevation of the Scoglio Galera is due to the glacio-hydro-isostatic signal that counterbalances the tectonic signal. Therefore, the uplifting rate for this location can be estimated at about 0.65 mm/yr for the last 1806±50 BP. The present day biological marker, the Dendropoma platform, is in agreement with these observations as well as with the tidal notch along the inner side of the pools of the fish tank. These markers developed in an environment characterized by a relative null vertical motion, and consequently are at the same elevation. This balance also includes the recent eustatic increase of 13 cm, as estimated for the Mediterranean by Lambeck et alii (2004b).

On the base of these data, the archaeological markers at Scoglio Galera did not record coseismic significant displacements along the vertical, including the December 8th, 1905 Ms 7 earthquake, which is the largest in this region in the last centuries, in agreement with the proposed dislocation model by Piatanesi and Tinti (2002). However, cosmic movements may have occurred before the construction of the fish tank.

In the Aeolian Islands, the relative sea level change is driven by volcano-tectonic land subsidence. For the last 2 ka, Lipari underwent to a mean subsidence at 5.79±0.01 mm/yr that caused a RSLC at 12.3±0.7 m. This imply an important VLM gradient between the subsiding Aeolian islands and the uplifting Calabria, that are located only a few tenth of km distant.

**Figure 51.** Plot of the predicted sea level curve (from Lambeck et alii, 2011) compared with the elevation of the archaeological markers of the fish tank (diamonds with error bars for age and elevation). The archaeological indicators remained at the same elevation with respect to local sea level since 1860±60 ka BP. The data at the top right of the curve show the current mean elevation value of the markers, still matching the local sea level (From Anzidei et alii, 2013).

**Figure 52.** Archaeological evidence of land subsidence at Lipari Island, in contrast with the uplift at Briatico. The plot shows the predicted sea level (from Lambeck et alii, 2011) compared against the elevation of the Roman age pier at Lipari, dated at 2100 ± 100 yrs BP, with uncertainties for age and elevation (black diamond with error bars). A relative sea level change at 12.3 ± 0.7 m has been estimated from the position of the archaeological indicator, that falls 10.86 ± 0.70 m below the predicted sea level at 2.1 ka BP (1.31 m) and the recent eustatic rise (0.13 m). Sr is the observed subsidence at rate at 5.79 ± 0.1 mm y⁻¹, while Tr is the rate of the estimated tectonic and volcanic contribution to the observed subsidence at 5.17 ± 0.1 mm y⁻¹. Sr and Tr have been reduced for the intervening eustatic change SLCe at 0.13 m as estimated by Lambeck et alii (2004b) (from Anzidei et alii, 2014).
The Capo Vaticano promontory is part of the Calabrian Arc (Fig. 53a) that represents one of the major seismogenic areas in the Mediterranean region and is characterized by large historical earthquakes (Boschi et alii, 1995, 1997). Most seismic events occurred along two overlapping extensional belts located along the Tyrrenian side of southern Calabria and the Ionian coast of eastern Sicily, and spatially coincide with strongly uplifted sectors. The vertical movements are recorded by Pleistocene marine terraces and by Holocene paleoshorelines (Fig. 53b). In particular, in north-eastern Sicily and southern Calabria, data on Holocene vertical tectonic motion have been obtained by elevation measurement and radiometric dating of biological and morphological markers (Stewart et alii, 1997; De Guidi et alii, 2003; Antonioli et alii, 2003, 2006; Ferranti et alii, 2007; Scicchitano et alii, 2011a,b; Spampinato et alii, 2012). In the Calabrian Arc vertical tectonic motion during the Pleistocene and Holocene is characterized by two source contributions (Westaway, 1993; Ferranti et alii, 2006, 2007): a regional component, slow and constant, that is reflected in large (~100 km) wavelength movements (steady uplift); a local component, of more limited (~10 km) wavelength, attributed to fault slip (co-seismic uplift). Although long-term uplift of Capo Vaticano has been studied since long time (Miyauchi et alii, 1994; Tortorici et alii, 2003; Cucci and Tertulliani, 2006; Bianca et alii, 2011), data on Holocene coastal deformation were sparse until Spampinato et alii (2014) and Lo Presti et alii (2014) performed a detailed mapping of morphologic, biological and archaeological indicators present all over the Capo Vaticano peninsula coastlines.
Quaternary extensional tectonic has accompanied a strong regional uplift which developed spectacular flights of marine terraces (Ghisetti, 1981; Dumas et alii, 1987; 1988; 1991; Westaway, 1993; Miyauchi et alii, 1994; Bianca et alii, 1999; Catalano and De Guidi, 2003; Tortorici et alii, 2003; Bianca et alii, 2011). The elevation of marine terraces and their offset across the main faults has been used to establish the relative contribution of regional and fault-related sources to uplift. Maximum cumulative uplift rates averaged since the Middle Pleistocene are estimated at 1.7 mm/yr (Westaway, 1993; Catalano Catalano and De Guidi, 2003). According to Westaway (1993), 1.67 mm/yr of post-Middle Pleistocene uplift of southern Calabria was partitioned into 1 mm/yr due to regional processes and the residual to displacement on major faults.

The Capo Vaticano area is a structural high bounded by two antithetic normal faults (Tropea and Mileto) oriented NE-SW (Fig. 54). To the northeast, the Briatico area represents the overlap zone between the en-echelon Tropea and Vibo faults. This area, is interpreted as a relay ramp by Tortorici et alii (2003). The occurrence of a NE-SW striking, NW-facing normal fault system in the north-western offshore (Fig. 54) has been displayed by a SSE-NNW striking single-channel seismic profile (Vp3 in Trincardi et alii, 1987), showing a prominent normal fault scarp, located at about 10 km from the coastline. The bathymetric map constructed by Argnani and Trincardi (1988) shows another ENE-WSW trending fault scarp, located at about 5 km from the coastline (Fig. 54). To the southwest, the promontory is abruptly truncated by two major WNW-ESE striking, SSW-dipping normal faults (Coccorino and Nicotera faults) which separate the Capo Vaticano structural high from the Gioia Tauro Basin. Landward, to the south-east, antithetic SE-facing fault segments (Mileto fault) separate the Capo Vaticano horst from the Mesima Basin (Fig. 39).

The geological backbone of the Capo Vaticano peninsula mainly consists of granites and gneiss of the Palaeozoic basement, covered by discontinuous remnants of Miocene and Pliocene carbonate and terrigenous deposits (Burton, 1971), on top of which eight distinct orders of well preserved Quaternary erosion surfaces and marine terraces have been recognized. Their number and their distribution are different (Westaway, 1993; Miyauchi et alii, 1994; Tortorici et alii, 2003; Cucci and Tertulliani, 2006; 2010; Bianca et alii, 2011). Previous studies identified a NE tilt of the promontory as reflected by a decreasing elevation of the Pleistocene marine terraces toward the NE. On the SW side of Capo Vaticano (Fig. 54), the tilt pattern is interrupted. According to Tortorici et alii (2003) the tectonic setting of the promontory is due to the combined Middle-Upper Pleistocene activity of both regional uplift and faulting. Detailed mapping, performed by Spampinato et alii (2014) and Lo Presti et alii (2014), all around the coast of Capo Vaticano (Fig. 54a) allowed to confirm the existence of the uplifted shorelines, previously described by Antonioli et alii, 2009, and to find evidence of other markers such as marine deposits, barnacle rims, Dendropoma rims and notches (retrieving material for shoreline dating), and geoarchaeological sites. Survey documented the occurrence of four Holocene paleo-shorelines raised at different altitudes. The uppermost shoreline (PS1) is represented by a deeply eroded fossiliferous beach deposit, reaching an elevation of ∼2.2 m above the present sea-level, and by a notch whose roof is at ∼2.3 m. The subjacent shoreline PS2 is found at an elevation of ∼1.8 m and is represented by a Dendropoma rim, a barnacle band and by a wave-cut platform. Shoreline PS3 includes remnants of vermetid concretions, a barnacle band, a notch and a marine deposit, and reaches an elevation of ∼1.4 m. The lowermost paleo-shoreline (PS4) includes a wave-cut...
platform and a notch and reaches an elevation of ∼0.8 m. Along the PS4 Lo Presti et alii (2014) found a millstones quarry carved on well-cemented Miocene sandstones, about 50-60 m thick, located between 40 cm a.s.l. and ~40 cm a.s.l. Radiocarbon dating of material from individual paleo-shorelines points to an average uplift rate of 1.2–1.4 mm/yr in the last ∼6 ka at Capo Vaticano. Data suggest that Holocene uplift was asymmetric, with a greater magnitude in the south-west sector of the promontory, in a manner similar to the long-term deformation attested by Pleistocene terraces. The larger uplift in the south-western sector is possibly related to the additional contribution, onto a large-wavelength regional signal, of co-seismic deformation events, which are not registered to the north-east. Have been recognized four co-seismic uplift events at 5.7–5.4 ka, 3.9–3.5 ka, ∼1.9 ka and <1.8 ka ago, superposed on a regional uplift that in the area, is occurring at a rate of ∼1 mm/yr.

The Santa Domenica area

Santa Domenica is located about seven kilometers to northeast of Capo Vaticano (Fig. 54b). The coast is carved into Miocene sandstone, and for this reason dissolution features, such as notches, were better developed there. Ancient sea-level markers were surveyed in this area at several localities, namely at Lo Scoglio and at Petri i mulinu. In locality Lo Scoglio were found two notches that are clearly distinguishable along the coastline of eastern side (Figg. 55-56).

The lower notch appears well developed and preserved and no biological remains were found associated with the notch. The roof of the lower notch was measured at an elevation of ∼0.90 m and the maximum concavity at an elevation of ∼0.70 m (Fig. 56). A fossil barnacle of the species Chthamalus depressus has been found above the lower notch at an altitude of ∼1.50 m a.s.l. The upper notch is larger and wider than the lower notch, with a roof at ∼2.30 m (Fig. 56). Along the coastline of the Scoglio Riaci was found a wave-cut platform that has a width of ∼0.70 m and extending from ∼0.40 m down to the present coastline. Along the north side of Scoglio Riaci another wave-cut platform was found; its inner edge has been measured at an elevation of ∼1.40 m.

