The formation of Cenozoic mountain belts in the Mediterranean realm was preceded by
tens of millions of years of subduction, forming volcanic arcs, and frontal contractional
systems. In addition, subduction usually involves slab rollback and formation of oceanic
backarcs. Although such structure must have influenced the orogeny of Mediterranean
mountain belts, no active analog has been mapped with modern crustal-scale seismic
methods. Here, we study the entire Calabrian subduction system to map the structure
resulting from Tethys lithosphere subduction and slab roll back, in a process that must be
akin to that operating during a phase of the formation of the Mediterranean orogenic belts.
We present a crustal-scale cross section of the entire Calabrian subduction system.
obtained from on- and off-shore wide-angle seismic data. The 2D P-wave velocity section
shows spatially abrupt (< 5 km of profile distance) structural and petrological transitions
from the Ionian sedimentary wedge and Calabrian arc, to the rifted NW Calabrian margin,
where the Quaternary Aeolian arc is emplaced. The margin, then, transitions northwards
into the Marsili backarc region, where exhumed mantle and localized volcanism occurred
during its formation. This complex structure implies rapid temporal and spatial changes
between magmatic and amagmatic processes, and between compressional and extensional
regimes during the evolution of this subduction system. We find that some terranes
involved in the Alpine orogeny share petrological and tectonic similarities with some
domains of the Calabrian subduction system. Based on the results of this study we propose
the Calabrian Arc system as an analog for the subduction structuration that preceded the
formation of Alpine orogenic systems.

Keywords: Calabrian Arc, subduction; Mantle exhumation; Travel-time tomography;
Wide-angle seismic data; Alpine orogeny; Mediterranean active arcs
1 Introduction

The Mediterranean realm includes a system of Cenozoic arcuate orogenic belts formed during the Alpine Orogeny (Dewey et al., 1973) (Fig. 1a). Mantle tomography images and plate reconstructions support that the collision events that formed these orogens were preceded by subduction of Tethys lithosphere (e.g. Dewey et al., 1973; Schettino and Turco, 2011). The evolution of the subduction systems must have often included slab retreat (e.g. Handy et al., 2010), causing the migration of volcanic arcs, extension of the overriding plate and formation of backarc basins possibly similar to the Tyrrhenian (Prada et al., 2015, Loreto et al., 2020) and the Alboran-South Balearic basins (Gomez de la Peña et al., 2018), and forearc contractional systems (e.g. Stampfli et al., 1998; Polonia et al., 2011; Marroni et al., 2017), determining thereby, the pre-collisional structuration of the lithosphere. The Gibraltar and Calabrian backarc are no longer opening, and their domains are currently being inverted, which may indicate a current initial phase of a collision (Giaconia et al., 2015; Zitellini et al., 2020). Yet, the structure of Alpine-type belts has been associated to the imbrication of the domains formed during continental rifting (e.g. Reston and Manatchal 2011; Mohn et al., 2014), and the major lithospheric-scale structuration created during subduction has been largely ignored, possibly because of the fragmentary geological record in mountain belt that makes reconstructions disputable, and the lack of present-day potential analogs.

Reconstructing the structure of past subduction systems from present-day orogens is a challenging task because of the tectonic and metamorphic overprinting that terranes suffer during orogeny, and the incomplete outcrop information. An alternative way to infer such structure is by exploring the present-day subduction of the Tethys lithosphere in the Mediterranean, as occurs under the Calabrian, Gibraltar and Hellenic arcs (Fig. 1a)
Possibly similar to Alpine subductions (e.g. Maffione and van Hinsbergen, 2018), the upper plate of these present-day subduction systems has been shaped by slab retreat, backarc opening, and accretionary forearc systems. Therefore, understanding of the forearc-to-backarc structure of present-day subduction systems in the Mediterranean may provide insights into the processes that shaped Alpine systems before collision.

In this work, we focus on the Calabrian subduction system, which includes the Tyrrhenian backarc basin, the Aeolian volcanic arc, the NW Calabrian margin, the Calabrian arc and the contractional Ionian wedge (Fig. 1a and 1b). Vintage (i.e. 1970) refraction studies provided 1D crustal velocity information across the subduction system, bringing first-order approximations on the crustal structure of the Calabrian arc (e.g. Cassinis et al., 2003). Regional earthquake tomography of the Calabrian region has provided insights on the structure of the slab and the petrological nature of the deep lithosphere and asthenosphere (e.g. Caló et al., 2009). However, these studies do not provide enough resolution to assess in detail the structure and petrological nature of shallow lithospheric domains (i.e. crust and uppermost mantle) across the subduction system.

