Global and Planetary Change 263 (2026) 105486

Contents lists available at ScienceDirect

Global and Planetary Change

journal homepage: www.elsevier.com/locate/gloplacha

Integrated gravity modelling of the onshore-offshore transect Ligurian Sea – Po Basin: Revisiting the lithosphere architecture of the Northern Apennines (Northern Italy)☆

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ARTICLE INFO

Editor: Dr. Sun Jimin

Keywords:
Gravity modelling
Orogenic system
Dismembered orogenic Wedge
Northern Apennines Lithosphere
Alps-Apennines orogenic System
Inheritance in Orogenic system

ABSTRACT

Orogenic system, which may often have onshore and offshore portions, requires different methodologies of investigation that, on their hand, provide results not always easy to compare. This is a drawback that typically hampers the reconstruction of unifying geological models of the orogenic belts and associated sedimentary basins. These limitations are even more evident in orogenic systems that were affected by late synorogenic extension such as the Apennines of Italy. In the attempt to create a unifying geological model of the Northern Apennine orogen, we integrated onshore geology, marine geology with geophysical data (magnetic and gravimetric anomalies, heat flow, refraction seismology, surface wave tomography, particular noteworthy seismic reflection profiles) and previous lithospheric studies to constrain a new gravity model along a transect that crosses the whole orogenic system from the Ligurian Sea, through the northwestern portion of the Northern Apennines adjacent to the Alps (Ligurian knot), to the Po Basin, up to the Southern Alps of Italy. The new gravity model fits with densities of a lithosphere made of different crustal types: from extended continental crust, 18 km thick in the offshore oceanic domain of Ligurian Sea, to the continental crust ≤40 km-thick including up to ~20 km sedimentary cover beneath the Po Basin in the onshore continental domain. In between, the offshore-onshore transition domain, ~150-km wide and up to 25 km thick, includes most of the inner Northern Apennines' thrust stacks which are thinned and dismembered by extensional faulting. Within this transition domain, around ~20 km depth, an eastward-dipping High-Velocity/high-Density Body (HVDB) is highlighted. It abruptly ends within the Ligurian Sea, where it is ~7 km thick, progressively thins out toward the NE and disappears at ~36 km depth beneath the watershed zone of the Northern Apennines. The new onshore-offshore lithospheric transect provides a unifying and updated geological model of the Northern Apennines and sheds new lights on their complexity in relationship with the colliding Europe and Africa continental plates that shape the Alpine-Apenninic orogenic system in the Mediterranean Region.

1. Introduction

The building up of orogenic wedges shapes the Earth surface by creating and controlling a variety of sedimentary basins with onland and marine portions (Einsele, 2000; Schellart et al., 2007; Frisch et al., 2011; Allen and Allen, 2013; Ju et al., 2022). The foreland basins system are the typical sedimentary basins associated to the continental collisional orogen which may suffered coeval or later extensional phases associated

☆ This article is part of a Special issue entitled: '**Tectonics, sedimentation and magmatism in sedimentary basins – processes and societal relevance**' published in Global and Planetary Change.

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https://doi.org/10.1016/j.gloplacha.2026.105486
Received 31 October 2025; Received in revised form 10 April 2026; Accepted 13 April 2026
Available online 21 April 2026
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to the formation of intramountain and/or back-arc basins (Kingston et al., 1983; Ingersoll, 1988; Ingersoll and Busby, 1995; Einsele, 2000; Kearey et al., 2013; Allen and Allen, 2013; Evenick, 2021; Ju et al., 2022), all these basins being framed within the Wilson's cycle of variable time extents (Woodcock, 2004; Wilson, 2019). This implies that some orogenic system may have portions offshore and onshore which are investigated with different methodologies: e.g. offshore are marine geology surveys with remote analysis and point sampling (core or borehole) generally surficial but can reach variable depths beneath the sea floor, marine seismic reflections; e.g. onshore are field geology surveys, onshore seismic reflection and point sampling (core or borehole) that can reach kms depths beneath the topography. These methodologies provide data that are not always comparable because of their different resolutions or analyzed by different research groups expert in those specific methodologies and/or investigated area.

Taking as a case study the extensively studied foreland basins system and back-arc basins associated with the Northern Apennines of Italy (among the others, Bartole et al., 1991; Bartole, 1995; Mauffret et al., 1999; Fiorentino et al., 2025 for offshore portions; Molli et al., 2010 and Conti et al., 2020 for onshore portions), we propose a new offshore-onshore lithospheric transect with the aim to generate a unifying geological model of the northwestern sector of the Apenninic orogenic system and associated sedimentary basins. Along a 465 km long and SW-NE trending section, we collected available geological data (Bernini et al., 1990; Argnani and Ricci Lucchi, 2001; Bernini and Papani, 2002; Carlini et al., 2013; Molli et al., 2018), seismic reflection profiles (Pieri and Groppi, 1981; Cassano et al., 1986; Argnani and Ricci Lucchi, 2001; Carlini et al., 2013) published in conferences (Camurri et al., 2001; Rizzini et al., 2006), and other geophysical data (Cassinis et al., 2005; Grad et al., 2009; Maus et al., 2009; Brandmayr et al., 2010; Di Stefano et al., 2011; Förste et al., 2014; Fuchs et al., 2021) including seismic refraction (Blundell et al., 1992; Makris et al., 1999; Scrocca et al., 2003; Cassinis et al., 2005; Finetti, 2005; Solarino and Cassinis, 2007).

All the collected data were integrated to constrain a new gravity model from the Ligurian Sea (offshore) through the northwestern portion of the Northern Apennines up to the Po Basin and the Verona province (onshore) (Fig. 1). The Ligurian Sea and the northwestern portion of the Northern Apennines being adjacent to the termination of the Alps in the so called "Ligurian Knot" (Laubscher et al., 1992). The new integrated gravity model and lithospheric cross section allowed us to revisit the geological model of the northwestern portion of the Apennines orogenic system. A complete suite of variably extended continental crustal types and sedimentary basins changing through time highlight and track the complex and variable architecture of the Apennines orogenic system. The revisited geological model allowed also to discuss the paleogeographic evolution of the northwestern portion of the Apennines orogenic system which was created by the long-lived, and still ongoing, convergence history between Europe and Africa plates (Stampfli and Borel, 2002, 2004; Handy et al., 2010; Marroni et al., 2010; Van Hinsbergen and Schmid, 2012; Le Breton et al., 2017; Marroni et al., 2017).

This history includes: 1) the convergence- and collisional-related deformations and basins formed during the closure of the Piedmont-Ligurian Ocean (Early/Late Cretaceous to Eocene), previously separating Europe from Adria plates; 2) the asthenospheric flows and related rifting which created the present-days Ligurian Sea (Oligocene-Middle Miocene) and the Tyrrhenian Sea (Late Miocene-Present); 3) the almost coeval (Oligocene-Present) formation of foreland basins system, one of the most recent foredeep of them being located in the Po Basin since the latest Messinian.

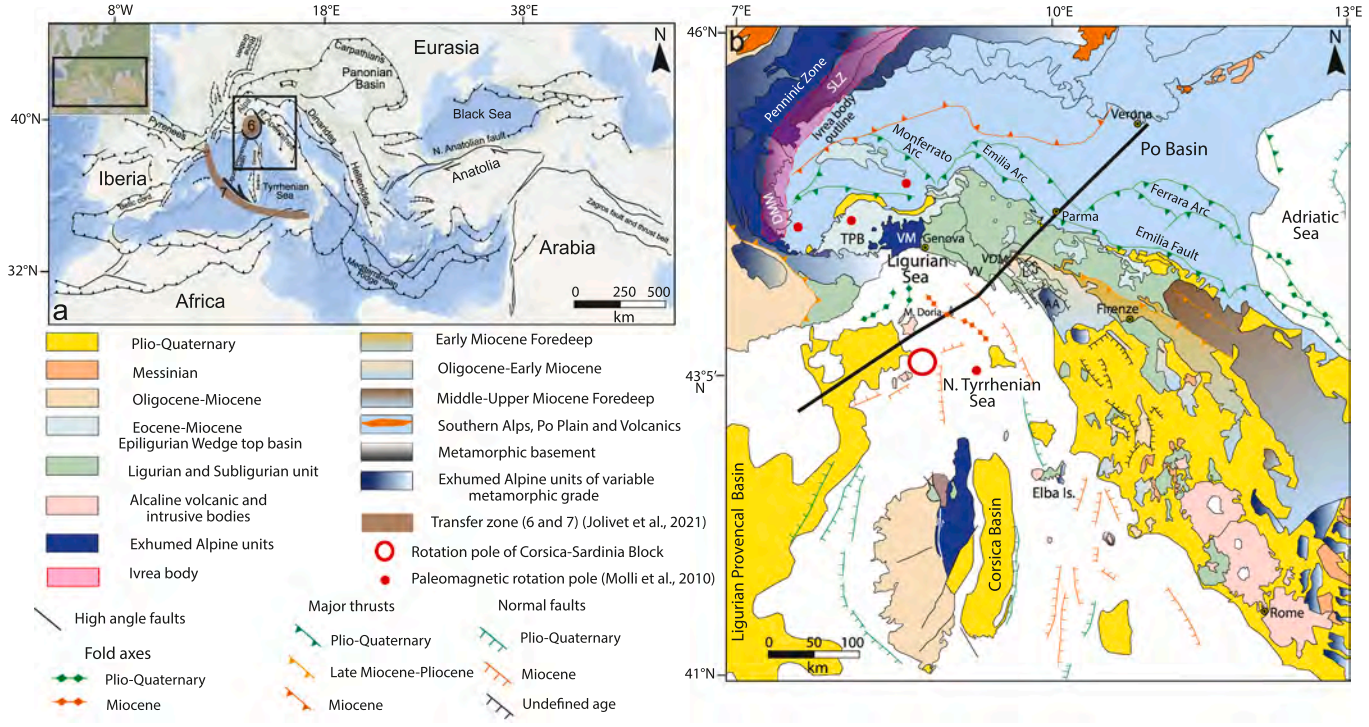


Fig. 1. a) Location of the study area in the tectonic framework of the Mediterranean Region with the inset after GeoMapApp (www.geomapapp.org on 08 October 2025) (Ryan et al., 2009). b) Regional geological synthesis (modified after Molli et al., 2010 and Camurri et al., 2001). The location of the rotation pole of the Apennine in the Ligurian Sea is reported (red circle after Réhault et al., 1984; Speranza et al., 2002; Gattacceca et al., 2007) and is close to one of the transfer zones segmenting the Mediterranean back-arc basins (number 6 of Jolivet et al., 2021). The other rotation poles (small red dots) are from Molli et al. (2010). The map is a synthesis of various works (Bigi et al., 1990; Bartole et al., 1991; Laubscher et al., 1992; Bigi et al., 1992; Makris et al., 1999; Mauffret et al., 1999; Rollet et al., 2002; Cassinis et al., 2005; Carminati et al., 2010; Lustrino et al., 2011; Conti et al., 2020; Fiorentino et al., 2025). AA – Alpi Apuane, DMM – Dora Maira Massif, SLF – Sesia Lanzo zone, TPB - Tertiary Piemont Basin, VM – Voltri Massif, LFS – Lunigiana fault system, VM – Val di Magra graben, VV – Val di Vara graben. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

These geodynamic processes are subsequent one to the other in the order they are listed above and, consequently, they created the sedimentary basins whose infill is strictly related to surficial processes, either marine or continental, acting at the corresponding times. The geodynamic evolution and the related sedimentary basins are commonly framed within the slab roll-back of the subducting lower plate (Adria, i.e. northern portion of Africa) and consequent extension of the upper plate (Europe) including the dismemberment of the earlier Alpine-Apenninic orogenic wedge (Doglioni, 1991; Gueguen et al., 1998; Faccenna et al., 2001; Carminati and Doglioni, 2005).

Although the new lithospheric cross section conforms with the geodynamic framework outlined above, it also allows a re-evaluation of the processes affecting the two colliding and oppositely subducting plates, Europe and Adria. These processes are responsible for the development of the Alpine-Apenninic orogenic system, including the possible remnants of the earlier (Alpine) lithospheric assemblages not fully overprinted by the most recent (Oligocene-Present) geodynamic events.

2. Geological overview

The geodynamic evolution of the study region is part of the Mediterranean area (Fig. 1) which results from a long geological history started from Permian-Mesozoic Tethyan rifting, separating Europe from Africa, evolving through lithosphere stretching, passive continental margin and continent-ocean transitions development, and oceanization in the Western Ligurian Tethys (Stampfli and Borel, 2004; Handy et al., 2010; Van Hinsbergen et al., 2014; Le Breton et al., 2017; Van Hinsbergen et al., 2020 and reference therein).

Starting in Late Cretaceous, plate motion reversal and convergence (e.g. Boccaletti et al., 1971; Faccenna et al., 2004; Schmid et al., 2004; Marroni et al., 2017) reworked extensional structures by triggering positive inversion of former major extensional fault systems and uplifting and exhuming former rifting domains, which are still locally preserved in the tectonic architecture of the Alpine-Apenninic orogenic system (Manatschal and Müntener, 2009; Mohn et al., 2010; Agard, 2021). The Alpine-Apenninic orogenic system was initially a double vergent orogen with two fold-and-thrust belts (Doglioni, 1991; Gueguen et al., 1997; Carminati and Doglioni, 2012): i) the Europe–vergent fold-and-thrust belt, or Alps prowedge related to the subducting Europe plate, and ii) the Africa-vergent fold-and-thrust belt or Alps retrowedge related to the shortening of the upper Africa plate and forming the early Southern Alps and Apennines. Orogenic evolution likely progressed through two diachronous oppositely dipping phases of subduction: a Cretaceous east-dipping “Alpine subduction”, and an Eocene-Oligocene to Present west-dipping “Apennine subduction”. The latter initiated along the retrowedge of the Alps, where oceanic or thinned continental lithosphere was present (Doglioni, 1991; Molli et al., 2010; Marroni et al., 2017; Carminati and Doglioni, 2012). Most of the oceanic lithosphere of the Ligurian Tethys was consumed by subduction and dragged deep into the mantle before being partly exhumed. Ophiolite tectonic slices are preserved in the Western Alps and in the North-Western Apennines, either as high grade metamorphosed slices (Penninic zone of the Western Alps and the Voltri Massif) or as low grade to non-metamorphic nappes (Ligurides in the NW Apennines) (Scafidi et al., 2009; Malusà et al., 2015, 2021; Schmid et al., 1989, 2017; Vignaroli et al., 2010; Malatesta et al., 2012; Starr et al., 2020) marking the suture zone of the collided Europe and Adria plates.