Figure 55. Sketch map of Capo St. Domenica; Scoglio Riaci and Petri I Mulinu are the site in which geoarcheological and biological markers are preserved.

Figure 56. Raised morphological markers of the Holocene paleo-sea level occurring along the coast of St. Domenica (Lo Scoglio); UN: upper notch, LN: lower notch.
In locality Petri I Mulinu a notch partially covered by a beach deposit has been recognised. This notch appears well preserved and, although several lithophaga holes are present, no biological remains were found associated with the notch. The maximum concavity of the notch was measured at an elevation of ~65 m. Remains of a fossil barnacle band have been found above the notch. The band is ~0.20 m wide, and its top reaches an elevation of ~1.30 m. A fossil balanid shell was collected from the upper portion of the band at 1.26 m and yielded an age of 2914±137 cal BP (Fig. 57).

Figure 57. St. Domenica (Petri I mulinu): band of fossil barnacles. Inset showed the dated balanid.

In Petri I Mulinu locality, Lo Presti et alii (2014) described the presence of a wide millstones quarry carved on well-cemented Miocene sandstones. Millstones quarries (Fig. 58) are documented in Southern Italy since around 2500 BP (Amouretti, 1986; Amouretti and Brun, 1993; Brun, 1997). Their sizes and shapes fit the carving systems used at those times, characterized by cylindrical or slightly truncated cone shape wheels that turn perpendicularly above a subjacent horizontal wheel of similar size placed above a masonry base (Amouretti, 1986; Hadjisavvas, 1992; Brun, 1997; Rosada, 2007). Their use is reported since the beginning of the Hellenistic period, but a large spread in the Mediterranean basin is evident from the Roman to the modern age. After the XIX century, this traditional crusher system was progressively abandoned (Amouretti and Brun, 1993). Quarry identified in Petri I Mulinu cover an area of about 30mx20m and it’s partially submerged lying at ~40 cm depth. At this site, some millstones (broken or defective) as well as numerous rings indicating complete extractions are still preserved in situ: Although the Petri i mulinu quarry complex is very similar to those described by Lo Presti et alii (2014) along Italian coast, and more in general to those studied by Antonioli et alii (2017) along Mediterranean coast, the vertical movements afflicting the Capo Vaticano area didn’t permitt the authors to age the archaeological site.

Figure 58. Millstone quarry discovered at Petri I Mulinu.
STOP 9 - Vertical land movements and sea level change in the Messina Strait: the Holocene raised wave-cut platforms along the Scilla coast (Lat. 38°15′09″N – Long. 15°42′07″E)

Geological framework

Intense Quaternary extensional tectonics, coupled with coastal uplift, are well documented in the Messina Strait (Fig. 59), a highly seismic area that was struck on 1908 December 28 by a M 7.1 earthquake and ensuing devastating tsunami (Baratta 1910; Shick, 1977; Aloisi et alii, 2012 and references therein). This structural depression is bounded by active normal faults, marked by well-preserved scarps, which displace Pleistocene marine terraces and Holocene shorelines (Ghisetti 1981; Valensise and Fantosti 1992; Ferranti et alii, 2007, 2008a; Scicchitano et alii, 2011; Monaco et alii, 2017).

The WSW-ENE to SSW-NNE striking extensional basin of the Messina Strait formed as a consequence of the Pliocene-Lower Pleistocene axial collapse of the inner sectors of the Calabrian arc. The Upper Pliocene-Lower Pleistocene deposition was followed by the uplift of the border fault footwalls and subsequent development, during the Lower-Middle Pleistocene, of huge submarine fan-delta systems (Messina gravels and sands). Since the Middle Pleistocene, the strong regional uplift has caused the emission of these fan-delta systems. In the meantime, the interaction between the uplift process and the eustatic sea level fluctuations caused the formation of flights of marine terraces on the basin flanks. Uplift rates have been higher in the Calabrian sector where the normal faults show evidence of recent activity. In particular, in the Ganzirri area (Sicilian side) the MIS 5.5 terrace is located at altitude of 90 m a.s.l., while in the Villa San Giovanni area (Calabrian side) it is uplifted up to 170 m a.s.l. (Ferranti et alii, 2006, Monaco et alii, 2017). Coastal tectonic studies have shown that faults located at/or intersecting the coast (Scilla, Reggio Calabria and Armo faults, see inset in Fig. 59) have recent activity. Late Holocene coseismic displacements on the ~30-km long Scilla Fault (Westaway 1993; Jacques et alii, 2001) are suggested by Holocene marine platforms and beach rocks which are uplifted above sea level on the fault footwall (sites a and b in Fig. 59, Ferranti et alii, 2007, 2008a). The latter authors dated two coseismic events at ~3.5 and ~1.9 ky BP, with estimated slips ranging between 1.5 and 2.0 m and Me ~6.9–7.0 (see Stop 2). The Reggio Calabria Fault (Ghisetti 1984, 1992) was considered the source of the 1908 earthquake by Tortorici et alii (1995) on the basis of morphotectonic, macroseismic and seismological observations, but evidence of active deformation is scarce. In contrast, the Armo Fault shows clearer evidence of Pleistocene activity (Ghisetti 1984, 1992; Aloisi et alii, 2012), and coastal studies suggest a possible reactivation during the Holocene (Scicchitano et alii, 2011). Marine geophysical investigations (Del Ben et alii, 1996; Guarnieri 2006; Ferranti et alii, 2008b; Argnani et alii, 2009) also highlight the prevalence of active faults on the eastern part of the Straits.

The uplifted shorelines

Along the Scilla coast two Holocene uplifted shorelines have been identified (Marina di San Gregorio and Punta Paci, see Fig. 59 for location; Antonioli et alii, 2004; Ferranti et alii, 2007; 2008a). Morphological and sedimentary constraints allows an elevation estimate at ~3.0 m a.s.l. for the upper shoreline at Punta Paci, where a large wave-cut platform outcrops, clearly visible from the national road (Fig. 60). Age constraints for the
shoreline range between ~5-3.5 ka. The lower shoreline is characterized by a prominent barnacle band, lying at elevations ranging between ~0.8 and ~1.9 m a.s.l., and an algal rim bored by Lithophaga holes is found at ~1.4 m a.s.l. below the denser patch of the barnacle band. Duration of the lower shoreline is tightly constrained by radiocarbon ages of barnacles between 3.5 and 1.9 ka, and its inception is in good agreement with cessation of the older shoreline (Ferranti et alii, 2007; 2008a).

Integration of on-land and offshore geomorphological and structural investigations coupled to mapping and extensive radiometric dating of the raised Holocene beaches reveals that these are located at the footwall of the western segment of the Scilla fault (see Stop 1) and that uplift has both steady and abrupt components (Ferranti et alii, 2007; 2008a). Radiometric dating of the shorelines indicates that rapid co-seismic displacements occurred at ~1.9 and ~3.5 ka, and possibly at ~5 ka (Fig. 62). Co-seismic displacement show a consistent site value and pattern of along-strike variation, suggestive of characteristic-type behaviour for the fault. The ~1.5-2.0 m average footwall uplift during co-seismic slips documents Mw~6.9-7.0 earthquakes with ~1.6-1.7 ka recurrence time. The palaeoseismological record based on the palaeoshorelines suggests that the last rupture on the Scilla Fault during the February 6, 1783 Mw=5.9-6.3 earthquake (Jacques et alii, 2001) was at the expected time but it may have not entirely released the loaded stress since the last great event at ~1.9 ka. It worths to note that this event triggered a large rockslide west of Scilla, where part of Monte Paci fell into the sea (Fig. 59) giving rise to a tsunami that principally affected the town of Scilla and the lowland of Punta Faro, on Capo Peloro, Sicily’s northeasternmost cape (Fig. 61) (Fago et alii, 2014).

Precise compensation for sea level changes constrains Late Holocene steady uplift during the interseismic intervals at ~1 mm/yr, a value consistent with long-term (0.1-1 Ma) estimates of regional uplift (Westaway, 1993). Thus, Late Holocene total uplift at ~1.6-2.1 mm/yr is nearly equally balanced between regional and co-seismic components.

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STOP 10 - Recent tectonics of the Messina Strait: the Pleistocene terraces on the Campo Piale Horst and the on-land prosecution of the Scilla fault (Lat. 38°14′52″N – Long. 15°41′15″E)

Geological framework

High resolution swath bathymetry data and multichannel sparker profiles (Ferranti et alii, 2008b) show that recent faults in the northern and narrower sector of the Strait are arranged in two broad ~NE–SW trending arrays with opposing polarity (Fig. 63). The NW-dipping fault array on the eastern side of the Straits, which represents the offshore extension of the Scilla and Reggio Calabria faults, is wider (~5 km), and large offsets of tens of meters are observed in the Middle Pleistocene–Holocene sedimentary sequence (Ferranti et alii 2008b). By contrast, the fault swarm on the western side has more limited appearance and is made up of discontinuous segments. The arrays are connected by a NW–SE trending transfer zone located between Messina and Reggio Calabria (Fig. 63), which seems to control the current release of low seismicity (Scarfì et alii, 2009).

Figure 63. Seismotectonic setting of the Messina Straits region (from Aloisi et alii, 2012). Faults (thick solid lines barbed on the downthrown side, dashed where inferred or submerged) after Ghisetti (1992), Monaco and Tortorici (2000), Jacques et alii (2001), Ferranti et alii (2007; 2009): ARF, Armo Fault; CF, Cittanova Fault; MSGF, Motta San Giovanni Fault; RCF, Reggio Calabria Fault; SCF, Southern Calabria Fault; SEF, S. Eufemia Fault; SF, Scilla Fault. The focal mechanism (after Gasparini et alii, 1985) and damage distribution of the December 1908 earthquake (data from Baratta 1910; Boschi et alii, 1995; Monaco and Tortorici, 2007) are indicated. Towns are labelled in white boxes: RC, Reggio Calabria; Me, Messina. The projection is UTM-WGS84. The levelling data of Loperfido (1909) are reported as yellow circles with average vertical change values. The blue star shows the macroseismic location of Michelini et alii (2005). Inset shows the location of the study area in the tectonic setting of the Central Mediterranean.