To assess the forearc-to-backarc structure of this system, we present a detailed P-wave velocity ($V_p$) cross section of the entire subduction system using travel-time tomography from wide-angle seismic data (WAS) (Fig. 1) acquired in 2015 during the CHIANTI (Calabrian arc Hazards in IoniAN and TyrrhenIan seas) experiment (Fig. 1b). The 425 km-long model provides unprecedented information of the structure and petrology of shallow lithospheric domains across the Calabrian subduction system from the forearc to the backarc (Fig 1b).
2 The Calabrian Arc subduction system

Calabria has earthquakes activity up to ~500 km depth under the northern margin of Calabria, delineating a ~70° NW dipping slab (Spakman and Wortel, 2004; Chiarabba et al., 2005). Slab rollback has been driving backarc basin formation by upper plate extension, similar to other regions of the Western and Central Mediterranean since the Oligocene (~30 Ma; Malinverno and Ryan, 1986; Schettino and Turco, 2011). After the opening of the Liguro-Provençal basins, the east-southeast rollback of the Apennines-Calabrian subduction system initiated the Tyrrhenian backarc basin in the early Tortonian (~11 Ma) (Kastens and Mascle, 1990). During the Messinian (5.3-6 Ma) the potential under-thrusting of the continental lithosphere of Adria stopped slab rollback and backarc extension in the northern Tyrrhenian (Faccenna et al., 2001).

In contrast, the central and southern subduction front migrated towards the southeast (Spakman and Wortel, 2004), opening the central and southern Tyrrhenian Basin, and forming the Magnaghi and Vavilov basins during the Pliocene, and the Marsili basin during the Quaternary (Beccaluva et al., 1990; Kastens and Mascle, 1990) (Fig. 1b). Ocean Drilling Program (ODP) Leg 107 and Deep Sea Drilling Program (DSDP) Leg 42 recovered mid-ocean ridge basalts in the Magnaghi and Vavilov basins (Dietrich et al., 1977; Beccaluva et al., 1990). Further basalts were also sampled at the top of the basement in the Marsili basin (Kastens and Mascle, 1990) (Fig. 1b). Consequently, these basins have commonly been interpreted as regions where break up led to oceanic spreading. However, recent $V_p$ models across the Vavilov and Magnaghi basins support that their basement is mainly composed of exhumed mantle rocks, and that basaltic ridges in the area are localized magmatic events (Prada et al., 2015, 2016). The occurrence of exhumed mantle in these two basins is further supported by the drilling of serpentinized peridotites.
The southeastward migration of the subduction system resulted in rifting of the NW Calabrian margin and emplacement of the Quaternary Aeolian volcanic arc at < 1.2 Ma (Argnani and Savelli, 1999), the tectonic Calabrian Peninsula, formed by Paleozoic to Eocene metamorphic rocks covered by Cenozoic sedimentary units (Vitale et al., 2019), and the Ionian contractional wedge (Polonia et al., 2011) (Fig. 1a).

3 Acquisition, processing and picking of wide-angle seismic data

The CHIANTI seismic experiment was conducted onboard the Spanish R/V Sarmiento de Gamboa. The offshore data used in this study were acquired along two offshore profiles, WAS1 in the Ionian Sea, and WAS2 in the Tyrrhenian backarc region (Fig. 1b). The offshore data were recorded by 14 and 17 LC2000 Ocean Bottom Seismometers (OBS) of the Spanish National Research Council (CSIC) along profiles WAS1 and WAS2, respectively. OBS spacing along both lines was ~ 10 km. The seismic source was generated by two arrays of 9 airguns each, which released a total volume of 4760 cu. in. (78 l), and worked at 15 m of depth to provide a range of frequencies between 7-45 Hz, optimum for crustal-scale WAS experiments. Shot interval along both lines was 90s (~230 m at 5 knots). Land stations in Calabria were installed by the University of Calabria. In this study, we use 31 OBS records with useful data and 3 out of the 5 land stations installed in Calabria, which recorded the shots along both WAS1 and WAS2 (Fig. 1).

We relocated all OBSs using an in-house Metropolis-Hastings algorithm that minimizes the misfit function between the observed and synthetic first arrival travel-times of the water wave to find the most likely location of the OBS. We applied a bandpass filter (1-5-18-25 Hz) to all seismic records to enhance the seismic signal and manually picked
travel-times of first arrival and critical reflections. In the contractional wedge, near offset
(< 15 km) first arrival times from OBS record sections of profile WAS1 define a first
refracted seismic phase with an apparent velocity of 3-4 km/s that corresponds to refracted
P-waves through the shallow sedimentary sequence of the contractional wedge or Ps (Fig.
2a-b). Between 15-20 km offset a second refracted seismic phases with an apparent
velocity of 5-6 km/s can be identified in those OBS deployed near the shore (Fig. 2b). We
interpreted this latter phase as refracted P-waves through the uppermost basement of the
contractional wedge or Pb. Some sections show a prominent critical reflection at 15-20
km of offset, defining the transition between refracted phases Ps and Pb (Fig. 2a-b). We
interpreted this phase as P-waves reflected at the top of the acoustic basement or PnP (Fig.
2a-b). The land record sections of line WAS1 only allowed the identification of seismic
phases Ps and PnP, up to ~40 km offset (Fig. 2h).