Apenninic subduction produced an orogenic belt characterized by lower topography, shallower decollement, and deeper foredeep (Doglioni, 1994; Carminati and Doglioni, 2012; Faccenna et al., 2001), and has being associated with slab retreat, which triggered back-arc basin opening and development of the present Western Mediterranean basins (Réhault et al., 1984; De Voogd et al., 1991; Gorini et al., 1993, 1994; Pascal et al., 1993; Mauffret et al., 1995; Gueguen et al., 1998; Guennoc et al., 2000; Van Hinsbergen et al., 2014; Jolivet et al., 2021; Maillard et al., 2020). Since about 30 Ma, the back-arc opening favored the separation of the Corsica-Sardinia block from the European margin and its counterclockwise rotation by about 60°, forming the thinned crust of the Ligure-Provençal Basin (Gueguen et al., 1997; Carminati and Doglioni, 2012). At about 15 Ma, back-arc extension shifted eastward of the Corsica-Sardinia block, causing the opening of the Tyrrhenian Basin (Malinverno and Ryan, 1986; Kastens and Mascle, 1990; Zitellini et al., 2023; Sanfilippo et al., 2025).

The pole of rotation of the Corsica-Sardinia block is located within the junction area between Western Alps and Northern Apennines and the adjoining part of the northern Ligurian Basin (Vignaroli et al., 2008; Molli, 2008; Molli et al., 2010; Jolivet et al., 2021; Fig. 1). This pole accommodates the transition between the westward Apennine subduction and the eastward Alpine subduction and it is coeval to counterclockwise rotation of the Alps and the Northern Apennines (Maffione et al., 2008; Vignaroli et al., 2008; Crouzet et al., 2025) along the transfer zone evidenced by Jolivet et al. (2021) (their transfer zone number 6). Associated with this pole of rotation, to the south, there is also the Catalan-Baleares-Sicily Transfer Zone, which is the longest in the Mediterranean (Fernàndez et al., 2019; Jolivet et al., 2021) (Fig. 1a). It also accommodates the progressive rotation of the Corsica-Sardinia block and then the progressive opening of the southern Tyrrhenian Sea. This transfer zone starts east of the Pyrenees, as the Catalan Fracture Zone, then becoming the North Baleares transfer zone and reaches the north of Sicily (transfer zone 7 in Jolivet et al., 2021).

Tyrrhenian extension started affecting the Northern Apennines in Late Miocene times with the activation of low- and high-angle extensional fault systems, development of NW-SE-striking grabens and exhumation of the deep metamorphic units such as the Apuane Alps (Argnani et al., 2003; Fellin et al., 2007; Eva et al., 2014; Molli et al., 2010) (Fig. 1b). In particular, asymmetric low-angle extensional faulting is consistent with the eastward rototranslation kinematics of the Northern Apeninnes (Boccaletti et al., 1990; Elter and Pertusati, 1973; Faccenna et al., 2001; Le Breton et al., 2017; Molli et al., 2018) and fits with the present-day tectonics (Carminati and Doglioni, 2012; Facenna et al., 2014) as well as the seismic activity of both extensional and thrust fault systems (Eva et al., 2014, 2025). The eastward migration of the paired contractional and extensional belts of the Apennines (Elter, 1975; Carminati and Doglioni, 2012) determined the eastward shift of the foreland basins system (Castellarin et al., 1985; Ricci Lucchi, 1986; Boccaletti et al., 1990; Argnani and Ricci Lucchi, 2001; Carminati et al., 2003; Fantoni and Franciosi, 2010; Turrini et al., 2016). The younger foredeep basin is located in the Po Plain and filled by Plio-Pleistocene turbidite and deltaic sequences from both the Southern Alps and the Northern Apennines (Argnani and Ricci Lucchi, 2001; Rossi et al., 2002, 2015; Argnani et al., 2003; Ghielmi et al., 2010; Ghielmi et al., 2013). Major segmentation of the outer deformation front characterizes the Northern Apennines, producing the Monferrato, Cremona, and Ferrara salients, the latter two being separated by the Parma recess and representing the most external front of the Apennines (Pieri and Groppi, 1981; Livani et al., 2018) (Fig. 1).

Crustal thickening in the Northern Apennines progressed by major duplexing among adjacent paleogeographic domains and their internal contractional deformations, favored by few main decollement levels, including the upper Triassic evaporites and the shaly formations underlying the foredeep turbidites (Livani et al., 2018). Out-of-sequence thrusting concurred to accommodate crustal shortening, causing imbrication of metamorphic thrust sheets onto non-metamorphic ones (Molli et al., 2018). As a whole, the Northern Apennines are considered as a thick-skinned thrust system, but two main decollement levels – the basal decollement (in the Carnian units) and a shallower one in the Late Eocene-Oligocene/Miocene formations complicate this tectonic style by creating thin-skinned structures (Livani et al., 2018).

As a result of the time and space complexity of Cretaceous to recent evolutionary history, the three-dimensional tectonic architecture of northern Apennines is on the prosecution of the Ligurian Alps which shares similar HP/LT rocks present in the Western Alps and in Corsica

(Alpine Corsica) (Fig. 1). The alpine high grade metamorphic rocks end East of the Voltri group, along the Sestri-Voltaggio lineament (Molli et al., 2010), which together with the northwestern portion of Northern Apennine belong the so called Ligurian Knot (Laubscher et al., 1992). This complex knot of the Alpine-Apenninic orogenic system extends offshore, toward Corsica, and it will be extensively discussed after presenting the results of this work.

2.1. Previous lithospheric scale studies

During the last ~25 years, the study region has been investigated actively by geophysical methods, mainly seismic methods and their integration (Cassano et al., 1986; Makris et al., 1999; Contrucci et al., 2001; Rollet et al., 2002; Cassinis et al., 2003). The Moho depth models were derived from active and passive seismic experiments for Italy (Cassinis et al., 2005; Di Stefano et al., 2011; Spada et al., 2013) and implemented with gravity inversion for the Ligurian Basin (Chamot-Rooke et al., 1999).

In the '80-'90, the CNR-ENI-ENEL Deep Crust Seismic Profiles, acquired in the frame of the CROsta Profonda (CROP) Project, provided the crustal configuration of Central Mediterranean. The project was aimed to understand the main geodynamic processes responsible for the current geological arrangement of Italy, the definition and prevention of geological hazard as well as the search for energy resources as hydrocarbons and geothermal fields. A set of Multichannel Seismic profiles were acquired; they cover marine and terrestrial areas and they are available at http://www.crop.cnr.it/, in atlas with uninterpreted (Scrocca et al., 2003) and interpreted (Torelli, 1994; Finetti, 2005) profiles.

Local seismic tomography provided construction of 3D seismic velocity lithospheric models of high reliability and resolution for the Alps down to the depth of 60 km (Diehl et al., 2009). Local tomography study in West Alps (Dora-Maira dome) was validated by receiver function analysis results (Solarino et al., 2018, 2024). The new receiver function analysis, actively performed recently in the Western Alps, was validated and complemented by a S-wave velocity model from ambient-noise tomography (Nouibat et al., 2022, 2023). These studies provide additional constraints for better understanding the involvement of remnants of Variscan suture zones in the Alpine subduction zone prior to ultra-high-pressure (UHP) rock exhumation (Paul et al., 2022).

In the Ligure-Provençal basin, or North Ligurian Sea margin, the recent offshore active seismic reflection-refraction and wide-angle profiles SEFASILS (Dessa et al., 2020) and LOBSTER-P02 (Dannowski et al., 2020) were acquired to study the recent tectonic regime and to specify the type of the crust which resulted to be an extended and very thin continental crust or exhumed partially serpentinized mantle. The above geophysical studies constitute the database for constructing an integrated 3D crustal-scale geomodel of the Western Alps and Ligurian Basin (Paul et al., 2024).

Mantle tomography has been extensively used for studying the upper mantle structure of Italy and delineating the geometry of slabs and lithosphere-asthenosphere system (Panza et al., 1992; Piromallo and Morelli, 2003). The thickness of the lithosphere in Italy varies strongly from 30 to 50 km in the Tyrrhenian Sea to 130 km below the Alps. In the Apennines, it changes from 50 to 70 km in the north to ~100 km in the Calabria (Panza et al., 1992). Recently, the structural model for the lithosphere-asthenosphere system of Italy was investigated by means of Vs-depth distribution. This method provides a usable tool to constrain the model (plotted on grid 1° × 1°) as an assemblage of juxtaposed selected cellular models (Brandmayr et al., 2010). This model, constructed for the Western Mediterranean, shows the easterly uprising of the asthenosphere to ~30 km depth in the Tyrrhenian Sea suggesting the easterly-directed upper mantle circulation in the Western Mediterranean (Panza et al., 2007).

3. Data and methods

To construct the lithospheric transect from the Ligurian Sea to the Po Basin (Fig. 2), various geological and geophysical datasets were integrated.

3.1. Geological data

The main geological, tectono-stratigraphic units and structures are derived from published works (Bigi et al., 1990; Bigi et al., 1992; Conti et al., 2020) and synthesis of various authors (Molli et al., 2010, 2016; Schmid et al., 2017; Bartole et al., 1991; Laubscher et al., 1992; Mauffret et al., 1999; Rollet et al., 2002; Carminati et al., 2010; Fiorentino et al., 2025). These data are compiled in Fig. 1 to create a regional synthesis map of the onshore and offshore area of the Apenninic orogenic system and part of the Alpine one (Western Alps and Southern Alps).

3.2. Geophysical data

Seismic reflection profiles are geophysical investigation of particular interest for linking geological data with other geophysical data (Figs. 2–5). Some of these seismic profiles were published in conferences and thesis (Camurri, 2000; Rizzini et al., 2006; Rizzini, 2007) and published here for the first time (Figs. 6–8).

The geological constraints were used to interpret the three seismic reflection profiles whose acquisition and processing parameters, where available, are reported in Table 1. Seismic reflection profile CROP M-11 (Scrocca et al., 2003), or offshore **Line 1** (Fig. 2), was reprocessed, time

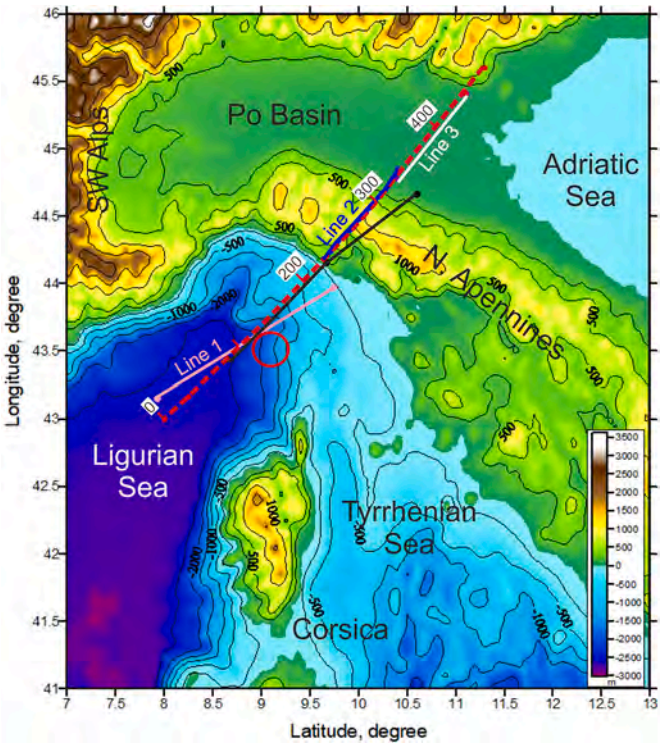


Fig. 2. Topo-bathymetry and trace of the lithospheric transect Ligurian Sea-Po Basin (thick red dashed line) and of the composing seismic reflection profiles: a) Line 1 or CROP M11 (lilla colour); b) Line 2 (blue colour) onshore Northern Apennines regional cross-section, kindly provided by E.N.I. (Camurri, 2000; Camurri et al., 2002; Molli et al., 2018); c) Line 3 (white colour) integrated geological-geophysical Po Plain cross section (Cassano et al., 1986; Pieri, 1983); d) the offshore-onshore wide-angle reflection and refraction profile from the Ligurian Sea to the Northern Apennines (black line) after Makris et al. (1999). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