Similarly, multichannel seismic profiles collected by Argnani et alii (2009) within the southern, broader part of the Messina Straits place the master faults on the Calabrian side. Specifically, a 30-km long, NW-striking and west-dipping listric fault located at the SW tip of Calabria cuts the seafloor (SCF, Fig. 63). On the other hand, offshore seismic profiles (Monaco et alii, 1996; Del Ben et alii, 1996; Argnani et alii, 2009) do not show evidence of low-angle faults and of their effects underneath the Messina Straits. According to Argnani et alii (2009), the lack of evidence of extensional faults large enough to cause an M ~ 7 earthquake within the northern and western sector of the Straits support the contention that the 1908 seismogenic fault is located along the south Calabria offshore. The lack of recognition of clear surface faulting, however, made it difficult to determine the source of the 1908 earthquake.

The few seismological recordings of 1908 coupled with co-seismic vertical displacements documented by levelling data of Loperfido (1909) have been predominantly interpreted to support a blind low angle (30° to 40°) normal fault dipping towards the SE nearly parallel to the Messina Straits and located along the Sicilian coastline (e.g. Valensise and Pantosti, 1992; Amoruso et alii, 2002; DISS Working Group, 2016; De Natale and Pino, 2014 and references therein). However, high-resolution swath bathymetry and seismic profiles (Argnani et alii, 2009; Doglioni et alii, 2012; Ridente et alii, 2014) show that the modeled through-going ~N-S striking faults below the Straits are not present. Conversely, another set of models considers NW-dipping, high-angle normal faults on mainland Calabria as causative sources of the seism (Schick, 1977; Mulargia and Boschi, 1983; Ghisetti, 1984, 1992; Bottari et alii, 1986; Westaway, 1992; Tortorici et alii, 1995; Bottari, 2008). In particular, the macroseismic picture (Fig. 63) and the youthfulness and fresh bathymetric expression of many of the faults...
in the eastern array indicates that these faults may be activated during large or moderate-sized earthquakes. This interpretation is consistent with the regional structure of the Messina Straits area, characterized by master faults on the Calabrian side and associated antithetic faults on the Sicilian side (Ghisetti, 1984; Montenat et alii, 1991). Recently, new field evidence along with a re-evaluation of the levelling and seismic data have been used by Aloisi et alii (2012, 2014) to identify the Armo fault, a NW-dipping normal fault exposed in SW Calabria, as a possible source of the 1908 event (Fig. 63).

The sequence of the marine terraces

Along the Calabrian side of the Messina Straits, between Villa S. Giovanni and Piano di Matiniti, a complete sequence of ten Late-Quaternary fluvial-coastal terraces is observable at elevations ranging from 40 to 520 m a.s.l. (Fig. 64) (Monaco et alii, 2017). The terraced deposits are formed by calcarenites, marine sands and conglomerates, more or less cemented, directly lying on the crystalline basement or on the early Pliocene (Trubi Fm.), infra-Pleistocene (Vino calcarenites) or late-middle Pleistocene (Messina Gravels and Sands Fm.) sedimentary covers (Ghisetti, 1981; Dumas et alii, 1982; Miyauchi et alii, 1994). The marine deposits generally pass upwards to continental reddish silt with sands and gravels levels, thick up to 20 m.

The terraced series is partly displaced by the Scilla fault that borders the Campo Piale horst to the north. The scarp of the onland western segment of the Scilla fault is characterized by up to 70-m-high triangular facets, suggesting recent activity, and by a cataclastic zone in the crystalline bedrock, including NW-dipping fault planes (Fig. 65). The lowest (I order) terrace extends along to the coast, with inner edge around at 40 m a.s.l. It’s worth noting that it seals the Scilla fault north of Villa San Giovanni (Fig. 64). Terraces II, III, IV, V and VI extend around the Campo Piale horst, extensively outcropping along the south-west side, with inner edges at elevations of 60, 95, 120, 175 and 200 m a.s.l., respectively. The complete terraced sequence of the remaining orders outcrops only along the Campo Piale horst, where the terraces VII, VIII and IX show inner edges at elevations ranging from ~480 m to ~520 m. North of the Campo Piale horst, on the hangingwall of the Scilla Fault, the assignment of terraces to distinct orders is more complex. In this area only three terraced surfaces have been recognized: the terrace I shows a clear continuity parallel to the coastline; a second surface, with inner edges at elevations between 70 m and 90 m a.s.l., has been attributed to the terrace III outcropping on the Villa S. Giovanni area, suturing the western end of the Scilla Fault; a third surface, with inner edges at elevations among 100 m and 140 m a.s.l., has been attributed to the V order (MIS 5.5, 125 ka) by geomorphological correlations (see Miyauchi et alii, 1994; Jacques et alii, 2001).

Figure 64. Morphotectonic map of the northern sector of the Messina Straits area showing the major Quaternary faults, the distribution of Quaternary terraces and the location of dated deposits (from Monaco et alii, 2017). Location of Stops 9 and 10 are indicated.
As regards the age attribution, the finding of *P. latus* in the Reggio Calabria area (Bonfiglio, 1972; Dumas *et alii*, 1987) and the absolute dating obtained by thermos luminescence (TL) and optically stimulated luminescence (OSL) methods (Balescu *et alii*, 1997), perfectly constraint the age of the terraced deposits between 60 and 330 ka. In particular, an age of around 60 ka has been attributed to the sandy deposits of the lowest terrace (I order), outcropping in the Acciarello place near Villa S. Giovanni (Balescu *et alii*, 1997, see Fig. 64), while the morphological correlation of the terrace V with the *P. latus* deposits (see also Miyashita *et alii*, 1994; Dumas and Raffy, 2004), has allowed to correlate this last to the MIS 5.5(= 5.e corresponding to about 125 ka). This allowed us to correlate the other orders of terrace with the other positive peaks of eustatic curve occurred between the MIS 3.3 and the MIS 9.3 (Fig. 66). Terraces II and VI order cannot be correlated with positive peaks (see also Dumas and Raffy, 2004) and for this reason they have been tentatively attributed to the MIS 4 (74 ka) and to the MIS 6.5 (167 ka), respectively (Fig. 66).

The uplift rates change in time and space and represent the sum of the regional signal, homogeneous in the last 700 ka (Westaway, 1993) and the co-seismic vertical deformation induced by the fault activity around the main Quaternary faults. These two components are added on footwall of the seismogenic faults, while on the hanging wall the co-seismic vertical movements are subtracted to the regional rate. The estimated uplift rates range from 0.8 mm/yr for the period 330-60 ka, on the downthrown block of the Scilla fault, to 2.0 mm/yr (twice the regional component of 1.0 mm/yr estimated by Westaway, 1993) for the period 330-200 ka, on the Campo Piale high. The constant elevation of the I order terrace suggests an uniform uplift rate of 1.4 mm/yr along the entire coastal area, and, consequently, a deactivation of the western sector of the Scilla fault in the last 60 ka, even though the offshore activity of segments belonging to the same system is not excluded (see Selli *et alii*, 1978; Ferranti *et alii*, 2008b). However, it is considered still active in its eastern sector (see above), since it is considered responsible of one of the sequence earthquakes on February-March 1783 (Monaco and Tortorici, 2000; Jacques *et alii*, 2001) and of the co-seismic uplifts of the coastal area between Scilla and Palmi area, situated on the footwall of the structure, in the last 5000 years (Ferranti *et alii*, 2007, 2008b). In general, the Calabrian side of the Straits has uplifted more quickly than the Sicilian side, where the elevation of the MIS 5.5 (125 ka) terrace (Bonfiglio and Violanti, 1983) suggests an uplift rate slightly smaller than the regional component.
21st September 2018 – From Reggio Calabria to Siracusa
Mt. Etna and south-eastern Sicily

Mount Etna is a 3,330 m high composite basaltic stratovolcano built up over the past 500 ka on the eastern coast of Sicily in the complex geodynamic setting of the Neogene-Quaternary convergence between the African and European plates (Ben Avraham et alii, 1990; Gvirtzman and Nur, 1999; Doglioni et alii, 2001; Branca et alii, 2011). The volcano lies at the front of the Sicilian-Maghrebian thrust belt and on Early-Middle Pleistocene foredeep clayey successions (Fig. 67) deposited on the flexured margin of the foreland that is the Pelagian block (Branca et alii, 2011). This succession is currently deformed by detachment folds related to the recent frontal migration of the thrust belt, as a response to the approximately N-S compressive regional tectonic regime (Bousquet et alii, 1987; Labaume et alii, 1990; Monaco et alii, 1997; Bousquet and Lanzafame, 2004; De Guidi et alii, 2015). Since the Middle Pleistocene, contractional structures of the orogen have been coupled with oblique extensional faults along the Ionian offshore. These faults form a lithospheric boundary extending between the Aeolian Islands and the eastern Sicily offshore, including the Malta Escarpment and the Alfeo-Etna Fault System, characterized by strong seismicity and active volcanism (Lanzafame and Bousquet, 1997; Palano et alii, 2012; Gutscher et alii, 2015; Polonia et alii, 2016).

The current morphostructural framework of the volcanic edifice is the result of a complex interaction of magmatic processes with regional tectonics and flank instability processes (Monaco et alii, 2010; De Guidi et alii, 2012; Azzaro et alii, 2013). The lower eastern flank of Mt. Etna is characterized by several morphological scarps (locally known as Timpe) which are the result of Late Quaternary normal faulting (Timpe fault system) (Monaco et alii, 1997, 2010; Azzaro et alii, 2012). The Acireale master right-normal fault morphologically controls a 10 km-long coastal stretch forming an up to 150 m high cliff and is characterized by a vertical slip rate of about 4 mm/yr in the past 35 ka (Azzaro et alii, 2012). It represents one of the northernmost segments of the offshore seismonogenic lithospheric boundary responsible for the historical destructive earthquakes (Bianca et alii, 1999; Argnani and Bonazzi, 2005; Palano et alii, 2012; Gutscher et alii, 2015; Polonia et alii, 2016). Accordingly, the coastal sector of eastern Sicily has been affected by a vigorous rift-driven tectonic uplift whose Upper Pleistocene rate, progressively decreasing southwards, was estimated by the measurement of the altitude of the MIS 5e marker (Ferranti et alii, 2006 and references therein). In particular, for the Mt. Etna area, long-term uplift rates of 1.3 mm/yr have been estimated. As regards the Holocene relative sea level change, the nominal elevation of measured paleo-shorelines allowed to estimate uplift rates of 2.5-3.0 mm/yr for the last 6-7 ka (Branca et alii, 2014).