From SE to NW in the NW Calabrian margin, OBS records along line WAS2 display a
first refracted seismic phase with apparent velocity of 6 km/s. This seismic phase is
interpreted as a P-wave refracted within the crust (Pg). At further offset (> 40 km offset),
we identified a second refracted phase, in this case with a faster apparent velocity of ~8
km/s. We interpret this phase as a P-wave refracted within the uppermost mantle (Pn).
The transition from Pg to Pn is marked by the arrival of critical reflections at the crust-
mantle boundary (PnP). An additional reflected phase can be identified in some receivers
located in the NW Calabrian margin. This seismic phase arrives at shorter offsets than
PnP’s, at 30-20 km on average (PgP in Fig. 2c), indicating the presence of a major intra-
crustal interface below the margin. Towards the NW, in the Marsili basin, OBS receivers
do not display clear Pg and PnP phases, just a single prominent refracted phase with
apparent velocity of ~8 km/s. These seismic records resemble to ocean bottom
hydrophone data recorded in the exhumed mantle region of the Magnaghi and Vavilov
basins during the MEDOC WAS experiment (Prada et al., 2016). For layer-stripping
modelling purposes, we also assigned this phase the P\textsubscript{s} label. Land station records of line
WAS2 show clear P\textsubscript{g} and P\textsubscript{m} phases (Fig. 2g).

In total, we picked 2837 P\textsubscript{s}, 857 P\textsubscript{s}P and 170 P\textsubscript{b} from WAS1 record sections, and 4378
P\textsubscript{g}, 675 P\textsubscript{g}P, 1090 P\textsubscript{m}P, and 3081 P\textsubscript{n} from WAS2 records. Picking uncertainty was set
between 30 ms and 100 ms based on the amplitude ratio of a 250 ms-long window before
and after the picked travel-time, following the approach of Zelt and Forsyth (1994).

4 Methods: travel-time tomography

We invert for V\textsubscript{p} and the geometry of the different seismic interfaces (i.e. Moho) using
the joint refraction and reflection travel-time tomography code TOMO2D (Korenaga et
al., 2000). The starting velocity model is parameterized as a regular grid hanging from
the seafloor, with 250 m node spacing both vertically and horizontally. Regularization
parameters are defined by horizontal and vertical correlation lengths (CL) that increase
from top to bottom in the velocity grid. In this study, we set 0.2 km and 1 km for the
vertical and horizontal CL at the top of the model, respectively, and 10 km and 20 km at
the bottom of the model, located at 50 km depth.

We have followed a layer-stripping strategy as in Prada et al. (2015) to resolve the sharp
velocity contrast attributed to the top of the acoustic basement in the contractional wedge,
and the crust-mantle boundary of the NW Calabrian margin. This way, along WAS1 we
built the model following a two-step inversion, each step consisting of 10 iterations. In
the first step, we inverted P\textsubscript{s} and P\textsubscript{s}P travel-times. The corresponding solution is included
and damped in the starting model for the second step, in which we inverted P\textsubscript{b} travel-
times together with P\textsubscript{s}, and P\textsubscript{s}P travel-times. We followed a similar approach to build the
model along profile WAS2. First, P$_g$ and P$_{nP}$ travel-times were inverted to solve the V$_p$
structure of the crust beneath the NW Calabrian margin, as well as the Moho geometry.
In a second step, we included and damped the solution of the first inversion within the
starting model for the second inversion step, in which we added P$_n$ travel-times to the
previous dataset. Once both models (i.e. WAS 1 and 2) were obtained, travel-times of
refracted phases of land stations in both profiles were jointly inverted to better constrain
the velocity structure of the Calabrian arc using crossing rays. P$_g$P travel-times along
WAS2 were inverted using the final velocity model as a reference to retrieve the geometry
and location of the intra-crustal reflector beneath the margin (Fig. 3). Acceptable ray
coverage beneath profile WAS1 in the Ionian is restricted to the first 10 km of depth of
the model, while tomographic constraints along WAS2 can reach down to 15-20 km of
depth (Fig.3b). The final root mean square (RMS) of refracted and reflected travel-times
for the final velocity model (Fig. 3) is ~60 ms and ~80 ms, respectively.

4.1 Uncertainty and resolution of the tomographic model

The uncertainty of the model parameters was assessed by means of a Monte Carlo
analysis (e.g. Korenaga et al., 2000). We followed the same layer-stripping strategy used
to obtain the preferred model to estimate the parameters uncertainty of each layer of the
model. We tested 100 realizations, each of them using a 1D V$_p$-depth profile and a flat
reflector as initial model, and a travel-time dataset perturbed with random Gaussian noise
(up to ±100ms). To generate the starting model, we have applied a random perturbation
of ±10% and ±6% for crustal and mantle velocities, respectively, of the reference velocity
model for each layer (Supplementary Table 1). A perturbation of ±4 km respect the
reference depth for each reflector was applied to generate each starting interface
(Supplementary Table 1). The set of correlation lengths were also randomly perturbed in
each realization. This way, we have used for each CL: 3±2 for the top horizontal CL, 15±5 for the bottom horizontal CL, 0.3±0.2 for the top vertical CL, and 4±1 for the bottom vertical CL.

All realizations converge after 20 iterations with an overall RMS of travel-times of 80 ms (Fig. 4a). We have computed the standard deviation of $V_p$ and depth of each reflector to quantify the range of variation (uncertainty) attributed to all random errors. These values show that the model presents standard deviations < ±0.1 km/s in most of the model, indicating a satisfactory control of seismic velocities. As expected, higher values (±0.2 - ±0.3 km/s) are observed in regions with a sharp vertical velocity contrast or limited ray coverage, like at the lower part of the wedge (Fig. 4). The uncertainty of each reflector ranges between ± 0.15 km in areas well covered by reflected rays, and ±4 km towards the edges of each reflector, where ray coverage is limited (Fig. 3b).