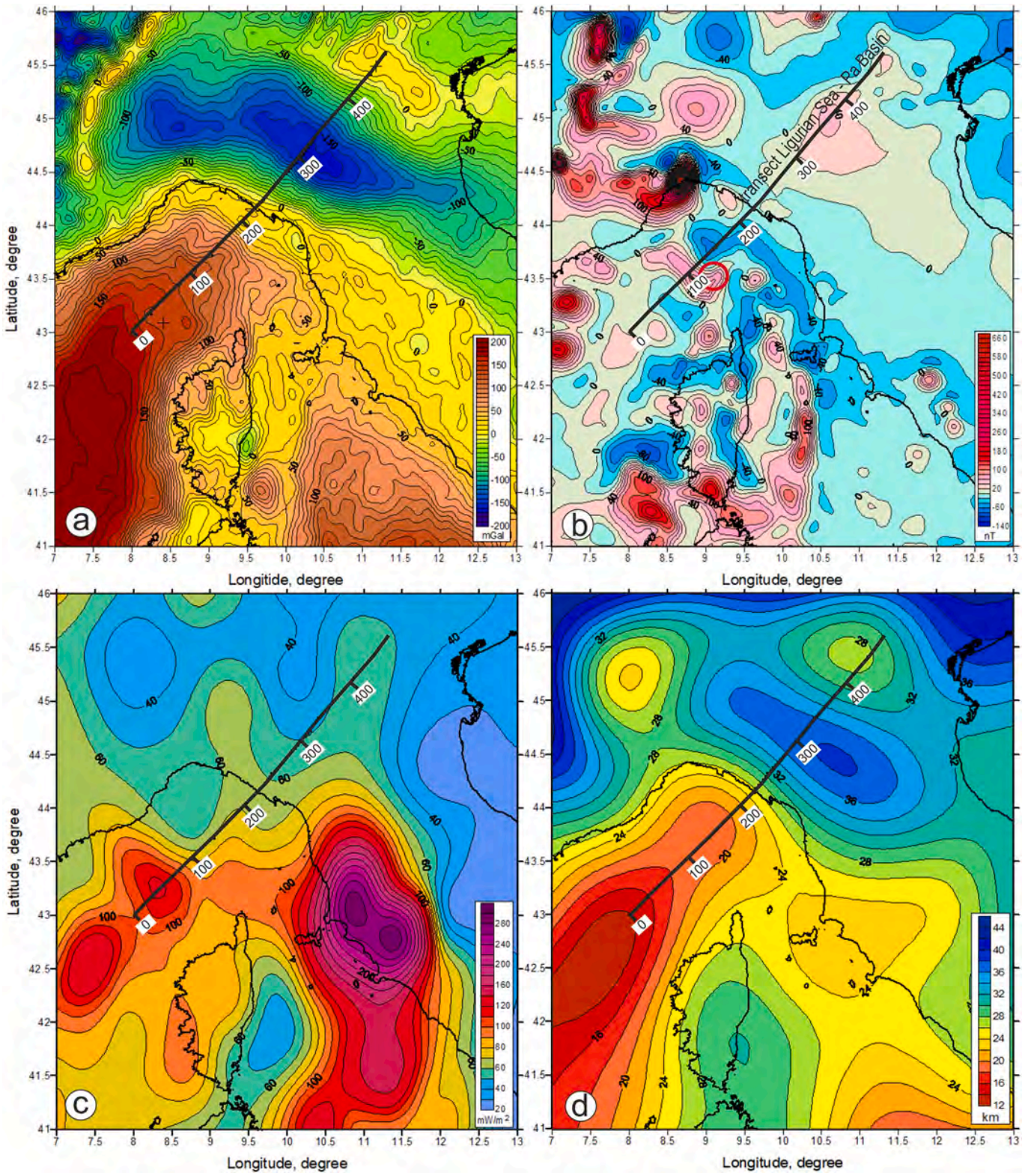


Fig. 3. Geophysical fields and the Moho depths map for the region of the Ligurian Sea - Po Basin transect. a) Gravity map, Bouguer anomalies from satellite model EIGEN6C4 (Förste et al., 2014; https://icgem.gfz-potsdam.de/calcgrid). b) Magnetic map from EMAG2 global magnetic model (Maus et al., 2009; www.ngdc.noaa.gov/geomag/emag2.html; http://geomag.org/models/EMAG2). Red circle shows the location of the rotation pole in the Ligurian Sea according to Réhault et al., 1984; Speranza et al., 2002; Gattacceca et al., 2007. c) Surface heat flow density from Heat Flow Data Base (Fuchs et al., 2021). d) Moho depth map (Grad et al., 2009). The black solid line is the trace of the updated lithospheric cross section Ligurian Sea – Po Basin. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

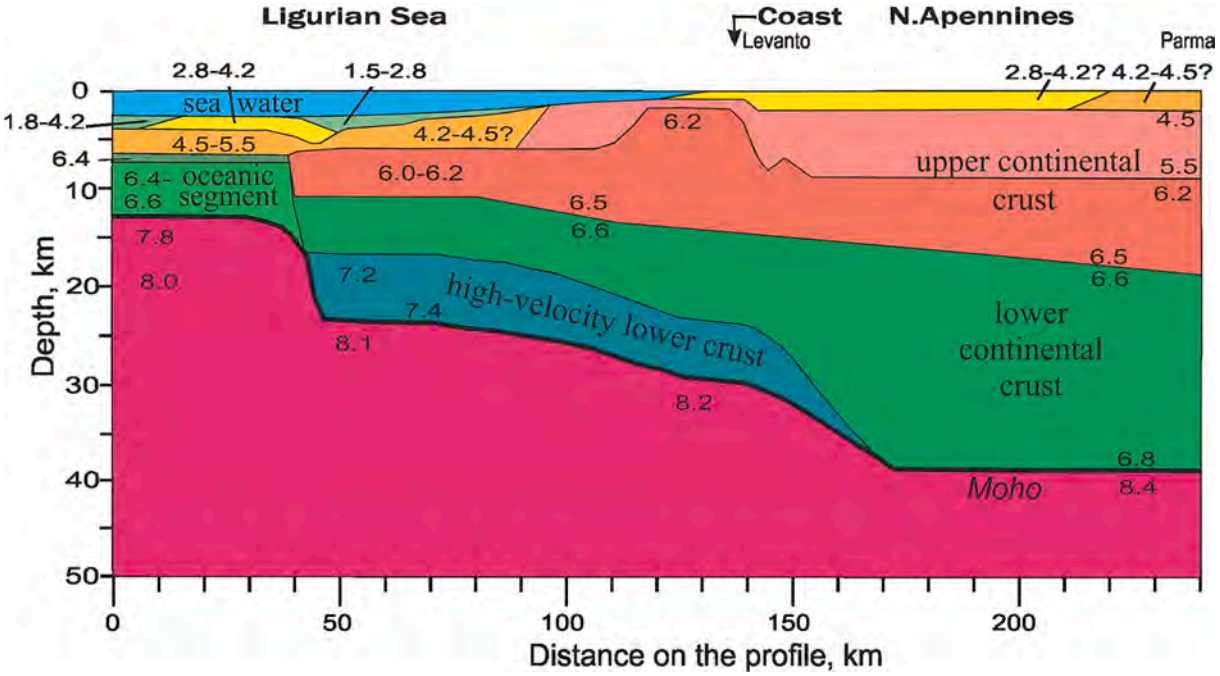


Fig. 4. Velocity model on the wide-angle reflection and refraction seismic profile from the Ligurian Sea to Northern Apennines (modified after Makris et al., 1999). Note the body with P-wave velocity 7.2–7.4 km/s above the present-day Moho (solid black line corresponding to the highest velocity contrast). It corresponds to the high-velocity lower crust. Note that this body is low-velocity with respect to the lithospheric mantle. It corresponds to the High-Velocity/high-Density Body (HVDB) presented later in this work (see Fig. 9 and Chapter 6. Gravity/Density model for the crust and upper mantle).

migrated and time-to-depth converted with Seismic Unix following the finite-difference full-wave post-stack migration which uses the time scale variant for depth with further conversion into the depth section as developed in the S. Subbotin Institute of Geophysics of National Academy of Sciences of Ukraine (Verpakhovska et al., 2022). This migration method, based on the finite-difference solution of the wave equation, implements wave field continuation, differs from that used in finite-difference reflection/refraction migration (Verpakhovska et al., 2025) and provides the velocity model, the migration parameters as well as the final migration image (Fig. 6b).

Seismic reflection profile and regional cross-section (see Camurri, 2000; Camurri et al., 2001) (**Line 2** in Figs. 2, 7) is time-to-depth converted according to interval velocities derived from stacking velocities of 5 CDP selected according to significant changes in the vertical velocity profile along the seismic profile; velocities ranging in between 2 and 5 km/s. In addition, five control wells (Pontremoli_01; Lama dei Cerri; Terenzo; Sivizzola, Vigatto) publicly available at ViDEPI database (Progetto ViDEPI - Visibilità dei dati afferenti all'attività di esplorazione petrolifera in Italia - ViDEPI), were transformed in time and tied to the seismic profile. The wells reach maximum 3 km depth from the surface and together with the five velocities vertical profiles they constrain the time-to-depth conversion.

Seismic reflection profile across the Po Plain, shown as the north-eastern **Line 3** in Figs. 2 and 8, was already converted in depth, integrated with all the geophysical data (gravity, magnetic, surface heat flow data) and hydrocarbon exploration wells as collected by Cassano et al. (1986). This interpreted profile was only redrawn and adjusted according to the stratigraphic units and geological domains of the other profiles.

The three seismic profiles were merged and integrated with other geophysical data which constrain the lithospheric architecture and characters around the composite cross section (Fig. 9). It has been created a compilation of gravity field model (EIGEN6C4 of Förste et al., 2014, https://icgem.gfz-potsdam.de/calcgrid; Zahorec et al., 2021), magnetic field data (EMAG2 of Maus et al., 2009; www.ngdc.noaa.gov/geomag/emag2.html; http://geomag.org/models/EMAG2), surface heat flow (Fuchs et al., 2021), crustal thickness (Grad et al., 2009; Cassinis et al., 2005; Di Stefano et al., 2011; Spada et al., 2013), Wide-Angle Reflection and Refraction (WARR, Makris et al., 1999), passive seismology based on the surface wave tomography (Brandmayr et al., 2010) or different types of waves (Panza et al., 1992; Panza et al., 2007; Piromallo and Morelli, 2003; Scafidi et al., 2009; Di Stefano et al., 2011; Solarino et al., 2018, 2024; Nouibat et al., 2022, 2023).

The geological data and the compilation of geophysical data provide the constraints used for the gravity modelling which will be described in Chapter 6. Gravity/density model for the crust and upper mantle.

4. Geophysical data compilation

4.1. Geophysical fields

The **gravity data** (Fig. 3a) are represented by the Bouguer anomalies of combined global gravity field model EIGEN6C4 (Förste et al., 2014; https://icgem.gfz-potsdam.de/calcgrid). This model is a combination of satellite gravity missions LAGEOS, GRACE, and GOCE augmented with the surface gravity data (EGM2008 model), which provides to reach very high resolution of the gravity field (up to degree and order 2190). The EIGEN6C4 model has been the first combined global gravity field model at so high resolution. Although a generalization of complete Bouguer anomalies from terrestrial data is available for the Alpine region (AlpArray Gravity Research Group -AAGRG- part of the European AlpArray program; Zahorec et al., 2021), we used the EIGEN-6C4 model for our modelling because of the discrepancy between the latter and the Alps-Array data base. In our study area, the discrepancy reaches ~4 mGal and, in the Ligurian Sea, the Alps-Array used the sparse network of ship-born marine gravity data for filling gaps of the EIGEN-6C4 model (Zahorec et al., 2021).

These gravity data show two contrasting anomalies of different sign in our region: the high-amplitude (160 mGal) gravity high over the Ligurian Sea and northern part of the Tyrrhenian Sea, which contrasts the negative gravity low onshore (−150 mGal) (Fig. 3a). The latter correlates well with the low of geoid heights (Carminati and Doglioni,

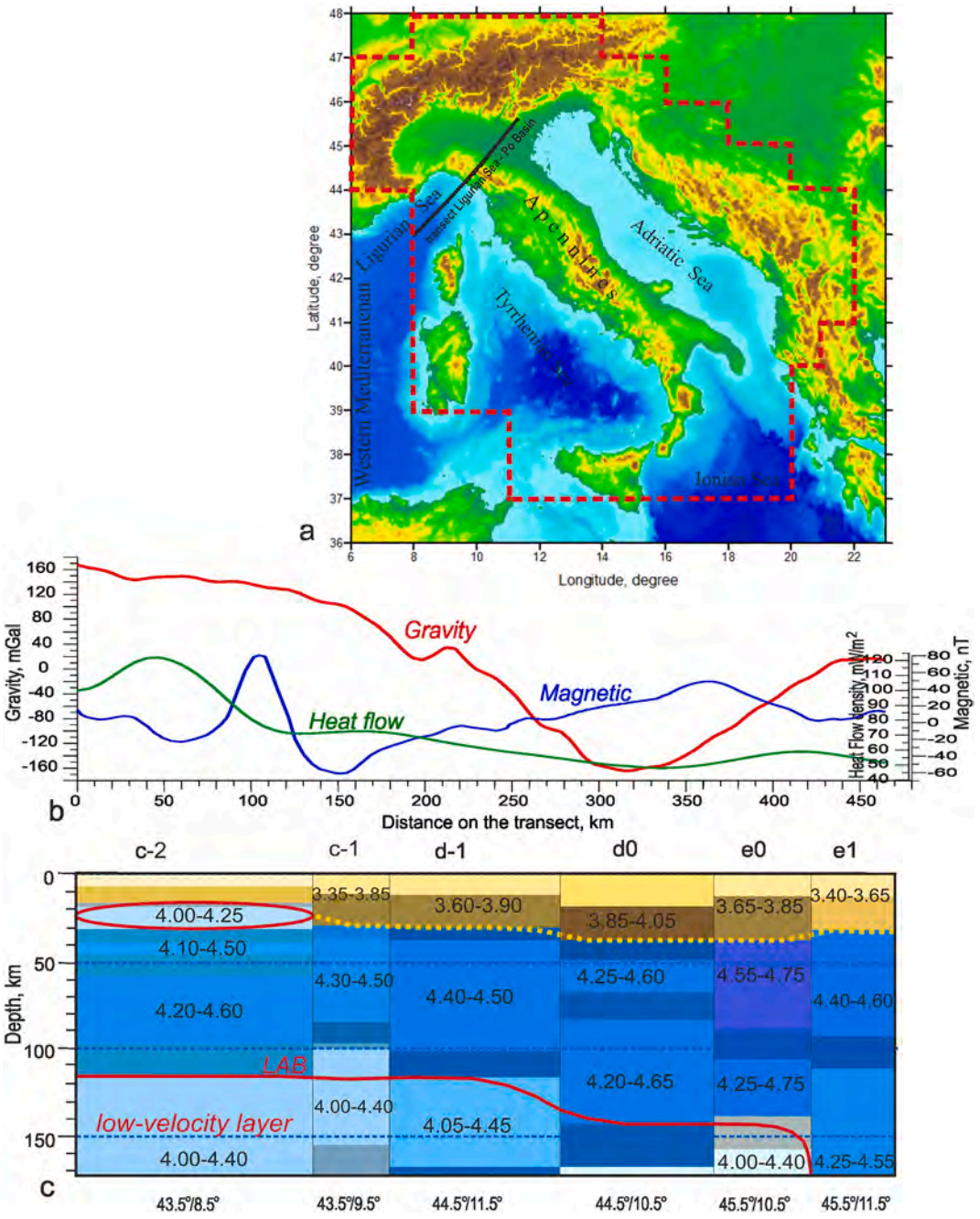


Fig. 5. Map of the studied area (a) covered by surface wave tomography of Brandmayr et al. (2010) outlined by red dashed line and the location of our transect Ligurian Sea-Po Basin (solid black line) with (b) the potential fields curves as well as heat flow density along the transect and (c) the S-wave velocity model to the depth of 180 km. The cellular grid is omitted in (a) for the sake of clarity. Numbers in (c) denote the S-wave velocities in km/s constructed for the Liguria Sea-Po Basin transect. The red solid line highlights the low-velocity layers. One in the subcrustal layer of the Ligurian Sea, at depth of 20–30 km; another under the entire transect at a depth of ≥120 km likely corresponding to the LAB (lithosphere-asthenosphere boundary. The numbers above and below the seismic tomography cross-section (c) correspond to the labels of the cellular columns of the seismic tomography model and their coordinates respectively as derived from Brandmayr et al. (2010). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

2012) that could be caused (at least in part) by the great thickness of the sedimentary and metasedimentary successions of the embricated thrust stack and foredeep infill in the Po Basin.