Conversely, the Pleistocene foredeep clayey succession and the overlying coastal-alluvial deposits (locally named Terreforti), located along the southern margin of the volcano have been involved in regional tectonic uplift at a rate of 1.2 mm/yr, forming in the last 240 ka a series of terraces, that, in turn, have been involved in the recent compressive deformation (Bousquet and Lanzafame, 1986; Ristuccia et alii, 2013). This produced a

Figure 67. A) Tectonic sketch-map of eastern Sicily and Ionian offshore. The Mt. Etna volcano (red polygon) lies mainly atop the Sicilian-Maghrebian Collision Zone, a fold and thrust belt sandwiched between the Peloritani Block in the North and the Hyblean Foreland in the south. Major tectonic boundaries in the Ionian Sea (e.g. the Malta Escarpment and the Alfeo-Etna Fault system appear to join in the Mt. Etna area. B) Structural framework of the Mt. Etna eastern and southern flanks showing a complicate deformation field resulting from the combination of several processes spanning from volcano-tectonic, sliding dynamics and regional tectonics (from Barreca et alii, 2018).
roughly 10 km long and W-E trending asymmetric south-facing anticline at the front of the chain (Terreforti anticline, Fig. 67B, Labaume et alii, 1990; Monaco et alii, 1997). Recent SAR and PSInSAR analyses (Lundgren et alii, 2004; Bonforte et alii, 2011) have highlighted the current growth of another large W-E oriented anticline in the north-western outskirts of Catania and in the Aci Trezza offshore (the Catania anticline, Fig. 67B; De Guidi et alii, 2013; Barreca et alii, 2018). This is characterized by maximum uplift rate of about 10 mm/yr along the hinge zone, and has been attributed either to gravitational spreading of the volcanic edifice (Lundgren et alii, 2004; Bonforte et alii, 2011) or to tectonic convergence (De Guidi et alii, 2013; Barreca et alii, 2018). The folding and the related N-S compression that produced the Terreforti anticline are older than the building of the main Mt. Etna edifice, so contractual structures south of Mt. Etna can be interpreted as the detachment response of a shallow thrust migrating within the foredeeps deposits at the chain front (Bousquet et alii, 1987; Labaume et alii, 1990; Monaco et alii, 1997; Bousquet and Lanzafame, 2004; De Guidi et alii, 2015).

South-eastern Sicily is characterized by thick Mesozoic to Quaternary carbonate sequences and volcanics forming the emerged foreland of the Sicilian-Maghrebian thrust belt (Grasso and Lentini, 1982). This area, mostly constituted by the Hyblean Plateau (Fig. 67A), is located on the footwall of a large normal fault system which since the Middle Pleistocene has reactivated the Malta Escarpment (Bianca et alii, 1999), a Mesozoic boundary separating the continental domain from the oceanic crust of the Ionian basin (Scandone et alii, 1981; Sartori et alii, 1991; Hirn et alii, 1997). In this area the vertical component of deformation has been recorded by several orders of Middle-Upper Quaternary marine terraces and palaeshorelines (Di Grande and Raimondo, 1982), which indicate a long-term uplift rate of about 0.5 mm/yr (Bianca et alii, 1999). This uplift rate gradually decreases towards the stable areas of the south-eastern corner of Sicily (Antonioli et alii, 2006; Ferranti et alii, 2006). Similar short-term values of vertical uplift have been determined through archaeological and geomorphological markers along the coast of south-eastern Sicily (Scicchitano et alii, 2008; Dutton et alii, 2009).

The uplifted Taormina coast

Geological framework

The Taormina region is characterized by thick Mesozoic to Quaternary carbonate sequences and volcanics forming the emerged foreland of the Sicilian-Maghrebian thrust belt (Grasso and Lentini, 1982). A contribution from regional and fault-related sources to uplift has been invoked to explain the development of several orders of Middle-Late Pleistocene marine terraces and Holocene paleoshorelines carved on Hercynian metamorphic rocks and Mesozoic limestone along the Taormina coast (Bonfiglio, 1981; Bonfiglio and Violanti, 1983; Firth et alii, 1996; Stewart et alii, 1997; Rust and Kershaw, 2000; Catalano and De Guidi, 2003; De Guidi et alii, 2003; Antonioli et alii, 2003; 2006; Spampinato et alii, 2012). According to Catalano and De Guidi (2003), long term uplift occurred with rates as high as 1.7 mm/yr, although Antonioli et alii (2006) estimated a more conservative rate of 1.1 mm/yr in the last 125 ka. As regards the short-term span, detailed morphotectonic analysis provided uplift rate of ~1.8 mm/yr in the last 5 ka, almost equally balanced between the steady and stick-slip components (due to three co-seismic uplift events, De Guidi et alii, 2003; Spampinato et alii, 2012). Abrupt displacements were attributed to co- and post-seismic flexural uplift of the footwall of an extensional structure, the Taormina fault that is inferred to run immediately offshore the studied coastal outcrops (Fig. 68).

Even though large (M~7) and destructive earthquakes during the last centuries can be associated to normal fault activity along south-western Calabria and south-eastern Sicily (Monaco and Tortorici, 2000), seismic historical data for the Taormina region (Mercalli, 1897; Baratta, 1901; Postpischl, 1985; Boschi et alii, 1995, 1997; CPTI Working Group, 2015) indicate a very low-level of seismicity since the old Greek civilization (VIII century B.C.). The area corresponding to the inferred Taormina fault is affected by low magnitude instrumental seismicity (M < 3.5; Azzaro et alii, 2006), mostly located offshore, with some stronger events which were are also felt along the coast (e.g. the March 28, 1780 event, Io = VII–VIII MCS, Mw = 5.6; Azzaro et alii, 2007).

Figure 68. Morphotectonic map of the coastal area between Capo St. Alessio and Capo Schisò (from Spampinato et alii, 2012). Late Quaternary marine terraces are reported from Antonioli et alii (2006).
STOP 11 – The Capo Schisò uplifted shorelines (Lat. 37°49’21”N – Long. 15°16’34”E)

Detailed mapping of geomorphological, biological and archaeological sea-level markers around the Capo Schisò volcanic headland, a few kilometres south of Taormina, north-eastern Sicily, has documented the occurrence of three Holocene paleo-shorelines raised at different altitudes. The uppermost shoreline (PS1) is represented by a fossiliferous beach deposit that is heavily eroded and only few small sections, at elevations ranging between ~3 and ~5 m above the present sea-level, are visible. The middle shoreline (PS2) was found at a maximum altitude of ~3 m and is represented by algal rims, remnants of barnacle bands and vermetid concretions, and by a fossiliferous beach deposit. The lowermost shoreline (PS3) includes remnants of algal rims, vermetid concretions, fossil barnacle bands and a beach rock, and reaches an elevation of 1.60–1.80 m.

Radiocarbon dating results, integrated with published ages from nearby paleo-shoreline outcrops, constrains for the Taormina region an average uplift rate of 1.7–1.8 mm/yr in the last 5 ka, and the occurrence of three coseismic uplift events at 4.4–3.9 ka, 2.1–1.8 ka and ≤1.0 ka ago (Spampinato et alii, 2012). Abrupt displacements are tentatively attributed to footwall uplift along an offshore normal fault. The analysis of historical seismicity reveals that a few earthquakes occurred in this area during the time span of the uplift events detected by our coastal investigation. In addition, the first event is too old; therefore it cannot be found within the historical seismicity catalogue (CPTI Working Group, 2004). The ~2.0 ka age of event II has two possible equivalents in the historical catalogue. The closest matches to this event are the 91 BC (~2.1 ka) earthquake and the 17 AD (~2.0 ka) earthquake.

![Figure 69. Ubicación of Capo Schisò respect to the famous area of Taormina.](image)

Of these, the 17 AD event has an estimated magnitude of ~5.1 and thus can be rejected. The 91 BC event has an equivalent magnitude of 6.3, and is compatible with our estimation. The macroseismic epicenter of the event, however, is located on the Calabria coast opposite to and north of Taormina. By considering the age of the event and the inherent uncertainty in macroseismic location of the causative structure, the possibility that the catalogue earthquake is the same than event II at the Taormina coast should not be neglected. Finally, seismic event III (~1.0 ka) could be referred to the 853 AD earthquake in northeastern Sicily (Boschi et alii, 1995; see also De Guidi et alii, 2003). Although, as discussed above, a large uncertainty exists on the duration of PS3 and thus on the age of event III, it is conceivable that this event is not much younger than ~1.0 ka B.P.; otherwise it should probably have been recorded in the historical seismicity catalogue. As for the 91 BC event, the macroseismic epicentre of the 853AD event, although located along the Sicilian coast, lies to the north around the city of Messina. However, we note that the equivalent magnitude of the 853 AD event is similar to that of the 91 BC event (CPTI Working Group, 2004). Because our record of the slip for both events is also similar, we tentatively propose that both events were caused by the structure responsible for coastal co-seismic uplifts at the Taormina coast.

![Figure 70. Fossil remains at Capo Schisò.](image)
STOP 12 - Raised paleo-sea level markers in the Aci Trezza area (Lat. 37°33'34"N – Long. 15°09'38"E)

Geological framework

Holocene relative sea level change and vertical deformation have also been evidenced in the southernmost coastal sectors of Mt. Etna volcano (Fig. 71) by several paleo sea level markers (Firth et alii, 1996; Branca et alii, 2014). This vertical deformation has been related to distinct sources: tectonic regional uplift, volcanic dome effect, local deformation along faults and folds (De Guidi et alii, 2015; Branca et alii, 2014). In the Aci Trezza area the uplifting has undergone an acceleration along the hinge of the active fold (the Catania anticline, Fig. 71). To the north, the Acireale master right-normal fault (Fig. 71) morphologically controls a 10 km-long coastal stretch forming an up to 150 m high cliff and is characterized by a vertical slip rate of 2-4 mm/yr in the past 35 ka (Monaco et alii, 1997; Azzaro et alii, 2012).