We have also evaluated the resolution of the tomographic model by performing a checkerboard test, following the approach in Zelt (1998). This way, a total of 72 checkerboard-type models with nine different squared cell sizes (4, 6, 8, 10, 12, 14, 16, 18, and 20 km), and with a maximum velocity perturbation of 5% have been tested. For each cell size, eight different patterns were created with positive and negative polarity, with and without 45º rotation, and with and without a diagonal shift of half of the cell size. Each checkerboard pattern was added to the preferred tomographic model to create each target model and compute the synthetic travel-times. Before inverting each synthetic dataset, we added a random Gaussian noise, with a standard deviation corresponding to the picking uncertainty, to each travel-time, as in the Monte Carlo analysis.

After inverting each dataset using the preferred model as starting model, we compute the semblance between the retrieved pattern and the corresponding target pattern. As in Zelt
(1998), we consider the pattern to be well retrieved for a semblance > 0.7 (red contour in Fig. 5a and 5b). We then compute the average semblance for each cell size and combine them all in a resolution map that depicts the maximum resolution depth of each cell size (Fig. 5c). The map shows that the minimum cell size that can be retrieved is 4 km along profile WAS2, while it is 6 km along WAS1. The resolution depth of the largest cell size (20 km) ranges between 15-20 km of depth beneath the arc/backarc, and 10 km of depth beneath the contractional wedge. The differences in resolution between the arc/backarc and the wedge is caused by the lower amount of travel-times (rays) observed in the seismic records of the Ionian region (Fig. 3b).

5 Results

From SE to NW, the final tomographic model shows the $V_p$ structure of the main crustal domains of the Calabrian subduction system, namely the Ionian contractional wedge, the Calabrian Arc, the NW Calabrian rifted margin, the volcanic arc (at ~290 km in Fig. 3), and the Marsili backarc basin in the Tyrrhenian (Fig. 3).

In the Ionian, the $V_p$ model delineates the Pre-Messinian wedge towards Calabria, across the Inner Plateau and the Squillace basin (Fig. 1b and 3). The shallow velocity structure across the Pre-Messinian and Inner Plateau sectors, which have distinct seafloor morphology, indicates a continuous crustal domain. The $V_p$ of the shallowest ~3 km displays a gradual increase towards the coast (km 0-175 in figure 3a). $V_p$ of <$~3$ km/s correspond to low consolidated sediment, particularly thick under OBSs 7-6, 3, and 1 (Fig. 3). The shallow $V_p$ structure in the wedge has good ray coverage (Fig 3b) and uncertainties <0.1 km/s (Fig. 4), with well-resolved lateral changes of ~10 km in width (Fig. 5). Underneath, $V_p$ steadily increases from 4.0 km/s to 5.0-6.0 km/s with depth (Fig 3a). $V_p$ >~4 km/s imply a well consolidated material, and $V_p$ 5.0-6.5 km/s at 10-15 km...
depth indicate crystalline rocks with little porosity (Fig. 3), although with depth ray coverage and resolution decrease and uncertainty increases (fig. 4). Near the coast, the ~4-5 km deep Squillace basin (beneath OBS 17-16 in Fig. 3) $V_p$ increases from 1.70 to 4.0-4.5 km/s at 6-7 km depth, underlaid by $V_p > 5$ km/s that supports shallow continental basement rocks. A relatively continuous intra-wedge regional wide-angle reflector roughly follows $V_p$ 3-3.5 km/s, shallowing from 10 km depth beneath OBS 17 to ~5 km depth beneath OBS 7-4. The $V_p$ does not abruptly changes across the reflector, which may indicate a rapid decrease in porosity and/or fracturing with depth rather than an abrupt boundary.

The structure beneath mainland Calabria is partially resolved with rays from both Ionian and Tyrrhenian marine shots. $V_p$ increases from ~4.0-4.5 km/s at ~1 km depth to 6.5-7 km/s at 20-22 km depth, above the Moho mapped by $P_{n}P$ reflections in a sector of NW Calabria (at 220 km in Fig. 3). These values are in agreement with Moho depth from the 1979 Calabria transect (Cassinis et al., 2003), and receiver functions (Piana-Agostinetti et al 2009). The Calabria crust $V_p$ range and low vertical velocity gradient (~0.1 s$^{-1}$) match typical continental crystalline crust $V_p$ distribution (Christensen and Mooney, 1995) as shown in Fig. 6.

The NW Calabria rifted margin continental crust thins from ~20 km under the coastline to 10 km at the NW edge (km ~325 in Fig. 3). Along this margin, $V_p$ increases from 1.8 km/s to 3.5 km/s at ~1.5 km below seafloor and then abruptly to 5.0 km/s at 2 km below seafloor roughly marking the basement top (Fig. 3). The basement $V_p$ is laterally homogeneous and with a gentle vertical gradient from ~ 5 to ~7.0 km/s near the Moho. An intra-basement reflector delineates a slight decrease in vertical gradient for $V_p > 5.7-6.0$. The basement velocity under Lametini volcano of the Aeolian arc (Fig. 1b) is well-
constrained and comparatively higher, of 6 km/s at ~1 km beneath top basement to ~7.0-
7.3 km/s in the lowermost crust (km ~290 in Fig. 3). The margin uppermost mantle \( V_p \)
increases from 7.7-7.8 km/s to 8.2 km/s in 5 km under the Moho (Fig. 3).