Between these two huge gravity anomalies of opposite sign, there is a wide gradient zone located south of ~44° latitude and average values varying from −30 to +30 mGal (Fig. 3a). It is trending NW-SE and corresponds to the area of extensional tectonics testified by the occurrence of normal faults in the Northern Apennines (Fig. 1).

In the NW corner of Fig. 3a, it is also visible the stripe of gravity high (up to +30 mGal), NNE - SSW directed and corresponding to the Ivrea body (Boriani and Sacchi, 1973; Ansorge et al., 1979; Kissling, 1993; The ECORS-CROP Gravity Group et al., 1989; Zahorec et al., 2021) which is clearly distinguishable from the background of the mosaic pattern of negative gravity anomalies of the Western Alps.

The **magnetic field data** (Fig. 3b) were taken from a global Earth Magnetic Anomaly Grid (EMAG2) of 2 arc min resolution and 4 km altitude above geoid (sea level) (Maus et al., 2009; www.ngdc.noaa.gov/geomag/emag2.html; http://geomag.org/models/EMAG2). EMAG2 is a global grid of magnetic anomaly data derived from satellite, ship, and airborne magnetic measurements. The main magnetic field

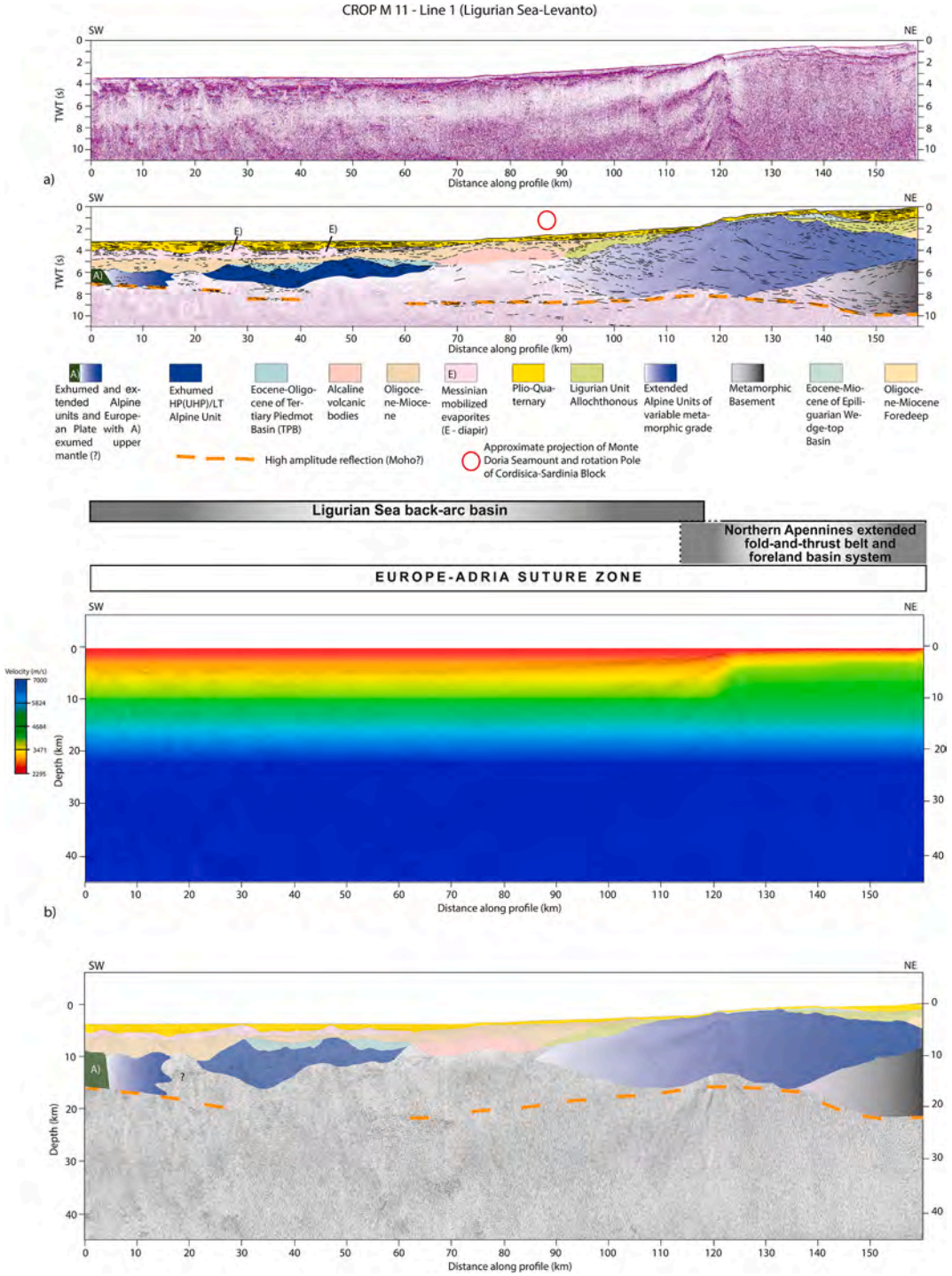


Fig. 6. Offshore seismic profile Line 1 (Ligurian Sea-Levanto) a) original data without interpretation and with interpretation (in time – TWT – section) and b) velocity field applied for post-stack time-to-depth conversion of the seismic reflection profile and interpreted seismic reflection profile depth converted (see text chapter 3. Data and methods for the time-to-depth conversion). Interpretation in a) and b) is based on Torelli (1994) and Rizzini et al. (2006).

heterogeneities are concentrated in the region of Corsica and adjacent parts of the Ligurian and Tyrrhenian Seas. They are marked by complex anomalies of different signs, sizes and configurations, which form a wide (up to 250 km) domain of magnetic crustal inhomogeneity centered on the Corsica Island (Fig. 3b). This block is traced northwards to the northern part of the Ligurian and Tyrrhenian Seas where it is bounded, to the east, by a N-S trending gradient zone in the Tyrrhenian Sea passing through the Elba Island (Fig. 3b). The strong magnetic heterogeneity of this crustal domain can be related to the origin of the Corsica-Sardinia block. This block, rifted and separated from the European continental margin, rotated counterclockwise to reach its present position resulted in the opening of the Ligurian Sea, to the west, and the Tyrrhenian Sea, to the east (Carminati and Doglioni, 2012). Thus, in the magnetic field we can see the imprint of two back-arc basins – Ligurian Sea and Tyrrhenian Sea – bounding the Corsica-Sardinia block.

The magnetic field pattern of the Ligurian Sea shows disordered magnetic anomalies; whereas, in the Tyrrhenian Sea, the lack of magnetic anomalies is apparently due to the high thickness of sediments

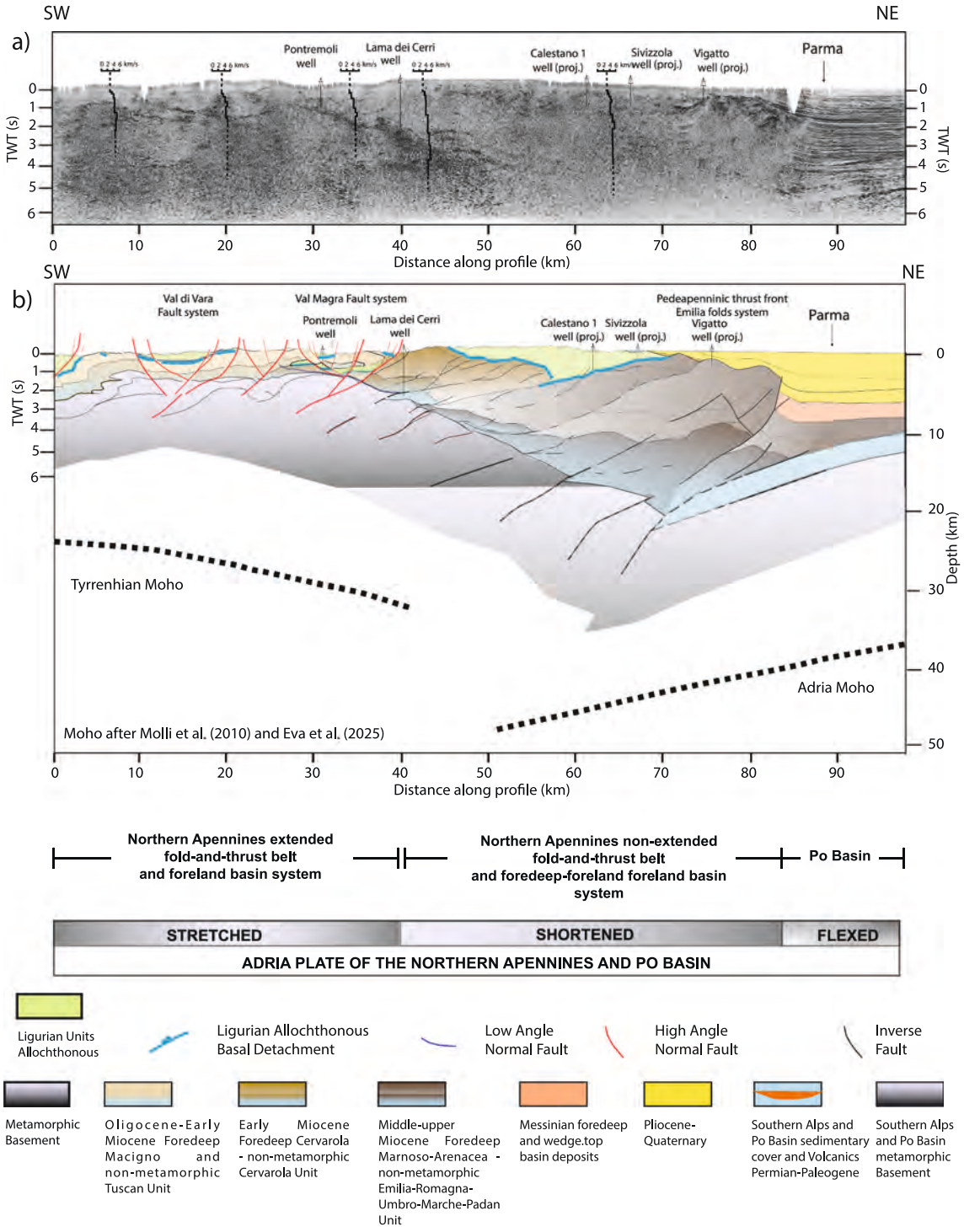


Fig. 7. Onshore seismic profile Line 2 (Levanto-Felino) a) uninterpreted in TWT (s) and with velocity logs and wells used for time-to-depth-conversion (calculated depth on the right of Fig. 6b); the velocity logs (a) are derived from stacking velocities of 5 CDP (after Camurri et al., 2001). b) Interpreted and modified seismic reflection profile after Camurri et al. (2001), Marroni et al. (2010) and Eva et al. (2025).

(Fig. 3b). The absence of magnetic anomalies characteristic of oceanic crust (symmetrical to the spreading center) suggests that the crust of the Ligurian Sea is an extended continental crust (Chamot-Rooke et al., 1999) or "atypical oceanic domain" (Rollet et al., 2002) more similar to transitional type of the crust, formed beneath small back-arc basins with unusual thick sedimentary layer (Menard, 1967; Izquierdo-Llavall et al., 2022; Filina et al., 2022; Yegorova et al., 2010, 2013) or in continent-ocean transition zone (COT) (Contrucci et al., 2001).

The surface heat flow map (Fig. 3c), compiled from the results of Heat Flow Data Base (Fuchs et al., 2021), represents the data release 2021 from 14,003 publications containing the global heat flow data generated between 1939 and 2021, obtained in equal proportions on the onshore and offshore domains. The data extracted from the database for the study region of our transect were filtered to obtain the smooth regional surface heat flow distribution. The map shows a large anomaly of >250 mW/m^2 in the central part of our study area (Fig. 3c). This N-S elongated anomaly covers the region of the northern Apennines in Tuscany and extends further southwards offshore into the Tyrrhenian

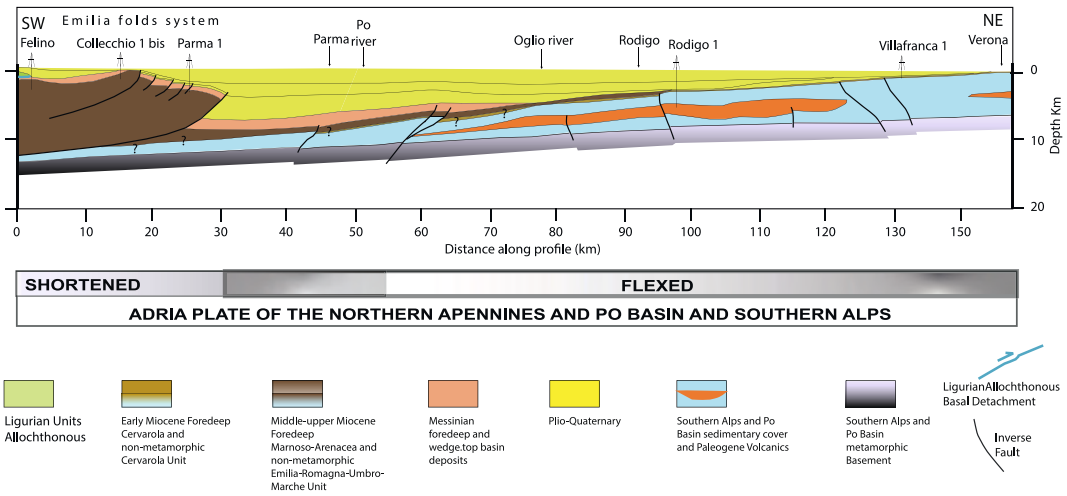


Fig. 8. Interpreted Line 3 Parma-Verona modified after Cassano et al. (1986).