The general process of uplifting is only locally interrupted by subsidence related to flank sliding of the eastern sector of the volcanic edifice (see Azzaro et alii, 2013 for a complete review), measured at docks and other manmade structures, characterised by complex interaction with fault systems. GPS (Bonforte and Puglisi, 2006) and PSInSAR data indicate a strong subsidence due to flank sliding of the triangular coastal sector located between Stazzo and Riposto villages (Giarre wedge of Bonforte et alii, 2011), that is bounded westwards by the San Leonardello normal fault (Fig. 71). Aseismic slip with vertical rates of about 1.0 cm/yr, measured in the last 20 years along the southern sector of this fault (Monaco et alii, 2010; Carveni et alii, 2011) is probably related to the subsidence and sliding of the coastal sector to the east.

Description of the stop

Morphological and biological markers of raised Holocene shorelines, which are represented by beach rocks, wave-cut platforms, balanid, vermetid and algal rims, provided precise time-space constraints for the dynamics of the lower south-eastern flank of Mt. Etna volcano (Branca et alii, 2014). The timing of coastal uplift has been determined by radiocarbon dating of shells collected a different elevations from the raised paleo-shorelines and, to correctly assess the total amount of tectonic uplift of the coast during the Late Holocene, the elevation-age data of sampled shells have been compared to the local curve of Holocene sea-level rise. In particular, shells of Lithophaga sampled between 1.55 to 6 m a.s.l. within a serpentid and algal reef encrusting the 500 ka old basalt pinnacles of the Ciclop Island (offshore of Aci Trezza village, Figg. 72c-73a), yielded calibrated AMS ages between 1.8 and 6.0 ka BP (Firth et alii, 1996). More recently, a Gasteropod collected at the top of the reef (6 m above sea level) yielded a calibrated age of 6316.5 ± 130.5 (Branca et alii, 2014). These data have large uncertainty since the Lithophaga holes are not related to a notch and the paleodepth cannot be accurately estimated (Lithophaga presently lives in the Mediterranean from 0 to -20 m). However, taking into account the age of the top shells and the predicted sea level at that time (-9.3 m; Lambeck et alii, 2004), an uplift rate of about 2.5 mm/yr has been estimated.

Even though volcanics along Mt. Etna coastline usually do not preserve tidal notches, indication of recent uplift are provided also by the two wave-cut platforms clearly carved on the 500 ka old hyaloclastic pillow-brecia and pillow lavas, at the bottom of the impressive Aci Castello rock, south of Aci Trezza (Figg. 72c-73b).

A small exposure of the higher platform can be observed only in the southernmost side of the rock where it is surmounted, at an elevation of 4.0-4.5 m a.s., by a 15–3.9 ka old lava flow (Branca et alii, 2011). The lower platform is very large and surrounds the entire rock, extending from ~2.0 m a.s.l. down to 0.9 m above b.m.s.l. Remnants of calcareous reef deposits, bored by Lithophaga, encrust the southern portion of the lower platform.
Figure 72. (a) Geometry and distribution of detected seismic units along the sparker line L1 (see c for location); the occurrence of the seaward extension of the Catania anticline and of the Acireale fault (see Fig. 71) to the northeast is evident. (b) A seismo-stratigraphic model reconstructed from seismic facies analysis; and (c) correlation of the seismic units with age-constrained volcanics and sedimentary deposits outcropping on land (age of volcanics from Branca et alii, 2011). S1, S2 and S3 are erosional surfaces; SBM: sea bottom multiple (from Barreca et alii, 2018).

A sample of this deposit was collected at an elevation of 1.9 m a.s.l. and its radiocarbon dating yielded an age of 6750 ± 130 cal BP (Branca et alii, 2014). In particular, the Acireale and the San Leonardello master faults divide the coastal sector in two main domains characterized by different vertical movements: the northern coastal sector located in the hanging wall, corresponding to the Giarre wedge delimited northward by the termination of the Pernicana fault, and the south erosional sector, developed on the footwall blocks (Fig. 71). The northern coastal sector shows historical and current subsidence, superimposed on the long-term (since 20 ka) uplift recorded by the LGM platform, that concurs with the seaward sliding dynamics of this block of the volcano edifice (Bonforte et alii, 2011). Here, the subsidence related to the sliding of the Giarre wedge is prevalent with respect to the long-term regional uplift. On the contrary, the southern As a whole, this coastal sector, including an LGM platform (Fig. 71), is characterized by long- (since 20 ka) and short-term uplift that is in contrast with the general seaward sliding dynamics of the eastern flank of Mt. Etna volcano. In particular, the GPS and interferometry data (Bonforte and Puglisi, 2006; Bonforte et alii, 2011) reveal a fairly stable behaviour of the coastal sector between S. Tecla and Capo Mulini, at the footwall of the Acireale fault (Fig. 71), providing evidence that the subsidence related to the sliding process compensates for the regional and fault-related uplift. Conversely, between Capo Mulini and Catania we observe a general uplift of the coast during the last 20 ka, highlighting an acceleration of the vertical motion at the hinge of the Catania anticline, in contrast with the seaward sliding (Fig. 71). The seaward extension of both the Catania anticline and the Acireale fault system is clearly visible in the seismic profile of Figure 73 (Barreca et alii, 2018).

Figure 73. a) Aci Trezza, Holocene serpulid and algal reef encrusting 500 ka old basalts. b) Wave-cut platforms on the 500 ka old pillow lavas forming the Aci Castello rock.
STOP 13 - The terraced sequence of Catania-Monte Po (Lat. 37°29’28’’N – Long. 15°02’15’’E)

Geological Framework

The area located between the southern edge of the Mt. Etna volcanic edifice and the Catania Plain, known as the “Terreforti” (Fig. 74), is characterized by a belt of terraced hills with elevation up to 325 m a.s.l. The terraced sequence lies on an Upper Pliocene to Quaternary sedimentary succession including a number of volcanic intercalations. This succession represents the sedimentary cover of the Sicilian-Maghrebian orogenic front overthrusting the Hyblean foreland domain, in response of the continental collision between the African and the European plate margins (Lentini, 1982; Labaume et alii, 1990). In such a geodynamic context, the Catania Plain represents the remnant of a foredeep domain, filled by Pliocene-Pleistocene sediments and volcanics, and by the Holocene alluvial-coastal deposits of the Simeto River (Longhitano and Colella, 2001). The sedimentary succession of the Terreforti hills is deformed by an asymmetric south-facing anticline, about 10 km long and ~W-E trending (the “Terreforti anticline”). It has been interpreted as a thrust propagation fold at the front of the orogenic system, related to the lately migration of the orogenic thrust belt, as a response to the regional N-S compressive tectonic regime (Monaco et alii, 1997; Labaume et alii, 1990), GPS velocity fields (Mattia et alii, 2012; Palano et alii, 2012; De Guidi et alii, 2015) seismological (Lavecchia et alii, 2007) and interferometric synthetic aperture radar data (Lundgren et alii, 2004; Bonforte et alii, 2011) suggest that contractional processes are still active.

Since the Middle Pleistocene, contractional structures have coexisted with extensional tectonics, as suggested by the occurrence of NNW-SSE to NW-SE striking oblique (normal-dextral) fault systems running offshore, where the Malta Escarpment (Bianca et alii, 1999; Monaco and Tortorici 2000, 2007; Palano et alii, 2012) and the North-Alfeo belt (Gutscher et alii, 2015; Polonia et alii, 2016) occur. Accordingly, the overall orogenic belt of eastern Sicily has been affected by a rift-driven vigorous tectonic uplift, with rates progressively decreasing southwards (Bordoni and Valensise, 1998; Ferranti et alii, 2006). In the Catania Plain area, this process has been also active during the Holocene with rates of about 1 mm/yr (Spampinato et alii, 2011). Uplift process, combined with relative sea-level change, has caused the terracing of fluvial and coastal deposits of this area (Monaco, 1997; Monaco et alii, 2002; Ristuccia et alii, 2013). Current tectonic activity is also responsible for the destructive historical earthquakes (M≥7) that occurred in south-eastern Sicily (e.g. 1169 AD, 1693 AD events, Baratta, 1901; Postpischl, 1985; Boschi et alii, 1995; Rovida et alii., 2011; Panzera et alii, 2011; D’Amico et alii, 2011). The location of seismogenic sources is a topic still widely debated: normal faults located along the Ionian offshore and/or compressional structures located to the north and to the south of the Catania Plain, between the front of the Appenninic-Maghrebian Chain and the northern margin of the Hyblean foreland (see DISS Working Group, 2016).

Figure 74. (A) Map of the terraced deposits. (B-C) Profiles across the different orders terraced surfaces. (D) Particular of deformed terrace T6 (from Ristuccia et alii, 2013).
The sequence of marine terraces

On the northern slope of the Monte Po ridge, along the municipal road connecting the Monte Po and the San Giorgio quarters of Catania city, the entire Pleistocene sedimentary succession outcrops (CtU1-CtU3 depositional units, Fig. 75), dipping at about 15° southwards. The oldest unit (CtU1) is represented by the pre-Etnean clays of Early Pleistocene age, up to 600 m thick, interpreted as the result of steady sedimentation in a relatively shallow marine gulf. The marly clays are overlain by gravels and sands (CtU2), up to 50 m thick and highlighted by the breaking of the slope, traditionally attributed to the Middle Pleistocene, interpreted as the product of coastal shallow marine sedimentation. In this area, there are evidences of the Etnean fissural early activity that originated products with tholeiitic-transitional affinity (pre-Etnean volcanism). These products, coeval or subsequent to the deposition of the CtU1 and CtU2 are mostly represented by the 320-250 ka old lava plateau (Gillot et alii, 1994) cropping out west of Catania in the Paternò area (Fig. 74). The third depositional sequence (CtU3) is represented by Middle-Upper Pleistocene marine shallow-water to alluvial polygenic conglomerates and coarse sands, forming massive or indistinctly stratified bodies, up to few tens of meters thick. Sediments of CtU3 have been mapped as terraced deposits belonging to seven orders, extending west of Catania (Fig. 74). It is noteworthy a good west-east lateral continuity of this alluvial-coastal terraces, especially the lowest ones (1st to 4th orders), and a limited extension of the 5th order terraces, restricted to the western sector. Terraces generally show a sub-planar southward sloping surfaces (3°-5°), with the exception of the 6th and 7th ones that are folded (Fig. 74). In particular, the 6th order surface forms a large anticline with an axis coincident with that of the main fold of the chain front (the Terreforti anticline).