The backarc region displays an abrupt lateral change in \( V_p \) structure at km ~330 (Fig 3a).
The Marsili basin has no Moho boundary (no \( P_mP \) reflections, Fig. 2e-f) and the transition
to mantle \( V_p \) occurs with comparatively steep vertical velocity gradients of ~1 s\(^{-1}\) for \( V_p \)
of ~4.5-7.0 km/s reaching \( V_p \sim 7.5-8.0 \) km/s at ~5 km below seafloor. The general Marsili
basin \( V_p \) structure is locally interrupted by a gentler vertical velocity gradients of the two-
layer \( V_p \) structure under the 3-km-high Marsili volcano. The upper layer increases rapidly
from 3 km/s near the surface to 6 km/s at 6-7 km of depth, with a 0.6 s\(^{-1}\) gradient (Fig. 3).
The second layer \( V_p \) increases from 6 km/s to ~8 km/s with a 0.2 s\(^{-1}\) gradient (Fig. 3). The
\( V_p \) model beneath small seafloor ridges under OBS 28 and 29 contains local sub-vertical
anomalies with relatively lower \( V_p \) (Fig 1 and 3).

6 Discussion

6.1 Geological interpretation

6.1.1 Ionian wedge

The upper 4-5 km of the internal wedge is characterized by low velocity regions (i.e. <
2.5 km/s). These low velocity anomalies may indicate comparatively higher porosity in a
region of fluid overpressure. Some of these low velocity anomalies (i.e. OBS 6-7 in Fig.
1 and 3) are spatially coincident with mud volcanoes mapped with multibeam bathymetry
(Gutscher et al. 2017), that are associated with fluid and mud expulsion (Loher et al.,
2018). Previous tectonic interpretation based on MCS data shows that low velocity
anomalies beneath OBS 6 to 1 are spatially coincident with compressional structures
Thus, we interpret that these low velocities may indicate regions of the wedge where active shortening is causing fluid outflow through thrust faulting.

6.1.2 The NW Calabrian margin

The basement of the NW Calabrian margin and that beneath Calabrian exhibits a similar vertical $V_p$ profile than the global continental crust compilation of Christensen and Mooney (1995) (Fig. 3 and 6). The Moho beneath the NW Calabrian margin shallows from 20-22 km of depth beneath Mainland to ~10 km of depth, reflecting the result of the Late Miocene backarc extension of the margin (~11-7 Ma ago; Kastens and Mascle, 1990).

The intra-crustal reflector modeled at 7-8 km depth roughly delineates the $V_p$ ~5.8-6.0 km/s (Fig. 3). The boundary does not mark an abrupt change either in $V_p$ or in gradient with depth, and we speculate that it may represent a transition with depth in rock fracturing and associated fluid percolation, possibly related to a change from shallower fault-controlled brittle deformation to deeper plastic lower crust deformation. A similar result is found in the Galicia Interior basin (Perez-Gussinyé et al., 2003), where the resulting reflector delineate the 6.2-6.3 km/s $V_p$ at 9-10 km depth in a ~15-km-thick basement. There, coincident seismic images and thermal modelling support that the reflector marks the boundary between a faulted upper crust and a ductile lower crust at the time of rifting.

6.1.3 The volcanic arc

The NW Calabrian margin is characterized by $V_p > 7$ km/s beneath the volcanic arc (Fig .3). These velocities are anomalously high for extended continental crust settings, which typically range between 6.69-6.93 km/s (Christensen and Mooney, 1995). To assess the
petrological nature of these rocks, we compare the vertical velocity structure of the Aeolian arc with (1) the vertical $V_p$ structure of the NW Calabrian margin at 275 km of profile distance, (2) the empirical $V_p$-depth distribution of different rock-types of the Izu-Bonin-Mariana arc in the Pacific (Kitamura et al., 2003), and (3) the $V_p$-depth reference function for continental crust (Christensen and Mooney, 1995) (Fig. 3a). The $V_p$-depth distribution of the arc reveals a compositional differentiation from felsic/intermediate (i.e. 6.0 to 6.8 km/s; Christensen and Mooney, 1995) in the upper crust to mafic material in the lower crust (Fig. 3a), where the $V_p$-depth distribution coincides with the $V_p$-depth distribution for gabbroic rocks of Izu-Bonin-Mariana arc (Fig. 6a; Kitamura et al., 2003).