Table 1
Acquisition and processing parameters of the Line 1, Line 3.

Seismic dataset	Acquisition parameters	Main processing Sequence
Line 1 (CrOP M 11)	Recording date November 1993 Energy Source: 4906 cu inch air guns Streamer: 4500 m Shot interval 50 m Group interval 25 m Group 120 Depth 8 m Coverage 30-fold Coverage: 3600% Recording length 1700 ms Filters High cut 77 Hz Sample rate 4 ms	Resample 8 ms Channel reduction To 90 F-K Filter 2–12 ms Analysis interval 15 km
Line 2	Recording date 1990 Energy Source: Dynamite Stacking multiplicity: 600% to 1200% Number of channels recorded 24 Spread configuration: number of groups 144 distance between group 30 m Minimum offset distance: 28,75 m Maximum offset distance: 2145 m	Static correction: reduced to 500 m DP Time variant prefiltering Predictive Deconvolution: GAP 12 ms (200 ms operator - 3 windows) NMO Correction Automatic medium-short wavelenght residual statics (Surface consistent) Stack 2400% fold coverage Migration in time Random noise attenuation Time-Variant Filtering Trace equalization
Line 3 (Seismic Line B in Pieri, 1983)	Recording date 1990 Energy Source: Dynamite Stacking multiplicity: 600% to 1200% Number of channels recorded 24 or 48 Interval between input channels: 75 or 50 m (interpolation from 75 to 50 m) Minimum offset distance: 50 m Maximum offset distance: 2325 m	Static correction: reduced to sea level plane Deconvolution: DBS – MIN. Phase Inv. Filter (32-200 ms operator - 2 windows) Frequency filtering: Variable within 10–50 Hz Automatic Statics: Surface consistent Migration: Finite difference Source of velocity for Processing: Seismic velocity analysis and constant velocity stack

Sea. This anomalous domain of very high heat flow corresponds to the Tuscan Magmatic Province, the igneous activity of which resumed from latest Miocene until present days (Peccerillo, 2005). From ~8.5 Ma to 0.2 Ma a discrete igneous activity generated series of volcanos and plutonic stocks mostly in the Tuscan region and offshore northern Tyrrhenian Sea. These rocks are represented both by crust- and mantle-derived melts as well as liquids of pure mantle and hybrids (crust + mantle) origin (Serri et al., 1993; Carminati et al., 2010; Brogi et al., 2025).

Another area of high heat flow, that includes two northeastwards elongated anomalies of up to 120 mW/m^2, is recorded in the central part of the Ligurian Sea where thin extended continental crust ("atypical oceanic crust") occurs (Fig. 3c). The northeastern anomaly is crossed by our new lithospheric transect.

4.2. *The crustal structure*

4.2.1. *Moho boundary*

The Moho boundary was constrained by European Moho compilation and Moho depths maps for Italy (Grad et al., 2009; Cassinis et al., 2005; Di Stefano et al., 2011; Spada et al., 2013). The Moho boundary is an important constraint for density/gravity modelling since it is the first-order seismic boundary with a velocity contrast, and consequently, with a significant density contrast. This is of particular importance in density/gravity modelling of transition zones from thin oceanic/sub-oceanic crust to thick continental crust, as it is the case of our transect that crosses the Northern Apennines where the Moho varies in the depth range of 15–17 km (Ligurian Sea) to 40 km (Po Plain).

A number of models of the Moho depths are known for the study region. These are the Moho maps of Italy, constrained by seismic data from active and passive experiments (Scarascia et al., 1994; Cassinis et al., 2005; Di Stefano et al., 2011; Spada et al., 2013; Molli et al., 2010) and by the results of 3D gravity modelling taking into account seismic data (Chamot-Rooke et al., 1999). Some of these models identified Moho domains as characteristic of the continental crust of the European and Adriatic plates, and of the sub-oceanic crust of the Ligurian-Tyrrhenian Basin (Molli et al., 2010; Di Stefano et al., 2011; Spada et al., 2013). Apart from these regional models, the Moho model of Europe (Dèzes and Ziegler, 2002) is in great demand. A digital high-resolution map of the Moho depths for the whole European Plate (Grad et al., 2009) is of particular interest for our study. It is compiled from more than 250 data sets of individual seismic profiles, including 3-D models obtained by active and passive seismic experiments. This last compilation was used for approximating the Moho interface for our gravity modelling along the study transect (Fig. 3d).

The Moho map in Fig. 3d clearly shows two distinct crustal blocks: the Adriatic plate with a Moho depth of 28–36 km; a thin crust block of the Ligurian Sea, striking SW-NE, where the crustal thickness reaches 16–18 km depth in the central part of the deep marine basin. The NE

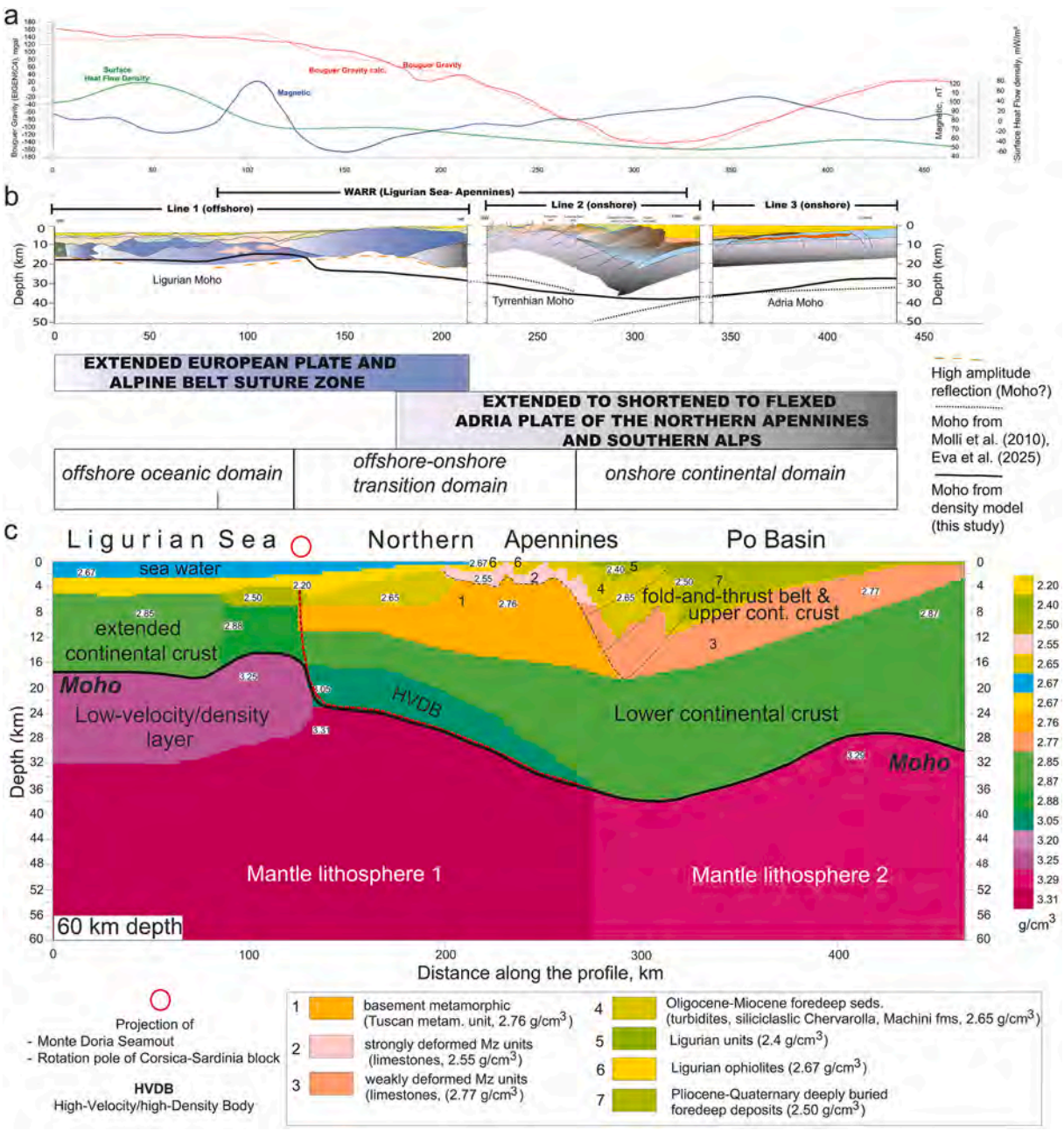


Fig. 9. a) The observed and calculated gravity (red solid and dashed lines respectively) supplemented with magnetic field and surface heat flow plotted along (b) the composite Ligurian Sea-Po Basin transect consisting of three profiles shown with the legend in Figs. 6–8 and (c) the gravity modelling for the crust and upper mantle at depth interval 0–60 km (numbers denote the density values in g/cm^3 obtained with the modelling; numbers 1–7 indicate the constituent parts of the accretionary prism and basement of the Northern Apennines). For the velocities values see Fig. 5. The gravity modelling was carried out with the GRAV3D software (Tchernychev and Makris, 1996). The velocities and densities of main complexes and layers are in Table 2. The new integrated gravity model shows the high-velocity/high-density lower crust body HVDB (dark green infill). Red dashed line indicates the transform fault in the crust of the continental slope offshore-onshore transition domain. The transform fault continues as a mantle (Moho related) shear zone (see chapter 7- Discussion for details). The pole of rotation (red circle) of the Corsica-Sardinia block is close and projected on the transect. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

portion of the Moho below the Ligurian Sea adjoins the thinned crust of the northern Tyrrhenian Sea with the Moho depths deepening to 24–22 km together with its eastern projection which continues onshore into the Tuscany region (Fig. 3d). The Adriatic Plate includes the Northern Apennines and the Po Basin with a continental crust of increased thickness up to 28 km and 37 km respectively.

4.2.2. Wide-angle reflection and refraction studies

Wide-angle reflection and refraction offshore and onshore seismic experiment, fulfilled in the Ligurian Sea to the Northern Apennines (Makris et al., 1999), provides good constraints for crustal structure from the thin crust of the Ligurian Sea to the thick continental crust of the Northern Apennines and the transition from one to the other. This experiment allows also to estimate the P-wave velocity for the sedimentary succession, crystalline crust and uppermost mantle (Fig. 4). The continental crust beneath the Northern Apennines has thickness of ~40 km including the gradually increasing thickness of sediments up to 10 km. The crystalline portion of the continental crust thins up to 10 km across a ~ 150 km-wide zone extending from the SW portion of the Apennines to the Ligurian Sea. The crystalline crust below the Ligurian

Sea has thickness of ~7 km overlain by two-layered sedimentary succession approximately 5 km thick. Mantle velocities in the region range from 7.9 km/s beneath thin crust of the Ligurian Sea to 8.3–8.4 km/s and beneath the continental crust of Northern Apennines (Makris et al., 1999; Paul et al., 2024; Götze et al., 2024). According to Makris et al. (1999), below the crustal portion, a peculiar feature is a wedge, 7–8 km to 0 km thick and velocities of 7.2–7.4 km/s, dipping to the NE below the Northern Apennines (Fig. 4). Remarkably, this wedge shows low velocity compared to mantle velocities and high velocity compared to crustal velocities; it is an anomalous body in the studied lithospheric transect. Following Makris et al. (1999), this is a sub-Moho/sub-crustal body. Thus, it should belong to the upper lithospheric mantle. Nonetheless, the highest velocity/density contrast (i.e., Moho) occurs at the base of the wedge (Fig. 5 in Makris et al., 1999) posing it toward the base of the crust or lower crust. In this work, we name this wedge High Velocity/high-Density Body (HVDB) and we consider it an anomalous lower crustal body. It is of particular interest, because it arises questions about its interpretation and role in the frame of the Alpine-Apenninic orogenic system (see chapter 7. Discussion).

4.3. *Upper mantle structure derived from surface wave tomography*

The upper mantle structure, useful to constrain the upper mantle heterogeneity and the lithosphere-asthenosphere system below the transect, was constrained by the surface wave tomography model of Italy (Brandmayr et al., 2010) represented by 1° × 1° grid cells and supplemented by a set of 1° cellular columns (tables) with the parameters of Vs (shear wave velocities) and depth down to 300–400 km depth. The values and cells of Brandmayr et al. (2010) have been extracted along the study transect by an assemblage of corresponding cellular columns down to 150–180 km depth (Fig. 5). In more detail, its description and interpretation are given further in the context of discussing the crust and upper mantle structure of the transect (see Subchapter 4.5.4. Lithosphere-asthenosphere system).

The model used could be further converted into density structure. The obtained upper mantle structure has been also compared to previous studies based on different modifications of seismic tomography and using different types of waves (Panza et al., 1992; Panza et al., 2007; Piromallo and Morelli, 2003; Scafidi et al., 2009; Di Stefano et al., 2011; Solarino et al., 2018, 2024; Nouibat et al., 2022, 2023; Fernàndez et al., 2019).

5. Geological-geophysical data compilation: the new Ligurian sea-Po Basin lithospheric transect

To constrain the geometry of the major tectono-sedimentary units, the geometries of sedimentary successions and basement along the transect from the Ligurian Sea to the Po Basin, we used three seismic reflection profiles derived from published (Argnani et al., 2003; Cassano et al., 1986; Bernini et al., 1997) and unpublished works (Camurri, 2000; Camurri et al., 2001; Rizzini et al., 2006; Rizzini, 2007; Torelli, 1994) (Fig. 2). They constitute three sub-sections which illuminate the upper crustal level of the transect: 1) the SW offshore part describes the Ligurian Sea-Levanto cross-section (Line 1 in Figs. 2, 6), 2) the Levanto-Felino profile (Line 2 in Fig. 2) imaging the central onshore segment (Fig. 7), and 3) the Parma-Verona profile (Line 3 in Fig. 2) inside the onshore Po Basin (Fig. 8). The three seismic reflection profiles are projected and merged onto the SW-NE directed modelled transect and they create a composite cross section (Fig. 9) which provides constraints on the structure of the upper (down to the depths of ~15–20 km) crustal portions as derived from the results of time-to-depth conversion which has been fulfilled both in the present (Fig. 6) and previous (Cassano et al., 1986; Camurri et al., 2001; Figs. 7, 8) works.