The younger unit (Ct_U4) is represented by up to 60 m thick of alluvial deposits related to the depositional activity of the Simeto River, interpreted as the result of Holocene to Present high stand sedimentation.

At the top of the hill (Stop 1, altitude 210 m a.s.l.) it is possible to clearly observe the depositional characteristics of the oldest terraced deposits of the CtU3 unit outcropping in the Catania area (6th order) that extend along the south-western edge of the urban area, at an altitude between 220 and 150 m. These are represented by fluvial-coastal polygenic conglomerates with a sandy matrix and with indistinct stratification, about 20 m thick.

The conglomerate consists mainly of metamorphic and sedimentary clasts, coming from the tectonic units of the Sicilian-Maghrebian chain, and less than 5% from volcanic clasts of pre-Etnean origin. OSL age determinations of alluvial-coastal terraced deposits cropping out between Mt. Etna volcano and the Catania Plain (Ristuccia et alii, 2013) provided good constraints for correlating the seven orders of terraces with as many sea-level high-stands each corresponding to a distinct phase of the global eustatic curve (i.e. MIS 9.3, 7.5, 7.1, 5.5, 5.3, 5.1, 3.3; Waelbroeck et alii, 2002). In particular, the 6th order terrace has been correlated to MIS 7.5 (236 ka), the 4th order terrace to MIS 5.5 (124 ka), the 3rd order terrace to MIS 5.3 (100 ka) and the 2nd order terrace to MIS 5.1 (80 ka). It is worth noting that the obtained OSL dating is in agreement with the time assignment of the same deposits suggested by Monaco et alii (2002), based on indirect dating of the distinct terraces through the chronological attribution of volcanic clasts included in conglomeratic levels. In particular, only volcanic clasts belonging to the 320-250 ka old pre-Etnean activity have been found in the 6th order deposits. Despite the indirect chronological attribution of the 1st order terrace, it is worth to note that the investigated area represents one of the few in southern Italy documenting MIS 3 deposits currently outcropping, while they are often intercepted in wells (see Ferranti et alii, 2006, 2009).
A mean uplift rate of 1.2 mm/yr during the last 330 ka has been calculated, probably related to regional processes. Moreover, the geometry of terraces indicates that local tectonic processes, related to the migration of the front of the Sicilian orogen, coupled with sea-level changes during the Middle Pleistocene. The two oldest terraces are in fact folded, suggesting activity of the frontal thrust between 236 and 197 ka. The morphological analysis of the terraces and the deformation pattern also suggest the occurrence of an important tectonic component in the uplifting of the oldest terraces (6th and 7th orders) which is related to Middle-Upper Pleistocene contractional processes at the front of the Sicilian chain. Despite the inner edge of the 6th order terraces is located at 270 m a.s.l., this surface reaches an elevation of 325 m a.s.l. along the axial zone of the Terreforti anticline. Taking into account that this terrace formed between 236 ka and 197 ka (this latter is the age of the first subsequent unfolded terrace), a fold growth rate of 1.4 mm/yr along the hinge can be estimated.

Observing the panorama towards the Catania Plain we note the almost complete sequence of the lower-order terraces, showing sub-horizontal morphology, with a slight dip towards the south. The Stop 2 (altitude 120 m a.s.l.) is devoted to the observation of the sedimentological features of the 4th order terraced deposits, attributed to MIS 5.5 (124 ka). Cross-bedded sands are here intercalated in thick conglomeratic banks (Fig. 76b) containing also clasts belonging to the early Etnean alkaline activity (Timpe volcanics, 200-100 ka old).
STOP 14 – The archaeological site of Megara Iblea (Lat. 37°12′18″N – Long. 15°10′59″E)

Megara Hyblaia is an ancient Greek colony built alongside a large Quaternary calcarenite plateau facing the Augusta Gulf, at an elevation of 10–15 m above sea level. It is located inside the modern Augusta harbour between two rivers, the Cantera to the north and the San Cusumano to the south. The most significant archeological marker is a submerged stone structure, at about 4 m off the present coastline near the northeastern corner of the plateau, previously observed by foreign visitors who was able to recognize it during an exceptional low-tide episode. This stone structure ("banchinamento" readable as "quay building"), later reconsidered by Villard and Vallet (1953) and Gras (1995), was interpreted as a harbour pier because of location, block typology, and building technique. It is 24.50 m long and 5.30 m wide and it was built in the so called “Greek style" technique that is typical of landing or military structures of the Greek world. This technique is characterised by the use of large parallelepipied calcarenite blocks, as long as 1 m and without any joins or transversal blocks, arranged in four overlapping rows: the first one seems to be placed straight on the rocky seabed and no foundation level has been detected; the second one forms a large submerged platform. In the eastern sector, the sea bottom relative to the second row is 1.20 m deep at the pier foot and 0.80 m (corrected height −0.88 m) at the pier head. Although only part of the four rows of blocks is still in place, one might hypothesize the existence of a complete four-row structure; if we consider the top of the mentioned structure and its original functional surface, the palaeo-sea level should be at −1.48 m depth. Further underwater archaeological prospecting carried out along this stretch of the sea floor has revealed the occurrence of an irregular structure, still under study, built by means of several remnants of blocks and rough stones fully covered by seaweeds (Basile, 1995). It is located about 22 m to the north of the “Banchinamento Orsi" at 1.20– 1.40 m depth and has been interpreted as a sort of L-shaped jetty, but it could represent a break water in front of the Cantera river mouth.

Figure 77. Sketch map of the archaeological site of Megara Hyblea; the position of the submerged “Banchinamento Orsi" harbour pier is shown; b) plan and c) cross-section of the “Banchinamento Orsi" harbour pier (from Villard and Vallet, 1953).
The Mediterranean Sea is well known to have a very long record of tsunami inundations (Soloviev et alii, 2000; Maramai et alii, 2014) and since the second millennium BCE (Before Common Era) hundreds of historical accounts report information on tsunami events originated mainly in Eastern Mediterranean Sea, Greece, southern Italy and southwest of Portugal. Moreover, in order to extend back in time the estimates on tsunami magnitude and frequency at a specific location, a dedicated geological approach on the Holocene/late Quaternary coastal sediments looking for paleotsunami deposits distribution, their characteristics and number may help in filling this information gap.

In Italy, this effort started about 20 years ago, well before the 2004 IOT (Indian Ocean Tsunami), along the coast of Apulia (Mastronuzzi and Sansò, 2000; 2004; Mastronuzzi et alii., 2007; Gianfreda et alii, 2001; De Martini et alii, 2003) and later on several studies were done along the coast of Eastern Sicily (see De Martini et alii, 2012 for an overview).

In this contribution we present the results of a multidisciplinary investigation (De Martini et alii, 2010), involving geomorphologic surveys, coring campaigns, laboratory analyses and different dating methods, applied at a coastal site named Priolo Reserve, located in Augusta Bay area (Siracusa province). Additional data on paleotsunami signatures are available from a site offshore the Augusta Bay (Smedile et alii, 2011) as well as in a following paper by Smedile et alii, 2012, where a combination of inland and offshore data on paleotsunamis found in Augusta Bay area is provided.

**Priolo Reserve site**

We performed aerial photographs and satellite images interpretation along Augusta Bay coastline in order to identify lowlands and lagoon where sedimentation, preservation and dating of tsunami deposits of the past could be favorable. This geomorphologic approach as well as the field survey, was quite difficult because a series of well-known petrochemical facilities and both the NATO and Italian Navy bases nowadays occupy large part of Augusta Bay coastline.

One of the sites selected was the Priolo Reserve area (De Martini et alii, 2010), located in the central-southern part of Augusta Bay coast, exactly in the hearth of the industrial area. Since 2.000 this site has been a protected area, founded and assigned to the LIPU (Italian League for Birds Protection) in order to provide refuge for migratory and resident birds as well as to protect the salt pans where special plants grow in saltmarshes and wetlands (more info at http://www.salinepriolo.it/). Priolo Reserve site is a 0.5 km² shallow lagoon separated from the sea, both from NE and SE, by up to 4.5 m high sand dunes; notably in its northern sector salt pans were installed since Greek times.

We collected 16 cores on which stratigraphic and sedimentological description together with high resolution pictures were performed both directly in the field and in the laboratory. Moreover, in order to highlight even small scale sedimentary features X-ray analysis was done on few selected cores. The maximum depth of the collected cores is 4.2 m at a maximum distance of 530 m from the present shoreline.

Qualitatively and quantitatively analyses were carried out on benthic foraminifera with the aim to reconstruct the paleoenvironmental evolution. Macrofauna was qualitatively analysed in order to study their taphonomic and morphoscopic characteristics. The chronology of the investigated stratigraphy is based on both radiocarbon (3 samples, 2 shells belonging to the *Cerastoderma glaucum* species and 1 shell of *Pirenella conica*, see Table 2 in De Martini et alii, 2010 for details) and tephracronology (6 samples, see Table 4 in De Martini et alii, 2010 for details) dating methods.

In Priolo Reserve site the collected sequence shows a not uniform stratigraphy apart from a 1-3 cm thick layer of black fine volcanic ash found in almost all the cores close to the surface being related to the 2002-03...
Etna volcano eruption. In general, important sedimentological changes are common moving from the central part of the lagoon, where clay to silty clay deposits dominate, towards south and east, where the sand of the dunes clearly influence the local sedimentation, resulting in about 1.5 m of brownish fine to medium well sorted sand close to the surface, interpreted as eolian deposits. The most interesting sector is located in the southwestern part of the reserve (Fig. 78), where the uppermost 1.5 m of the stratigraphy is made by gray to brown silty clay and clayey layers with vegetal remains, bivalve, gastropods and shell fragments. This fine sequence (Fig. 79) is interrupted by two distinctive bioclastic layers (PR01 and PR02 at 0.3 and 0.5 m in depth, respectively), by one peculiar detritic deposits (at 0.9 m in depth, PR03) and finally by a clear blackish volcanic ash layer (at 0.7 m in depth). Moreover, a distinct sandy layer (PR04) was found at about 1.6 m in depth within a 0.7 m thick deposit dominated by gray clay to silty clay with small and scarce shell fragments.