6.1.4 Marsili backarc basin

In the backarc region we compared average vertical velocity distribution at either side of the Marsili volcano, with reference $V_p$-depth functions from modern tomographic studies for oceanic crust (Grevemeyer et al., 2018a), exhumed mantle regions from the central Tyrrhenian and Gulf of Cadiz (Prada et al., 2015), as well as exhumed mantle regions from ultra-slow spreading centers (Grevemeyer et al., 2018b). Oceanic crust and exhumed mantle references have different $V_p$-depth trends reflecting their petrology (Fig. 6b). The oceanic crust reference has two-layer structure associated to the petrological differentiation between the widespread basaltic layer 2 and the gabbroic layer 3 (e.g. Grevemeyer et al., 2018a), while the exhumed mantle field appears as a continuous, comparatively steeper velocity gradient of 0.6-0.7 s$^{-1}$ of the upper ~5-6 km of basement, reflecting the decreasing degree of serpentinization with depth (e.g. Prada et al., 2016). The comparison with our observations supports that the basement of the Marsili backarc basin is made of exhumed mantle similar to the Magnaghi and Vavilov basins in the Tyrrhenian.
Previous regional studies interpreted that the Marsili basin is floored by oceanic crust produced by ultrafast oceanic spreading (e.g. Nicolosi et al., 2006; Manu-Marfo et al., 2019). However, our results show that the structure that can be related to oceanic spreading, that is the Marsili volcano and neighboring ridges, are isolated velocity anomalies in the tomographic model, and that the bulk of the basin has a velocity gradient matching the exhumed mantle $V_p$-depth reference. This result, together with similar observations from the Magnaghi and Vavilov basins (Prada et al., 2016), indicate that mantle exhumation and emplacement of localized oceanic ridges and large-volcanic edifices (Fig. 1b) have been recurrent events during the opening of the Tyrrhenian backarc.

The occurrence of exhumed mantle in the Tyrrhenian basins results paradoxical. Stratigraphic record analysis of the Marsili Basin allowed to infer that the basin opened ~2 Ma ago (Kastens and Mascle et al., 1990; Argnani and Savelli, 1999), so that mantle exhumation should have occurred at a rate of 3.5-4.5 cm/yr (Kastens and Mascle et al., 1990), although magnetic anomaly analysis suggests a much faster rate (~19 cm/yr; Nicolosi et al., 2006). This rate is exceptionally fast considering models of partial decompression melting, which suggest that mantle unroofing should not occur at rates >~2 cm/yr (e.g. Greveemeyer et al., 2018b). Geochemical analyses of serpentinized peridotites from site 651 show that these rocks are strongly depleted in lithophile elements during recurrent partial melting events (Bonatti et al., 1990). This suggests the presence of a depleted mantle source, which could restrict the production of partial decompression melting during lithospheric extension (Perez-Gussinyé et al., 2006).
6.2 The forearc-to-backarc structure of the Calabrian arc: resemblance with Alpine subduction systems

The 450 km long transect displays the contrasting Vp structures of the Ionian contractional wedge, the 20-22 km thick continental crust of the Calabrian Arc, the NW Calabrian rifted margin, locally modified by ~20 km-wide volcanic arc activity, and the exhumed mantle domain of the Marsili backarc basin, locally intruded by volcanic ridges (Fig. 7). These domains represent terranes made of distinct petrological suites and formed by contrasting deformation regimes, bounded by <4-5 km wide transitions, near the resolution limit of our data (Fig. 5). This complex terrane structuration involves a tectonic and magmatic evolution not fully understood. The NW Calabrian margin integrates the evolution of extended continental crust possibly by a Miocene to younger rift (Kastens and Mascle et al., 1990), modified by the emplacement of the volcanic arc, and the development of the Plio-Quaternary Paola basin, currently in a forearc position, with up to 5 km infill (Zitellini et al., 2020) that extends towards the NW of our transect (Fig. 1 and 3). The backarc Marsili basin is possibly Quaternary (Kastens and Mascle et al., 1990), and formed by amagmatic extension leading to mantle exhumation and the subsequent intrusion of a ~20 km-wide volcano and localized ridges. This basin is surrounded to the north, east and south by continental crust and volcanic complexes related to the subduction (Fig 1b), while towards the northwest, the basin leads to the Vavilov Basin, where exhumed mantle and localized oceanic ridges were emplaced during its formation (Prada et al., 2016).

This complex structuration resulted from subduction and rollback of the Jurassic Tethys lithosphere, which is the same geodynamic process that controlled the subduction phase of possibly many Alpine systems (e.g. Schmid et al. 1996; Wortel and Spakman, 2000; Maffione and van Hinsbergen, 2018). The complex structure of the Calabrian subduction
system is possibly not unique or anomalous, and similarly complex structures might have occurred at overriding plates of ancient Alpine subduction systems.

General proposals indicate that the structural inheritance of rifted margin architecture is key to explain orogenic configuration (e.g. Reston and Manatchal 2011). More specifically Alpine rock units originally from the Jurassic Tethys (Froitzheim and Eberli 1990) are interpreted to signify a structural inheritance that significantly influenced the orogenic development of Alpine belts (e.g. Mohn et al., 2014). Based on those concepts, it is proposed that closure of suspected narrow oceans lacks significant subduction-related magmatism, leaving little trace in mountain belts like the Alps and Pyrenees. It is argued that those orogens develop fundamentally by processes in which the original rifted margins structure controls the final orogen architecture (e.g. Chenin et al., 2017). Although the Pyrenees are overwhelmingly interpreted as the closure of a narrow rifted basin, the Alps are debated and the subduction phase may have lasted 25 Ma, from Early Paleocene to at least Late Eocene, with >500 km of subduction (Schmid et al., 1996), similar to the present-day subduction systems in the Central Mediterranean (e.g. Faccenna et al., 2001). Therefore, such subduction system could have created a structuration similar to the Calabrian Arc system. Furthermore, at a larger scale, opening during Tethys-Ionian slab subduction and rollback since ~25 Ma triggered the formation of the passive margin of the Gulf of Lions, the Ligurian basin and the Corsica-Sardinia continental block (Schettino and Turco, 2011), which adds further complexity to a pre-collision scenario.