5.1. *Line 1 - Ligurian Sea-Levanto cross-section*

The offshore part of the transect (Fig. 6) starts in the deep-water portion of the Ligurian Sea (Bartole et al., 1991; Bernini et al., 1997; Torelli, 2001; Rizzini, 2007; Rizzini et al., 2006) and it covers the oceanic domain of the transect. The Late Eocene(?)/Oligocene – Early Miocene *syn*-rift sequences, the well-stratified post-rift undifferentiated Messinian deposits, with evidence of deformation due to salt remobilization (diapirism – E in Fig. 6) and gravitational tectonics, are well recognized here (Rizzini et al., 2006) (Fig. 6).

In the central part of Line 1, around the projection of Monte Doria seamount, there is a block devoid of reflections interpreted as a crustal block covered by thin sediments (km 70–90 in Fig. 6). This block separates the deep-sea oceanic domain with thick sedimentary cover, above possible oceanic crust or exhumed mantle (Rollet et al., 2002; Dannowski et al., 2020; Dessa et al., 2020; Jolivet et al., 2021), of the Ligurian Sea from the domain with thinned continental crust, likely representing the transition from thin crust with extended/rifted continental crust (km 0–70 Fig. 6) to the continental crust of Northern Apennines (Figs. 6, 7, 9). This extended/rifted continental crust has to be superposed to the prosecution of the Corsica-Sardinia block with the exhumed alpine units (Alpine Corsica) which reach the Voltri group toward N-NW (Fig. 1). Thus, the extended/rifted continental crust is affecting the suture zone of alpine chain or early Alpine-Apenninic orogenic system.

The northeast end of Line 1 (km 90–150 Fig. 6) reaches the area that is located on the northward continuation of the Northern Tyrrhenian Sea and Tuscany region which, since Upper Miocene-Lower Pliocene, experienced extensional tectonics demonstrated by west-dipping low-angle listric faults, as well as antithetic high-angle normal fault systems bounding syn-rift fault-controlled basins (Figs. 6, 7, 9; Rizzini et al., 2006; Molli et al., 2018). One of these east-dipping syn-rift faults is around km 120 of the Line 1 (Fig. 6).

5.2. *Line 2 – Levanto - Felino cross-section*

In the Levanto-Felino section (Fig. 7), a surface has been mapped along the Northern Apennines indicating the top of the basement units (Camurri, 2000; Carlini et al., 2013; Molli et al., 2018). This surface dips mainly toward east, starts at 2–3 km depth near Levanto (km 0–10 in Fig. 7), is crossed by Pontremoli 1 well (Anelli et al., 1994), reaches 18 km depth below the Northern Apennines fold and thrust belt and outcrops in the Alpi Apuane (Fig. 1). In the area where the basement occurs at shallower depths, the high-angle Pliocene-recent extensional faults outlines two grabens - Val di Vara and Val di Magra forming the Lunigiana Fault system (km 10–40 in Fig. 7, Fig. 1). The Val di Magra basin, bounded by a system of antithetic master faults (Mulazzo and Compione faults, Lucca et al., 2018; Molli et al., 2021), has been investigated since long time (Bernini, 1988, 1991; Bernini and Lasagna, 1988; Bernini et al., 1997; Bernini and Papani, 2002). It dips southwestwards, has an offset of ~1 km; it is the most northeastern normal fault recognized to dismember the whole stack of tectonic units forming the Apennine fold and thrust belt (Fig. 7). These normal faults cross cut the about 3 km thick sedimentary cover whose thickness is the result of pre-Pliocene shortening and low-angle extension (Carlini et al., 2013; Molli et al., 2018).

The not-extended Northern Apennine fold and thrust belt (km 40–85 in Fig. 7) starts NE of the Lunigiana Fault systems and includes the watershed of the Apennines. It is characterized by the highest allochthonous Ligurian nappes which overlay and are out-of-sequence cross-cutted from beneath by three major tectonic units/thrust sheets: the Tuscan unit, the Cervarola unit, and the Emilia-Romagna-Umbro-Marche-Padan unit. These three tectonic units approximately correspond and control the Oligocene-Miocene foredeep basins which migrated toward NE in response to the advancing orogenic wedge fronts of the Apennines (Boccaletti et al., 1990; Argnani and Ricci Lucchi,

2001). The youngest post-Late Miocene and Pliocene-Quaternary foredeep basin of the Northern Apennines corresponds to the Po Basin and it is above the flexed Adria Plate (km 85–100 in Fig. 7).

5.3. Line 3 – Parma - Verona cross-section

The outermost and northeastward thrust sheet deforms the Middle Miocene foredeep basin and defines the inner front/s of the Late Miocene-recent Po Plain foredeep basin. This thrust zone comprises a narrow belt of thrust-related-folds, called pede-Apennine thrust front or Emilia Folds system (Emilia Folded Arc in Pieri and Groppi, 1981), marking the boundary between the Apennines foothills and the Po Basin (km 0–30 in Fig. 8). The Emilia folds system corresponds to a steep NE deepening of the basement which reaches the depth of ~20 km (Fig. 8). Strikingly, the pede-Apennine thrust front (Emilia folds system) appears to cut down across the basement, which is involved in the foot-hills structures (Figs. 7, 8). The deep, thick-skinned thrusts prosecute northeastward and reach higher decollement levels inside the sedimentary successions that led to the formation of the outermost and buried Northern Apennines thrust system beneath the Po Basin (Pieri and Groppi, 1981; Livani et al., 2018) (Fig. 1). Two decollement levels are recognized here: the lowest basal decollement located in Triassic evaporites (Carnian), while the highest upper detachment level within the Oligocene-Middle Miocene clay and marls that belong to the distal portions of the Oligocene-Miocene foredeep (Livani et al., 2018). Even though, basement involved frontal structures and inverted Mesozoic faults are also envisaged for the buried Po Plain thrust system (Franciosi and Fantoni, 2010; Maesano et al., 2015; Turrini et al., 2016; Livani et al., 2018, 2023; Amadori et al., 2019) (e.g., around km 60 in Fig. 8), especially in the Ferrara Arc (Pieri and Groppi, 1981) (Fig. 1). The load of the fold and thrust belt and the form of basal decollement have determined the dip of the monocline, which shapes the basement of the Po Basin and overlaying Mesozoic-Cenozoic sediments (km 30–160 in Fig. 8); the Cenozoic deposits fill the space created by the flexed basement and they thin from 15 km to 5 km toward the NE (Figs. 8, 9; Pieri and Groppi, 1981; Cassano et al., 1986).

A remarkable feature of the sedimentary fill of the Po Basin is an extensive volcanic body as thick as 2 km, conformable with and locally intruding the SW dipping monocline (Fig. 8). This body is represented by Paleocene-Oligocene volcanics coeval to those cropping out in the Veneto region (lavas, rhyolites, dacites and tuffites); they cause the wide magnetic anomaly along the Liguria Sea-Po Basin profile (Figs. 8, 9; Cassano et al., 1986). The monocline is the SW prosecution of the Southern Alps forming the retrowedge of the Alps.

5.4. Composite cross-section along the new Ligurian Sea-Po Basin lithospheric transect

Over a length of 465 km, the SW-NE composite cross section (Fig. 9) results from the projection of the three seismic reflection profiles described before (Fig. 6–8) onto the trace of the modelled lithospheric transect Ligurian Sea-Po Basin (Fig. 9). Following the compiled geological data (Fig. 1), the composite cross-section crosses: 1) the offshore oceanic domain with extended continental crust of the Alpine chain in the Ligurian Sea and part of the Northern Tyrrhenian Sea; 2) the offshore-onshore domain with complex stack of Apenninic thrust sheets which involve and superpose the metamorphic basement (Tuscan metamorphic unit and external basement sheets), the Mesozoic carbonate rocks, the partly exhumed and partly buried Oligocene-Miocene (pre-late Messinian) foredeep depocenters, the Ligurian allochthonous nappes; 3) the late Messinian to recent foredeep deposits of the Po Plain, which overlay the basement and the Mesozoic-Late Miocene sedimentary succession above; 4) the foreland area of the Northern Apennines in the Verona area which coincides with the Southern Alps foothills and retrowedge of the Alps.

6. Gravity/density model for the crust and upper mantle

All the above geological and geophysical data are needed to constrain the reliable density model for the crust and upper mantle along the transect Ligurian Sea-Po Basin. The main objective of our gravity modelling is constructing the density model from all available seismic profiles by converting them into density equivalent to derive the starting model, which is calculated by a trial-and-error method to fit the calculated and observed gravity fields.

For our modelling we used the GRAV3D software developed by Tchernychev and Makris (1996), which is based on Talwani et al. (1959) algorithm and uses an approximation of the model by prisms of constant density with the dimensions of 2.5 × 0.5 km (X, Y) in our case. This software, in 2D and 3D, have been applied for different areas (Makris and Yegorova, 2006; Yegorova et al., 2011, 2013; Makris et al., 2013).

The geometry of the initial density model is constrained by the seismic reflection profiles Line 1, Line 2, and Line 3 (Figs. 6–8) and the structure of the crystalline crust - by the deep wide-angle seismic profile of Makris et al. (1999). The depths to the Moho were taken from (Makris et al., 1999), taking into account the European Moho compilation (Grad et al., 2009) and Moho depths in Italy (Cassinis et al., 2005; Di Stefano et al., 2011; Spada et al., 2013). For the upper mantle, the S-wave seismic tomography model of Brandmayr et al. (2010) was used. For density parametrization of the initial model, the velocity values of seismic models were converted into densities using the conversion function of Nafe and Drake (Ludwig et al., 1970) and generalization of Brocher (2005) (Table 2). During the modelling, the geometry of the layers was fixed and further refinement of the model was achieved interactively mainly in terms of its densities.

A stable solution of the modelling (Fig. 9) was obtained for a deep-water oceanic domain with extended continental crust which passes to the thicker continental crust toward NE. The crustal thicknesses changes from a 14–18 km in deep-water oceanic domain of the Ligurian Sea, including the Monte Doria Seamount, to ≤40 km-thick onshore

Table 2
Velocity and density parameters of the Ligurian Sea-Po Basin cross-section.

Layer/Rocks	Vp, km/s (according to Makris et al., 1999)	Density, g/cm^3, from Ludwig et al., 1970; Brocher, 2005)	Density adopted for the final density model
Marine sediments I	1.8–4.2	1.8–2.4	2.20
Marine sediments II	4.5–5.5	2.45–2.66	2.5
Sediments (continental)	2.8–4.2	2.15–2.42	2.4
Deeply buried foredeep deposits (Po basin) - Pliocene-Quaternary	4.2–4.57-5.5	2.42–2.62	2.50
Foredeep deposits (turbidites, siliciclastic) Macigno, Cervarola (Oligocene-Miocene)	4.5–5.5	2.45–2.66	2.65
Strongly deformed carbonates (limestones) - Mesozoic	4.5–5.0	2.45–2.53	2.55
Weakly deformed carbonates (limestones) - Oligocene-Miocene	6.2–6.5	2.75–2.81	2.77
Ligurian units (no ophiolites)	2.8–4.2?	2.15–2.42	2.4
Ligurian ophiolites			2.67
Basement metamorphic	6.2–6.5	2.75–2.81	2.75
Continental lower crust	6.6–6.8	2.85–2.89	2.87
Oceanic crust	6.4–6.6	2.80–2.85	2.85
High-velocity/high-density body (HVDB)	7.2–7.4	2.95–3.08	3.05
Upper asthenopheric domain	8.0	3.24	3.25
Upper mantle transition	8.1	3.31	3.31
Upper mantle continental	8.2–8.4	3.3–3.4	3.29

continental domain which includes up to 18 km thick meta-sedimentary and sedimentary succession in the Po Basin. In between, from the continental slope of the Ligurian Sea to the watershed of northwestern Apennines, the offshore-onshore transition domain, ~150-km wide and crustal thickness which increases toward the NE from 14 km to 30 km. This is a wide transition zone that includes most of the offshore and onshore Northern Apennines' complex stack that is also thinned.

The obtained gravity model fits also the lithosphere model down to the depth of 155 km, providing constraints on the variations of the lithosphere-asthenosphere boundary (LAB) along the transect (Fig. 10).

Based on the characteristics of the lithosphere, three main domains can be distinguished in the Ligurian Sea-Po Basin transect (Fig. 9). We first describe the modelling results of the intermediate depth lithospheric model (60 km depth, Figs. 9, 10), starting from the offshore oceanic domain to the offshore-onshore transition domain and, finally, to the onshore continental domain of the new lithospheric transect.