The two 5 to 10 cm thick bioclastic units (PR01 and PR02) show abnormal concentration of shell fragments together with entire gastropods, unusual chaotic pattern, no grading and a sharp lower contact potentially erosional, as observable from X-ray image. The distinctive 2-3 cm thick detritic layer (PR03) presents an unusual assemblage made by rare ostracods, gastropods and notably by macromammal bone fragments. Finally, the about 3 cm thick sandy layer (PR04) is characterized by abundant shell fragments, thus resulting in a clearly coarser layer with respect to the fine clay to silty clay sediments above and below.

The performed analyses suggest that fine deposits of the whole sedimentary sequence of Priolo Reserve site (Fig. 79) are characterized by molluscs, ostracods and benthic foraminifera all specific of a lagoonal low-energy environment.

Differently, both PR01 and PR02 layers show the presence of a significant amount of planktonic foraminifera and a peculiar increment in the benthic foraminiferal specific diversity with respect to the surroundings fine
deposits; PR03 (the detritic peculiar layer) appears characterized by poorly preserved benthic (Cassidulina carinata, Cibicidoides pseudoungerianus, Melonis barleanum, Planulina ariminensis) and planktonic (Globigerinoides sp.) foraminifera, usually common in open marine environments; in the PR04 sandy layer both benthic (Asteriginata mamilla, Buccella granulata, Cibicides lobatulus, Elphidium spp., Nubecularia lucifuga, Rosalina spp.) and few planktonic (Globigerinoides sp.) foraminifera were found and classified.

**Chronological attribution**

By combining the above data, we may say that the two bioclastic layers (PR01 and PR02) share some peculiar characteristics: coarser grain size with respect to the dominant much finer sediments at the site, sharp basal contact potentially erosional, macro-micropaleontological marine content from the foreshore up to the open sea (differently from all the other finer sediments investigated at the site). In our interpretation, these two layers represent high-energy deposits of clear marine origin (storm or tsunami) deposited within a lagoonal low-energy environment. Interestingly, also for PR03 and PR04, their stratigraphic, sedimentologic and paleontologic characteristics imply a rapid change from lagoonal low-energy condition to high-energy marine deposition, thus supporting again the occurrence of sudden marine inundations. Finally, the presence of macromammal bone fragments in the PR03 deposits was interpreted as potentially related to a backwash wave able to carry on such variety of elements.

In order to try to discriminate between storms and tsunamis at Priolo Reserve site and to highlight the more probable mechanism of deposition for the 4 high-energy deposits of marine origin, we analysed the anemometric and ondametric data for the period 1989-2006 and we found that the strongest recorded storms in Augusta Bay area had a maximum inundation distance of about 55 m from the shoreline (see De Martini et alii, 2010 for details). If we consider the PR01, PR02, PR03 and PR04 distribution within the site we may easily understand that these deposits are well beyond the maximum storm flooding distance, thus supporting the tsunami origin as the most probable.

By integrating the available chronological constraints (notably all ages respect the stratigraphic order) derived from 3 radiocarbon ages and 6 samples of a distinctive plinian Etna eruption tephra dated back to 122 BCE (on the latter we derived a sedimentation rate of 0.33 mm/yr.), we may suggest that the uppermost 1.6-1.8 m of the studied stratigraphy is as old as 4000 years and in more detail the time of the marine inundations: PR01 should be slightly younger than 1420-1690 CE (the 1693 CE local tsunami being the best candidate), PR02 is close to 160-320 CE (with a possible link to the 365 CE Crete earthquake and tsunami), PR03 deposited around 800-600 BCE and finally PR04 is close to 2100-1630 BCE (potentially related to the Late Minoan Santorini event).
22th September 2018 – From Siracusa to Catania

Stop 16 - Penisola Maddalena
Stop 17 - Ognina
Stop 18 - Marzamemi
Stop 19 - Pantano Morgella
The Ionian coast of south-eastern Sicily, between the towns of Augusta and Siracusa, is characterized by the occurrence of anomalous calcareous boulders. They are mostly scattered along large terraces located 2–5 m above sea level, gently sloping towards the sea. Boulders are up to 80 t in weight and are arranged either in isolated elements or small groups composed of a few stacked elements. Several boulders show biogenic encrustations (serpulids, balanids, lithophaga) all over their surface which suggest that they were dragged from the mid-sublittoral zone. Direct observations on each boulder (distance from the shoreline, size and weight), together with statistical analysis of the storm regime of the area, allowed to operate hydrodynamic estimations useful to verify if tsunami or storm waves were responsible for their detachment and transport, while radiocarbon age determinations on marine organisms constrained the timing (Scicchitano et alii, 2008; 2012). Collected data, compared to historical catalogues, suggest that in the last 1000 years three seismic events with local sources could have triggered tsunami waves associated with the boulder deposits occurring in the area. The first two were probably triggered by the earthquakes of February 4, 1169 and January 11, 1693 which destroyed south-eastern Sicily. The third tsunami was probably generated by the strong earthquake which took place in the Strait of Messina on December 28, 1908. The major boulder accumulation has been found along the north-eastern coast of the Maddalena peninsula, south of the Siracusa natural harbour. The peninsula is a calcareous semi-horst gently tilted to the ENE, formed by Miocene sediments that along the coast are unconformably covered by Pleistocene calcarenites (unitweight = 2.28 g/cm³). Most boulders are up to 40 t in weight and are scattered at a distance of up to 70 m from the coast on a large terrace located 5 m above sea level, gently sloping towards the sea. The shoreline is marked by a steep cliff down to −5 m; the near-shore sea bottom topography is irregular, being characterized by two steps at −6/−10 m and at −10/−32 m. The blocks are mostly arranged in isolated elements and have also been found on a flat anthropogenic platform at 1–2 m a.s.l. inside an ancient Greek quarry located along the coast. Locally, small groups composed of stacked elements occur. In this site a very large boulder, characterized by a volume of 80 m³ and weight of 80 t, has been found at a distance of 45 m from the coast. According to our observations, during the winter 2006 the boulder B6 was removed inland for about 3 m by storm waves. As regards the distribution of the A-axis direction, it is more dispersed around NNE–SSW and SSE–NNW orientations. Samples of serpulids and balanids were collected from six distinct blocks and their age determined. The detailed morphological analysis of blocks revealed the occurrence of exposed fracture surfaces on at least two sides of boulders which are characterized by balanid and serpulid encrustations on the other sides. This suggests that they were carved out from the shoreline platform edge (midlittoral zone), along strata planes and joints. Some boulders show also karstic rock pools on the upper side which suggests that it was carved out and dragged from the mid-supralittoral zone. However, most of the boulders are smoothed and completely encrusted by marine organisms living in the mid-infra littoral zone as serpulids, balanids and lithophaga. All together, these features indicate a submerged pre-transport setting. The location of the boulders inside an ancient Greek quarry suggests that they could be relics of quarry operation, even though their morphological features and biogenic encrustations are indicative of a submerged pre-transport setting.
Figure 81. Position of boulders surveyed at the Maddalena Peninsula and submerged profile showing the near-shore sea bottom topography.

Figure 82. Laser Scanner survey of isolated boulder in Maddalena Peninsula.

Figure 83. Laser Scanner survey of boulder accumulation in Maddalena Peninsula.

Figure 84. Laser Scanner survey of isolated boulder in Maddalena Peninsula.

Figure 85. Laser Scanner survey of boulder accumulation inside ancient quarry in Maddalena Peninsula.
STOP 17 – Sea level and tsunami in the archaeological site of Ognina (Lat. 36°58′48″N – Long. 15°15′25″E)

The Ognina area is located 10 km south of Siracusa and is formed by two small promontories. The coast is mostly characterized by rocky platforms, carved on Pleistocene calcarenites, placed between 3 and 0.5 m a.s.l. and gently sloping seaward. The archaeological site was mostly located on a former small peninsula (the main part is the Ognina island), connected to the mainland by a narrow rocky isthmus, which is now submerged. It is a complex site with remnants of several archaeological phases, spanning the period from Neolithic to Byzantine. The positive structures survive to a height of only 3 m on the top of the islet, whereas negative structures (carved straight out of the rock) endure both on the island as well as on the coastal mainland. Some of the archaeological evidence found at Ognina in a partially modified position with respect to the time of use has already been discussed in considerable detail as a signal for sealevel change (Kapitaen, 1970; Castagnino, 1993; 1995; Basile et alii, 1988). The submerged rocky isthmus provided in the antiquity the settlement with sheltered leeward anchorages and beaching places, and it is still in place, from −0.20 m down to −3.30 m relative to the present sea level. On the tiny offshore island of Ognina a series of post-hole structures arranged in parallel alignments suggest the presence of a settlement established during the Neolithic Age, while a stable Maltese trading center (Parker, 1980) flourished during the Bronze Age (3.8–3.2 ka). Close relationships with Malta are suggested by certain vessels which are matched with Tarxien (3.8–3.4 ka) and Borg in Nadur (3.4–3.2 ka) cultures and which reflect a series of long-distance contacts within an organised system of maritime trade. Material evidence seems to suggest that during the Bronze Age the island of Ognina was a genuine trading post under the control of Malta. In the western sector of the islet, a partially submerged Bronze Age tomb of the rock-cut chamber type is carved in the calcarenites. This chamber is preceded by a long dromos with an elliptical opening, the floor of which is at −1.20 m (corrected height −1.21 m) below the present sea level. Taking into account the functional height, the palaeo-sea level should have been at ≤−1.81 m depth. Further meaningful indicators of sea-level change come from the adjacent coastal mainland and are located along the channel as well as along the coast southward of Capo Ognina. Alongside the channel, several partially submerged bollards are carved into the rock. Below the sea surface a bollard has been detected which forms a small artificial mushroom shape, the foot of that is 0.9 m below the present sea level. The sea bottom inside the channel has been detected at maximum −3.00 m (−2.97 m corrected for tide and pressure; Table 1). Taking into account at least 1.0 m of ship draught (Castagnino Berlinghieri, 2003), the palaeo-sea level should have been ≥−1.97 m depth. These two last data points are very significant if we relate them with the Bronze Age (3.8–3.2 ka) activity at the site by the Maltese and other seafaring people, as attested by archaeological evidence. It is worthwhile to note that the channel extends to the east where it is completely submerged, with bottom reaching a depth of −13 m b.s.l.. Along the edge of the channel there are tracks (carraie), previously discussed by Castagnino (1993;1995), that show clear sign of erosion and part of which are collapsed by the southern side of the channel border. Although shapeless shards of amphorae and common ware have been recovered from the submerged channel, it is rather difficult to assess the chronologic range of use, but it seems feasible to surmise that this road system was built to support the intense activity along the channel. Stone quarries are in fact located both on the north and on the south side of the present channel mouth; these are of uncertain age and currently partially submerged with floor located at maximum depths of 0.30 m below sea level. In addition, a partially submerged furnace of uncertain age has been found south of Capo Ognina.