The geological record of several Alpine-type mountain belts retain unequivocal evidence of terranes formed during subduction, sharing similarities with domains of the Calabrian arc. Conspicuous units found in Alpine belts such as the Apennines, Alps, Hellenides, Dinarides, and Taurides are the remnants of forearc accretionary systems (e.g. Stampfli
et al., 1998; Okay et al., 2006; Marroni et al., 2017; Maffione and van Hinsbergen, 2018), and volcanic arcs (Sharkov et al., 2014). But the most distinctive units implying subduction are serpentinized mantle rocks accompanied by basaltic rocks found in Alpine-type ophiolites (e.g. Moghadam and Stern, 2015). Although some mantle rocks from the Alps are related to mantle exhumation in Tethys rift context (e.g. Froitzheim and Eberli 1990), the majority of these ophiolites have geochemical signatures that support a supra-subduction origin and are spatially associated to magmatic rocks with Island arc affinity (e.g. Dilek 2003; Moghadam and Stern, 2015; Maffione and van Hinsbergen, 2018), similar to Tyrrhenian samples (Beccaluva et al., 1990). In particular, assemblages similar to Tyrrhenian serpentinized mantle rocks are described in the east and west Vardar Ophiolites in Albania (Maffione and van Hinsbergen, 2018), the Tsiknias ophiolites in Greece (Lamont et al., 2020), and Iran (Moghadam and Stern, 2015). This latter suit of serpentinized mantle rocks found in Mesozoic ophiolites, formed prior to the construction of the Zagros belt, share a peridotite spinel composition with Tyrrhenian peridotites drilled at ODP leg 107 (Fig. 1). Both rocks present high values of Cr# (> 40) and low values of Mg# (< 60), indicating a similar degree of depletion during melt extraction at convergent margins (see fig. 7 in Bonatti et al., 1990; fig. 13 in Moghadam and Stern, 2015).

Based on these observations, it is likely that the pre-collisional structure of many Alpine orogens from South Europe to the Middle East that experienced extended subduction periods, possibly accompanied by slab retreat, was formed by neighboring terranes with distinct petrological and tectonic characteristics separated by spatially abrupt boundaries, resembling to the structuration mapped here across the of the Calabrian Arc (Fig. 7).
7 Conclusions

The first modern WAS transect across the Calabrian Arc system provides a 2D $V_p$ tomographic image of the entire system along 450 km from forearc to the backarc region from joint inversion of refracted and reflected P-wave travel-times.

The Calabrian system structure is characterized by spatially abrupt changes (<4-5 km width) between domains formed by different processes and differing petrological and tectonic structures.

The frontal region where contractional tectonics dominate includes the offshore internal wedge formed mainly by consolidated rocks covered by 1-3 km of sediment with $V_p < 3.5$ km/s. The $V_p$ model of the wedge displays gentle undulations defining local low velocity anomalies (< 2.5 km/s) suggesting the presence of high porosity regions of the wedge. These anomalies are spatially coincident with thrusting and mud volcanism on surface, suggesting that thrusting leads to fluid outflow to the seafloor. The transition between the inner wedge and Calabria occurs under the Squillace basin, with an abrupt shoaling of the consolidated basement rocks.

The Calabrian Peninsula has a 20-22 km thick crust with a typical continental crust velocity structure. The continental Calabrian crust was extended and thinned in the Miocene to 16-9 km across ~100 km of the NW Calabrian margin. The rifted margin structure was subsequently modified during Plio-Quaternary time by magmatic arc activity across a ~20 km wide area and the development of the 3-5 km thick Paola forearc basin. The continental crust abruptly thins over ~5 km width to the Marsili basin.

The 200 km wide Marsili basin has a vertical velocity gradient typical of exhumed mantle domains, similar to Vavilov and Magnani basins in the central Tyrrhenian. Local velocity
anomalies under Marsili volcano and low ridges indicate that basaltic features are isolated intrusions in the basement of the basin, which is inferred to be composed by serpentinized peridotite.

Although it has been long speculated that orogen formation is strongly influenced by the inherited structure of rifted continental margins, most conceptual models and reconstructions have not accounted for the subduction phase that possibly preceded collision in most of the Alpine-type orogens from South Europe to the Middle East. We show the first detailed structure of an analog of a system created by Tethys slab subduction and rollback, which has created an unanticipated complex pattern of large-scale lithospheric domains, some of which are similar to terranes found in Alpine-type mountain belts.