6.1. Offshore Oceanic domain

In the offshore part of the Ligurian Sea-Northern Apennine-Po Basin

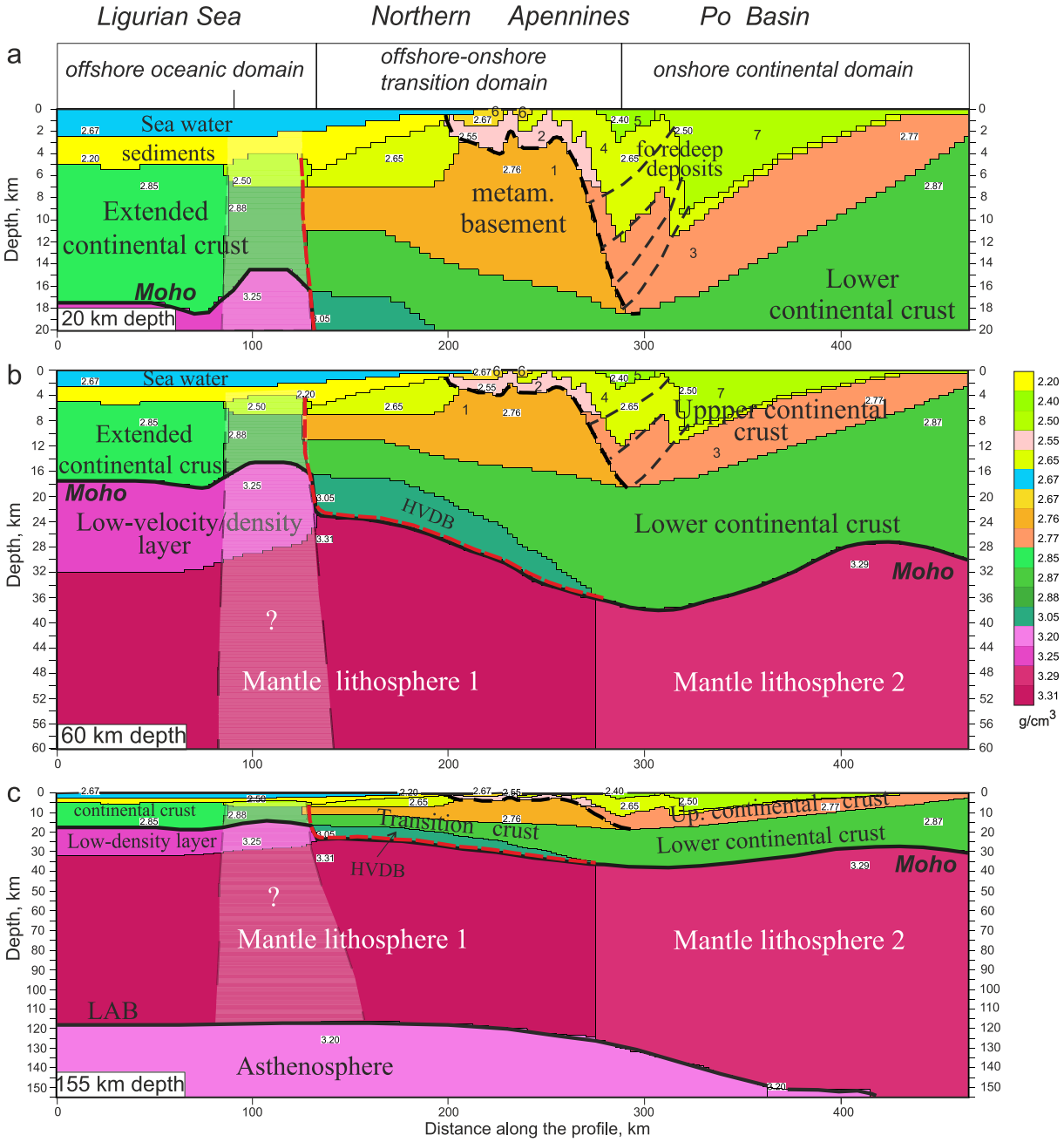


Fig. 10. Calculated gravity model for the crust and upper mantle of the Ligurian Sea-Po Basin transect shown at different depths: 20 km (a), 60 km (b) and 155 km (c). Numbers denote the density values in g/cm^3. The asthenosphere, or lower low-velocity layer, is considered to have S-velocities of 4.0–4.40 km/s. For the other velocities values see Fig. 5. Numbers 1–7 indicate the constituent parts of the fold-and-thrust belt and basement of the Northern Apennines (see also Fig. 9 and Table 2). HVDB is the High-Velocity/High-Density Body. Red dashed line indicates the transform fault in the crust and its possible prosecution as a mantle (Moho related) shear zone (for details see chapter 7. Discussion). The transform fault marks the boundary between the deep-water offshore oceanic segment and the offshore-onshore transition domain. The boundary between the two domains is also marked by the lithospheric block (whitened area) with thinner crust and a low-velocity /density layer in the mantle lithosphere. Vertical exaggerations are ~4.5:1 and ~ 3.05:1 for the upper and middle sections (a and b), and ~ 1:1 for the whole model (c). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

transect (Fig. 9), three segments are distinguished: the deep-water portion (km 0–85), which is separated by a narrow (40 km wide) block around the Monte Doria seamount from the continental slope (km 125–200) (Fig. 9). The first two segments form the offshore oceanic domain whereas the third segment is part of the offshore-onshore transition domain.

In the deep-water part of the profile, under ~2.5 km thick sea-water layer, the seismic image distinguishes a ~ 1-km thick layer of uncompacted sediments with linear horizontal layering underlain by a 1–1.5 km to 2 km thick layer of strong horizontal reflectivity, which deepens to a depth of ~7 km in the central part of the marine segment of the profile. They could be represented by post-rift Messinian deposits and *syn*-rift sequences of late Eocene(?)/Oligocene – early Miocene age (Bartole et al., 1991; Torelli, 1994; Torelli, 2001 – Unpublished; Rizzini et al., 2006) (Figs. 6, 9a). The density of these two layers was taken as 2.2 g/cm^3 as characterizing uncompacted marine sediments with velocity of 1.8–4.2 km/s (Table 2, Fig. 9). These two upper sedimentary layers were deformed by salt diapirism and gravitational tectonics manifested themselves in large number of Messinian salt diapirs (two of them marked in Fig. 6).

Deeper, the basement is distinguished as a fairly homogeneous area with a small number of reflections. The top of the basement and the Moho boundary are not reliably distinguished, most likely due to a weak seismic signal generated by airgun shooting though a high amplitude reflection might be attributed to the Moho discontinuity (Fig. 6). The latter boundary is identified only fragmentarily at the depth of ~13–14 km (7–8 s of TWT) and it locally coincides with the Moho used in our modelling (km 0–70 in Fig. 9). For our gravity model, the depths of the Moho in the offshore oceanic domain were taken at 17.5–18 km depth in accordance with the published Moho maps (Grad et al., 2009; Di Stefano et al., 2011; Spada et al., 2013; Cassinis et al., 2005).

The northeasternmost segment of the offshore oceanic domain, near the projection of Monte Doria seamount, exhibits a small uplift of both the rocks with Vp = 4.5–5.5 km/s (density of 2.50 g/cm^3) and the Moho depths to 14.5 km (Makris et al., 1999). This thinner crust segment (km 85–120 in Fig. 9) marks the ends of the offshore oceanic domain and the abrupt inception of thicker crust beneath the continental slope included into the offshore-onshore transition domain.

6.2. Offshore-onshore transition domain

The crust of the continental slope (~80 km length between km 120–200 offshore) and the crust of the innermost portion (seaward side) of the northern Apennines (~70 km length between km 200–270 onshore) belong to the offshore-onshore transition domain which ends around the Apennines' watershed (Fig. 9). This domain exhibits feature of thinned continental crust. This is the only portion of the model (at depth of 1.5–7 km) where we increased densities of the lower sedimentary successions; seismic velocities were indicated unclear by Makris et al. (1999) and a density of 2.65 g/cm^3 was attributed by assuming them as crystalline rocks. This corresponds well with the reinterpretation of seismic data on Line 1 that shows here the block of exhumed Alpine units of variable metamorphic grade (km 85–150 in Figs. 6a, 9).

The offshore-onshore transition domain, ~150-km wide and crustal thickness varying in the range from ~20 km to 32 km, includes also most of the Northern Apennines' embricate stack that is here thinned. However, around a depth of ~20 km, it is here to note a body characterized by Vp velocity of 7.2–7.4 km/s and density of 3.05 g/cm^3, thicknesses of ~7–8 km thinning out to 0 km toward the NE and dips toward the Po Basin (Figs. 9, 10). As explained in Subchapter 4.2.2 Wide-angle reflection and refraction studies, the noted body is high-velocity/high-density with respect to crustal velocities/densities ranges and named HVDB. It coincides with the body detected in the WARR of Makris et al. (1999) who considered it a subcrustal/sub-Moho body. However, according to the Moho map here presented (Fig. 3d) and the velocity/density contrast along the transect (Fig. 4), the HVDB should be above the Moho and part of the lower crust as shown in Fig. 9. The offshore-onshore transition domain also corresponds to a wide gradient zone of the gravity field from the gravity low of −160 mGal of the Po Basin to the high amplitude gravity field values (+140 mGal) in the Ligurian Sea (Figs. 9, 10).

6.3. Onshore continental domain

The onshore continental domain occupies the NE half of the composite profile extending from the Northern Apennines's watershed through the Po Basin (km 270–450), where Moho depth changes from 30 km to 38 km depth (Figs. 9, 10) and shows a smoother geometry compared to previous Moho profile plotted on the same composite transect (Molli et al., 2010; Eva et al., 2025). The upper part of the crust comprises the deeply buried (down to 18 km depth) sedimentary and meta-sedimentary foredeep deposits of the Northern Apennines fold-and-thrust belt represented by the complex stack of embricated thrust sheets with P-wave velocities and modelled densities reported in Table 2 and Fig. 9. The lower continental crust, with P-wave velocity of 6.6–6.8 km/s and density of 2.87 g/cm^3, is characterized by its downward bending from the thinned portion (~ 10 km thickness) below the Northen Apennines to the much thicker lower crust (~20 km) below the Po Basin (Fig. 9). In this way, the complex stack of the foredeep complexes that occupies much of the upper continental crust along the transect conforms the structure of the lower crust and of the Moho geometry which, together, cause the wide gravity low of −160 mGal (Figs. 9, 10).

6.4. Lithosphere-asthenosphere system

The densities in the upper mantle, estimated by gravity inversion procedure and constrained by surface-wave tomography (Brandmayr et al., 2010) (Fig. 5), show two major blocks of mantle lithosphere: one below the offshore oceanic and the offshore-onshore transition domains with estimated density values of 3.31 g/cm^3; one beneath the onshore continental crust domain with estimated density values of 3.29 g/cm^3. They could explain in part the overall NE-decreasing trend of the gravity field along the transect (Figs. 9 a, c and 10).

The S-wave velocity section, constructed along the transect, reveals also two low-velocity layers in the upper mantle (Fig. 5; see also Fig. 5 in Brandmayr et al., 2010): a thin (~15–20 km thick) sub-Moho layer with Vs = 4.0–4.25 km/s below the Ligurian Sea and corresponding to the offshore oceanic domain, and a deeper thick layer (Vs = 4.0–4.40 km/s) at the depths of ≥120 km below the major part of the transect (Fig. 10c). These velocity heterogeneities were integrated into the upper mantle of our density model with estimated densities of 3.25 g/cm^3 and 3.20 g/cm^3 (Fig. 10b, c). The deepest low-velocity/density layer is associated with the asthenosphere. Instead, the shallower thin subcrustal layer in the offshore oceanic domain with similar density/velocity characters is here related to the soft and heated upper mantle material which explains the high surface heat flow anomaly that has value of 110 mW/m^2 (Figs. 3c, 9). This anomaly is indicative of a heating event activated either recently or in unknown past, active till present-days and affecting the uppermost mantle in the NE segment of the Ligurian Sea. Considering the heat flow map (Fig. 3c), we suggest that this uppermost heated mantle zone prosecutes to the SW joining the soft upper mantle material and drives the subsidence processes that occur in the central deep-water portion of the Ligurian Sea, at least up to 42° latitude, where, unfortunately, seismic tomography data collected for this work are lacking (Brandmayr et al., 2010).

7. Discussion

The new density model along the composite transect Ligurian Sea-Po Basin shows the architecture of the sedimentary basins, structure of the

crystalline crust and upper mantle down to 150 km depth from the thin crust offshore to a thick continental crust of the Northern Apennines and Po Basin passing through a transition zone, which contains extended continental crust offshore, part of continental crust onshore and the embricated stack of thrust sheets made of meta-sedimentary and sedimentary successions reaching 18 km in thickness; this latter portion being in part thinned and dismembered by extensional events (Figs. 9, 10).

This new integrated gravity/density model allows revisiting the lithospheric architecture along the offshore-onshore Ligurian Sea-Po Basin transect and shed new lights on the deep configuration of the Northern Apennines – Western Alps junction area or Ligurian knot (Laubscher et al., 1992). The latter is the concentration of several loose ends of important tectonic zones of movements between the southern Alps, the western Alps, the Ligurian Alps, the Northern Apennines, and the Ligurian and Tyrrhenian seas (Laubscher et al., 1992) and it is the result of the complex interference of various tectonic processes related to collision of Europe and Adria with the formation of opposite dipping subductions, subduction shift and development of the Northern Apennines in a frame of slab retreat and opening of the back-arc Ligure-Provençal and Tyrrhenian basins (Laubscher et al., 1992; Molli et al., 2010; Handy et al., 2010; Argnani, 2012; Carminati and Doglioni, 2012; Schmid et al., 2017; Van Hinsbergen et al., 2020; Jolivet et al., 2021). The new transect reflects the long-lived tectonic memory of the northern sector of Northern Apennines near its junction with the Western Alps and facing the Southern Alps. The tectonic memory could preserve the imprints of the events occurred from Permian-Mesozoic separation of Europe from Africa to recent eastward propagation of the Northern Apennines foredeep basin associated with ongoing seismicity.

In the following, we attempt to highlight and discuss this tectonic memory by tracing back the different tectonic events occurred from most recent extensional ones (Fig. 11a, b) to those occurred during collision (Fig. 11c, d). We consider that memory of some of these events is testified by the HVDB highlighted in the revisited lithospheric transect (Fig. 9).

7.1. Opening of the Ligurian Sea and nature of its crust

The opening of the Ligure-Provençal and Tyrrhenian Basins is assumed to be related to the Oligo-Miocene geodynamic due to the northwestwards movement of the Adriatic Indenter (i.e. Adria plate), the counterclockwise rotation of the Corsica-Sardinia block (i.e. Europe plate and Adria-Europe suture zone) (Crispini et al., 2009; Handy et al., 2010; Van Hinsbergen et al., 2020; Crouzet et al., 2025; and references therein) and the slab roll-back of the Adria triggering back-arc basin opening (Réhault et al., 1984; Mauffret et al., 1995; Gueguen et al., 1998; Faccenna et al., 2001; Carminati and Doglioni, 2012; Maillard et al., 2020; Jolivet et al., 2021) (Fig. 11a, b). Around the rotation pole, located in the Ligurian Sea north of Corsica (43.5°N, 9.0°E) (Réhault et al., 1984; Speranza et al., 2002; Gattacceca et al., 2007) (Figs. 1, 11a, b), the Corsica-Sardinia block rotated counterclockwise in the range of 20–68° (Maffione et al., 2008) with most of the 45° of rotation between 20 and 15 Ma, whereas ~10° remain since 15 Ma (Crouzet et al., 2025; Gattacceca et al., 2007). The Corsica-Sardinia and Northern Apennines can play as rigid blocks (Crouzet et al., 2025) in agreement with the hypothesis of a proto-arcuate belt of Alpine-Apenninic orogenic system inherited from the subduction stage and active at different times in different domains during the Tertiary (Stampfli et al., 2004; Brunsmann et al., 2024; Van de Vyver et al., 2024).