Figure 86. Location of Ognina.
An important tsunami deposit is located on the edge of a coastal embayment incised within Miocene limestones. These deposits fill the back edge of a ria incised within Miocene limestones and are composed of three main stratigraphic units characterized by distinct sedimentological features. The two lower units, formed by cross-bedded sands and laminated clays, recorded the development of a small confined beach-barrier depositional system, influenced by frequent high-energy events. The upper unit, represented by chaotic coarser sediments, can be attributed to a destructive marine high-energy event. The physical properties of the composing stratigraphic units and the morphological setting of the study area, together with sedimentological and palaeoecological observations, accompanied by archaeological determinations and absolute dating, allowed to reconstruct a suite of storm and tsunami-related marine depositional processes that might have occurred in recent times along this area of elevated seismicity. In particular, absolute dating and archaeological determinations allow correlating the upper unit to a tsunami wave triggered by the 1693 AD catastrophic earthquake. The same depositional mechanism can also account for some of the coarse levels occurring into the underlying stratigraphic units.

Figure 87. Sketch map of the archaeological site of Ognina (left); inset shows the islet with the position of the partially submerged tomb; b) cross sections of the emerged (A–B) and submerged (C–D) sectors of the Ognina channel. On the right the sketch map of the tsunami deposit located along the coast of Ognina and down the description of the section.

Figure 88. (a) Picture of the tsunami deposit in Ognina; (b) photomosaic od the deposit; (c) Outcrop sketch showing the stratigraphic units composing the deposit.
Figure 88. (a) Detail of the calcareous bedrock at the innermost termination of stratal unit 3, note red colours sedimentary dikes filling pre-existing fractures organized in centrifugal directions; (b) archaeological remains inside stratal unit 3.

Figure 89. Archaeological remain inside stratal unit 2.
STOP 18- The submerged quarries of Marzamemi (Lat. 36°44'38"N – Long. 15°07'04"E)

Marzamemi is a small village located about 50 Km southward from Siracusa. In this area was found the biggest quarry located in south-eastern Sicily with a surface of about 36,800 m² and several levels of exploitation. It is carved into a marine terrace of Tyrrenian age (Lena et alii, 1988; Lena and Bongiovanni, 2004). The quarry shows evidences of different exploitation techniques; several imprints carved in the rock testify the extraction of huge blocks that were after subdivided into smaller ones using wedges. The quarry is vertically divided in three distinct levels: the shallower down to -0.4 m depth, the medium down to -1.05 m depth and the deepest down to -1.5 m depth. Various markers are present in the quarry, some of these are directly related to the exploitation, such as floors and cutting lines, and some others are related to the infrastructures linked to the quarry, such as stairs carved in the rock connecting different levels of the quarry. In this quarry 9 points have been measured most of whom are floors; the deepest one is the base of a stairs located in the seaward part of the area. The shallower points range in depth between -0.56 m and -0.76 m below sea level, the deepest lies between -0.95 m and -1.34 m below sea level.

![Map of Marzamemi area](image)

*Figure 90. Location of ancient quarry in Marzamemi area.*
Figure 91. Panoramic view of Marzamemi quarry.

Figure 92. Cut lines depicting a block inside quarry.

Figure 93. Block detached inside the quarry of Marzamemi.
STOP 19 - The tsunami impact evidence of the Pantano Morghella site

In this contribution we present the results of a multidisciplinary investigation (Gerardi et alii, 2012) involving geomorphologic surveys, coring campaigns, laboratory analyses and different dating methods, applied at a coastal site named Pantano Morghella, located in the southeastern tip of Sicily. From the available historical records, we know that at least 10 tsunami inundations (Maramai et alii, 2014) affected the about 240 km long eastern coast of Sicily in the last 2 millennia. We are also aware of the fact that this area represents the emerged portion of the Hyblean foreland of the Apenninic-Maghrebian orogen (Lentini et alii, 2006 and references therein) and it experienced very modest (below 0.2 mm/yr.) long term and Holocene vertical coastal movements (Ferranti et alii, 2006; Antonioli et alii, 2009). Aware of the fact that the deposition and preservation of tsunami deposits are strongly dependent on the geomorphologic setting and available accommodation space, we performed a detailed geomorphic study by means of aerial-photographs and satellite images interpretation coupled with field truthing surveys, along the southern part of the eastern Sicily coast.

Pantano Morghella site is a large (1.3 by 0.8 km in length and width) almost flat (elevation difference of less than 1.5 m over the whole area) wetland, usually partially submerged (less than 1 m water depth) apart from a short period at the end of the summer when it dries out completely (Fig. 94). It is a paleo-fluvial valley nowadays almost totally filled up by fine sediment, communicating with the sea by means of a narrow (less than 1 m large) channel built at the beginning of the 20th century for the installation of a salt pan. This site is surrounded by lavas and volcanoclastic deposits to the south, while limestones, calcarenites and marls outcrop along the northern and western side; interestingly, nowadays a 3 m high fossil dune system together with beach sands (Fig. 94) separate the site from the sea side.

At Pantano Morghella site, by using both hand auger equipment and vibracoring device, we collected 40 cores on which stratigraphic and sedimentologic description together with high resolution pictures were performed both directly in the field and in the laboratory. Moreover, in order to highlight even small scale sedimentary features X-ray analysis was done on few selected cores. Taphonomic and morphoscopic characteristics of macrofaunal remains were quantitatively analysed, while foraminifera were examined both qualitatively and quantitatively, when possible, on some selected cores for palaeoenvironmental reconstruction. A chronological reconstruction of the investigated stratigraphy took advantage from both radiocarbon (8 samples, all shells belonging to the Cerastoderma glaucum species, see Table 2 in Gerardi et alii, 2012) and optical stimulated luminescence (4 samples, see Table 3 in Gerardi et alii, 2012) dating methods.

Results

Sedimentological analysis of the cores revealed that the grain size of the deposits found in Pantano Morghella site is generally fairly uniform. In fact, clay and silty-clay dominate the sediments with only two abrupt changes in grain size (yellowish sandy levels). The higher of the two coarser units in the investigated stratigraphy (Fig. 95) was found in the easternmost part of Pantano Morghella site; here a 3 cm thick yellow sand with silty clay matrix showed up close to the surface overlaying more than 3 m of deposits dominated by clay and silty clay units; it is covered by a thin (max 10 cm thick) brownish soil, even if in some cores, it is immediately followed by a up to 7 cm thick transitional layer made by clay and sand and by the soil only close to the surface. This peculiar layer (called MOR-T01) is also characterized by a sharp basal contact, a variable thickness (3 to 30 cm) and by an inland extension of about 380 m.
The second coarse layer, lower in the stratigraphy, was found in the central and westernmost part of Pantano Morghella site; here a medium coarse yellowish sand rich in shell debris appear at about 1.3-1.5 m below the surface, interrupting an almost uniform section, 5.5 m long, of finer sediments (mainly clay and silty clay, Fig. 96). This peculiar layer (called MOR-T02), characterized by some clayey rip-up clasts, extends inland for at least 1200 m and its thickness decreases with the distance from the sea, from a maximum of 25 cm to a minimum of few cm, in the westernmost part of the pond. Notably, MOR-T02 was also investigated by means of X-ray imaging, showing fining upward grain size and a very sharp basal contact, potentially erosional.

The paleontological analysis performed on the sediments collected in the whole site pointed out to a lagoonal paleoenvironment (e.g. oligothic foraminiferal assemblage dominated by Ammonia tepida and Haynesina germanica together with charophyte gyrogonites and brackish molluscs) that characterize almost all the finer samples, pertaining to the clay and silty clay units (Fig. 95-96). Differently in the two sandy levels an important mixing of lagoonal and marine foraminifera taxa (near-shore to outer shelf benthic species with a slightly increment of planktonic specimens) together with a peculiar richness of shell debris made of fragments of sea urchins, corals, sponge spicules, molluscs and bryozoan are well documented. Moreover, a qualitative comparison between samples of sand collected on the foreshore and the two peculiar sandy levels collected by coring in Pantano Morghella site proved that the paleontological content and grains aspect (many smoothed bioclastic clasts) is very similar among them (Fig. 95).

**Chronological constrains**

By combining the above data, we may say that the two sandy layers (MOR-T01 and MOR-T02) share some peculiar characteristics: coarser grain size with respect to the dominant much finer sediments at the site, fining upward succession, presence of a thin transitional layer on top, sharp basal contact potentially erosional, macro-micropaleontological marine content from the foreshore up to the open sea (differently from all the other sediments investigated at the site but the sand beach). All these helped us in understanding that the two sandy layers are probably the result of a sudden and marked change in the depositional mechanism at the site suggesting the occurrence of exceptionally high-energy depositional events due to marine inundations (storm or tsunami).
Taking into consideration this latter interpretation, we analysed the available anemometric and ondametric data recorded by the closest meteo-marine station in Catania and we calculated the storm-influenced area at the Pantano Morghella site to be about 60 m from the shoreline. If we consider the MOR-T01 and MOR-T02 distribution within the site and their maximum distance from the shore (380 and 1200 m, respectively), these deposits are well beyond the maximum storm flooding distance calculated at the site thus making the tsunami origin as the more plausible mechanism.

Finally, by integrating the available chronological constraints derived from 8 radiocarbon ages and 4 OSL measurements (actually in very good agreement) we may suggest that the time of deposition for MOR-T01 and MOR-T02 should be bracketed between 1860 and 1940 CE and 240 and 500 CE, respectively.

By comparing these ages with the available historical tsunami catalog, we are quite confident that the two sandy layers investigated (MOR-T01 and MOR-T02) may represent the records of the 1908 Messina Straits and 365 Crete tsunamis.
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IGCP Project 639 - “Sea-Level Change from Minutes to Millennia”


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