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9 References


Figure 1. a) Bathymetry and topography map of the Mediterranean region showing the location of the main Cenozoic orogenic belts, and the location of the seismic cross section presented in this study. AL: Alboran Sea, LP: Liguro-Provençal, TY: Tyrrhenian, IO: Ionian. b) Bathymetry map of the Tyrrhenian and Ionian region showing the petrological nature of basement rocks from ODP sites (Kastens and Mascle, 1990), DSDP sites (Dietrich et al., 1977), and dredging samples (Colantoni et al., 1981). Yellow circles
depict the location of ocean bottom seismometers and land stations used to record shots along profiles WAS 1 and WAS 2. The record sections of numbered OBSs and land stations UC1 and UC3 are shown in Figure 2. White circles show the location of ODP and DSDP boreholes in the Tyrrhenian, while mud volcanoes in the Contractional wedge (CW) are depicted with dark red circles. The boundary between the internal and external wedge (dashed white line) and the location of mud volcanoes is taken from Gutscher et al. (2017), and Loher et al. (2018), while the location of the morphological domains of the Inner plateau and the Pre-Messinian wedge (PMW) are taken from Polonia et al. (2011). Red lines and triangles in the PMW are splay faults from Polonia et al. (2011). Earthquakes occurring within ±50 km of distance from the WAS profile are projected along the geophysical cross section and presented in the small inset to show the steep (~70°) geometry of the Ionian slab beneath the study area (earthquakes are from Chiarabba et al., 2005). AI: Aeolian Islands, CA: Calabria, LV: Lametini Volcano, Ma-Vb: Magnaghi-Vavilov basins, Mb: Marsili basin, MaV: Marsili Volcano, Pb: Paola basin.
Figure 2. Seismic record sections of OBSs (a)6, (b)16, (c)19, (d)24, (e)31, (f)34, and land station (g) UC3 record from profile WAS2, and (h) UC1 record from profile WAS1. The location along the line of each instrument is shown in Fig. 1b. The upper panels show the interpreted seismic phases, while the lower panels show the records with the picked refraction and reflection travel-times (blue bars), and the corresponding synthetic refraction (red dots) and reflection (green dots) travel-times obtained in the final model.
Figure 3. 2D P-wave velocity model of the Calabrian subduction system obtained from joint inversion of refracted and reflected travel-times. The inverted reflectors are intra-wedge reflector (IW), the intra-crustal interface (IC) and the Moho along the NW Calabrian margin. Ma.V: Marsili Volcano, P.B: Paola basin S.B.: Squillace basin. Red dots are OBS and land stations. Vertical red, orange, and brown arrows indicate the location of the vertical velocity-depth profiles shown in Figure 7a. Red arrow also shows the location of the volcanic arc, while the dark red triangle shows the location of mud volcanism neighboring WAS line 1 in Fig. 1. (b) Derivative weight sum of the tomographic model. This image can be used as a proxy for the refracted and reflected ray coverage through the grid during the inversion.
Figure 4. a) Distribution of RMS travel-times of the first (grey dots) and final (red dots) iteration of each Monte Carlo realization (100 in total) showing how the initial random distribution converges to a common solution (Final RMS ~80ms). b) Average final tomographic model of the 100 Monte Carlo final solutions. The red bands depict the error bar of each reflector. c) Standard deviation of the 100 realizations.

Figure 5. Retrieved (upper) and true (lower) checkerboard pattern of the (a) 8x8 km and (b) 12x12 km anomaly size. The red contour line indicates where the pattern is retrieved satisfactorily (i.e. where the semblance is ~0.7). Red dots are OBSs. c) Resolution map of the model showing the region at which we are able to retrieve each size of the checkerboard pattern.
Figure 6. a) Velocity-depth profiles of the region of the volcanic arc (red band), the NW Calabrian margin (yellow band), and mainland Calabria (dark brown) compared with the continental crust velocity-depth function of Christensen and Mooney (1995) and the velocity-depth function for tonalitic and gabbroic rocks of the Izu-Bonin-Mariana arc (Kitamura et al., 2003). The thickness of each profile from our model corresponds to the error bars inferred from the Monte Carlo analysis. Note that no error bar is drawn for the Calabrian profile as the Monte Carlo analysis was focused on the offshore regions. b) Average velocity-depth profiles of the NW (light green) and SE (dark green) Marsili volcanic area compared with the velocity-depth reference function for oceanic crust from Grevemeyer et al. (2018a), and exhumed mantle regions of the Tyrrhenian and the Gulf of Cadiz from Prada et al. (2015), and at ultra-slow spreading centers (USSC) Grevemeyer et al. (2018b).
Figure 7. Interpretative cross section of the entire Calabrian subduction system based on the result of this study and the depth-distribution of earthquakes from 20 to 60 km of depth (white circles) from Fig. 1b. The geometry of the Ionian subducting slab is taken from (Chiarabba et al., 2005), while the Ionian oceanic Moho is based on the results from Cassinis et al. (2003) and Piana-Agostinetti et al. (2009). Brown lines delineate interpreted normal faults along the NW Calabrian margin, while red lines are compressional features across the Inner plateau and the Pre-Messinian wedge observed by Polonia et al. (2011). Thick black lines are wide-angle seismic boundaries constrained by the tomographic model in this study. Thin black lines are isovelocity contours shown in Fig. 3.