The location of the rotation pole of the Corsica-Sardinia block is very close to our profile; it is near the Monte Doria Seamount where characteristic magnetic anomaly marks the transition from the deep-water offshore oceanic domain of the Ligurian Sea to the continental slope of the offshore-onshore transition domain (Figs. 1, 3b, 9). This implies that the transect has a portion, SW of the pole, which shows the opened Ligure- Provençal back-arc basin crossed at its NE termination (Fig. 11b); while NE of the rotation pole, the transect shows the roto-translation of the Corsica block separated from the Northern Apennine (deformed Adria) block as a consequence of the Late Miocene to Present Day Tyrrhenian extension (Fig. 11a). The motion vectors pointing to the NNW (blue arrow with red contour and red arrow with blue contour in Fig. 11) indicate that Corsica-Sardinia and Northern Apennine blocks were and are separating slightly obliquely with respect to the transect and they dismembered the Alpine-Apennine orogenic system. In particular, the Northern Apennine block dismembers the Oligocene-Miocene foreland basin system of Apennines; whereas Corsica-Sardinia block dismembers the suture zone of orogenic system as testified by exhumed HP-LT remnants close to the Monte Doria Seamount and the Alpine Corsica. The latter is dismembered together with its westerly dipping/subducting European plate and associated Paleogene Alpine pro-wedge foreland basin (Van de Vyveret al., 2024; Di Rosa et al., 2025). These remnants, derived from the dismemberment of previously collided/sutured continental plates, and the earlier (Eocene(?)-Oligocene) foreland basins system are believed to create the extended continental crust in the Ligurian Sea as recognized in previous works (Dannowski et al., 2020; Dessa et al., 2020; Jolivet et al., 2021).

7.1.1. The extended continental crust of the Ligurian Sea

The superposition of volcanic bodies and magnetic anomalies is a specific feature for the Ligurian Sea, where volcanism identified on seismic profiles is generally associated with positive magnetic anomalies (Fig. 3b and Fig. 11 in Rollet et al., 2002). Namely, the presence of ambiguous (non-typical oceanic) magnetic anomalies that show a mosaic discontinuous pattern are too complex to correlate with magnetic isochrones. This was one of the reasons to attribute the crust below the Ligurian Sea to extended continental crust (Rollet et al., 2002; Jolivet et al., 2015; Gailler et al., 2009) analogous of that assumed for the Tyrrhenian Sea where locally subcontinental mantle is exhumed (Mauffret et al., 1995; Chamot-Rooke et al., 1999; Sanfilippo et al., 2025). As a matter of facts, the thin high-velocity crust of the deep-water offshore oceanic domain of the Ligurian Sea cannot be classified as typical oceanic crust formed in spreading zones of the mid-ocean ridges at divergent plate boundaries characterized by magma intrusion and extrusion, which led to formation of a basalt pillow lavas, sheeted dikes and gabbro thin layer, which are accompanied by a specific pattern of magnetic isochrones, parallel to the ocean ridges (Grevemeyer et al., 2018; Arndt, 2023 and references therein).

The crust of this portion of the Ligurian Sea is more consistent with transitional type of crust, formed beneath the small (back-arc) ocean basins like the Aleutian Basin, Gulf of Mexico and Black Sea with unusual thick sedimentary layer (Menard, 1967; Izquierdo-Llavall, 2022; Filina et al., 2022; Yegorova et al., 2010, 2013), and in the continent-ocean transition zone (COT), where it may contain exhumed upper mantle (serpentinised peridotite), or in the rifted volcanic margins of North Atlantic and at the western continental margin of India (Laxmi Basin) (Clift et al., 2020).

This extended continental crust is SW of the km 85 on the transect and it reaches the NW part of the Tyrrhenian Sea according to the magnetic signature (Fig. 3b). It is below a ~ 2.3 km thick sea-water layer and a sedimentary succession as thick as 3.5 km with velocity and density of 1.8–4.2 km/s and 2.2 g/cm^3 respectively. A similar sedimentary thickness is derived by a joint interpretation of magnetic and seismic data (Schettino and Turco, 2006) suggesting the presence of post-rift and passive margin deposits which provide an overall subsidence rate of 0.1 mm/year (3.5 km sediments +2.3 km water accumulated in about 34 Myrs). The sedimentary cover is underlain by crystalline crust with Vp = 6.4–6.6 km/s and ρ = 2.85 g/cm3 (Figs. 9, 10). Comparison of the crustal seismic parameters of the Ligurian Sea with those from all oceanic crustal profiles (White et al., 1992) shows that the igneous crust below the Ligurian Basin has lower velocities (6.4–6.6 km/s) than the crust of average oceanic crust of 7–8 km thickness with velocities increasing from 6.7 km/s at the top to about

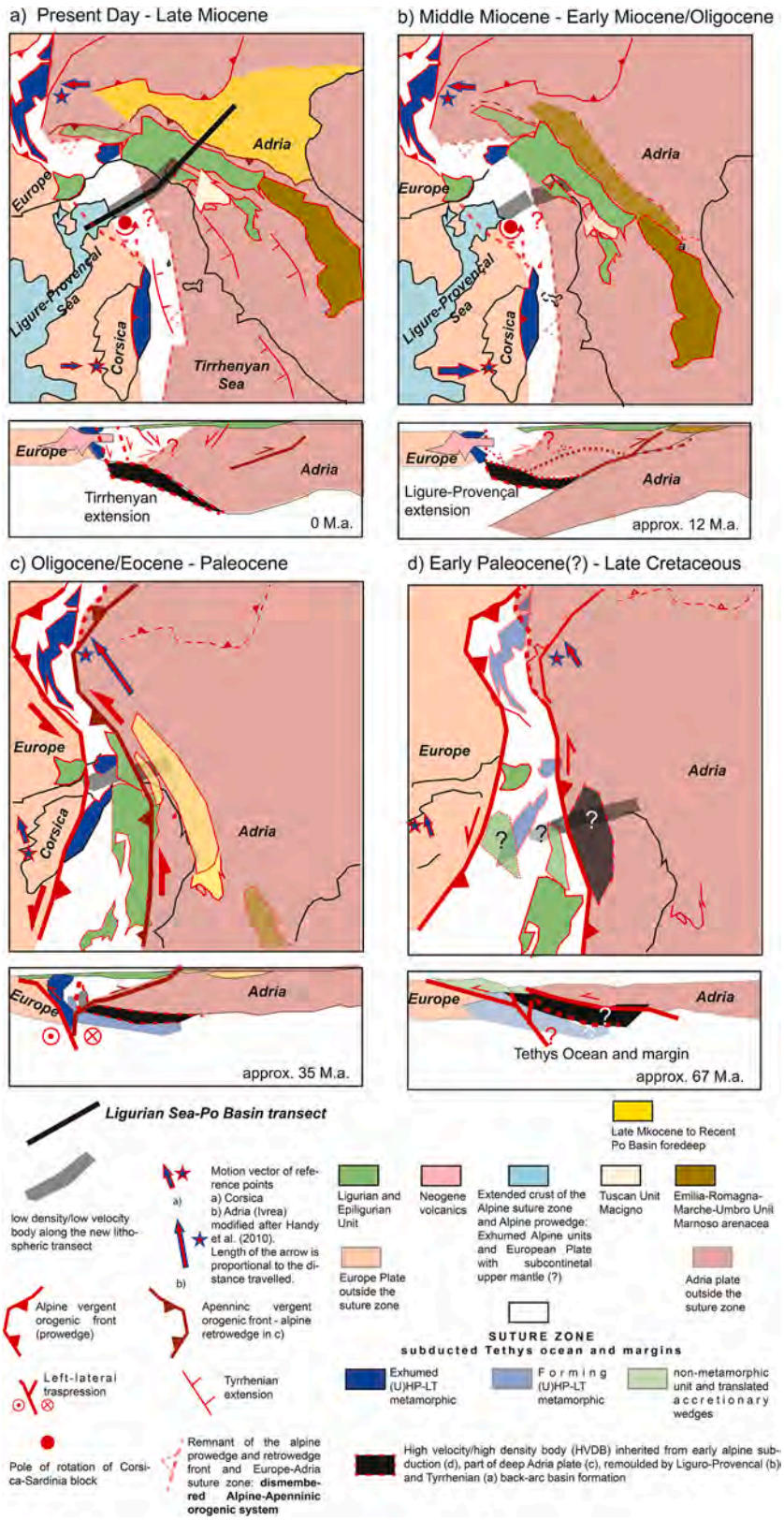


Fig. 11. A paleogeographic reconstruction of the Alpine-Apenninic orogenic system from present day-Late Miocene (a), to Middle Miocene-Early Miocene/Oligocene (b), to Paleocene-Eocene/Oligocene (c) and till Late Cretaceous (d) are here scketched taking into account the previous literature (Stampfli and Borel, 2002; Handy et al., 2010; Le Breton et al., 2017, 2021; van Hinsbergen et al., 2020). Each paleogeographic sketch and associated cross-section below show the proposed position of the High Velocity/high-Density Body (HVDB) (black and pale gray area) evidenced along the new lithospheric cross-section (Fig. 9c). For simplicity, it is drawn with a shape similar to the one in the new gravity model; its geometry, extension and paleogeographic position is still uncertain for b), c) and d). The motion vector arrows are from Handy et al. (2010) (their Fig. 7); the length of the arrow is proportional to the distance travelled by the reference points (Corsica and Ivrea) during the intervals of time in which is depicted.

7.2 km/s at its base. Then, the sub-Moho upper mantle, at ~18 km depth, shows modelled density up to 3.25 g/cm^3 as constrained by seismic velocity of ~8.0 km/s (White et al., 1992). These crustal and upper mantle characters further support the non-oceanic crust present in this portion of the Ligurian Sea.

The sub-Moho mantle region of the lower velocity and density beneath this segment of offshore oceanic domain, as described in subchapter 6.4 (Figs. 9, 10), and the distribution of the high heat flow zone (Fig. 3c) could represent an overheated and ductile (soft) region in the uppermost mantle lithosphere that originated recently by mobilization of the uppermost mantle. That caused the initiation of magmatic activity not only in the Ligurian Sea, but also evident as widespread tectonic and magmatic event in the Toscana domain and Tyrrhenian Sea during last 30 Ma (Lustrino and Wilson, 2007; Lustrino et al., 2011; Lavecchia and Bell, 2012; Brogi et al., 2025) associated with the uprising asthenosphere up to the subcrustal depths (Mongelli et al., 1989; Della Vedova et al., 1991; Panza et al., 2007; Frezzotti et al., 2009).

The formation of the extended continental crust of the Ligurian Sea, and the mobilization and uprise of the overheated subcrustal mantle abruptly ends against a subvertical transform fault zone (Figs. 9, 10). This fault zone marks the NE termination of the extended continental crust of the Ligure-Provençal back-arc basin and it can be considered the ocean-continent transition (OCT) zone of the oceanic crust occurring around 500 km to the SW of the studied Ligurian Sea-Po Basin transect (Rollet et al., 2002; Chamot-Rooke et al., 1999; Jolivet et al., 2021).

It is worth to remark that the abrupt termination of the offshore oceanic domain and, consequently, of the ocean-continent transition type continental crust against the transform fault zone also means that the continental crust and the HVDB in the offshore-onshore transition domain was likely wider than we detect it nowadays; in pre-Oligocene time, the HVDB might be extended more toward SW before being dismembered by extensional faults related to the opening of the Ligure-Provençal basin (Fig. 11c, d).

7.2. The extended suture zone of the collided Europe-Adria plates

On the new composite transect, the offshore-onshore transition domain is abruptly separated from the high-velocity offshore oceanic domain by a block ~40 km wide (at km 80–120) with steep flanks and marked by local magnetic anomaly (Figs. 9, 10). Here, the marine sediments as thick as 1.5 km cover the meta-sedimentary or volcanic rocks with Vp = 4.5–5.5 km/s and the crystalline crust of 6.4–6.6 km/s as thick as 3.5 km and ~ 7 km correspondingly, with the modelled densities of 2.50 g/cm^3 and 2.88 g/cm^3, underlain by the Moho that is uplifted here up to 14 km depth (Figs. 9, 10). This block corresponds to the change from the deep-water oceanic basin to the continental slope and it is located close to the Monte Doria seamount and the local, almost linear magnetic anomaly striking NW-SE (Figs. 3b, 9 and 10). This anomaly is also parallel to the transform faults (Rollet et al., 2002; Argnani, 2012; Jolivet et al., 2021). This block is very close or contains the projection of the rotation pole of the Corsica-Sardinia block (Réhault et al., 1984; Laubscher et al., 1992; Speranza et al., 2002; Gattacceca et al., 2007) (Figs. 1, 3b, 9 and 11). In addition, it should also contain the dismembered Paleogene Alpine pro-wedge foreland basin of Alpine Corsica (Malavieille and Molli, 2024; Van de Vyver et al., 2024; Di Rosa et al., 2025); the boundary between the Alpine Corsica and its prowedge foreland could not be defined offshore but tentatively traced in Fig. 11. Therefore, the offshore oceanic domain and the offshore-onshore transition domain contain: 1) the remnants of the Alps (Alpine Corsica) and exhumed units of HP rocks forming the suture zone of the collided Europe-Adria plate, about SW of rotation pole; 2) the remnants of the Northern Apennine Oligocene-Middle Miocene foreland basin system, about NE of the rotation pole. Both portions forming the Alpine-Apenninic orogenic system.

7.2.1. Remnants of the Alps (Alpine Corsica) and exhumed units of HP rocks of the suture zone

We associate these remnants to clear magnetic anomalies (Fig. 3b) related to block with exhumed HP/LT (or U(HP)/LT) metamorphic rocks. These rocks, represented by ophiolites and serpentinized ultramafic rocks, are exposed on the surface in the South Western Alps, enveloping the Gran Paradiso and Dora Maira massifs, and cropping out in the Voltri Massif in the Ligurian Alps in the junction area from Western Alps to the Northern Apennines (Molli et al., 2010 and references therein) as well as in Alpine Corsica. In the Ligurian Alps, recording *syn-* to late-collisional evolution, they experienced exhumation which occurred in a span time between 50 Ma and 35 Ma (Fig. 11c) (Molli et al., 2010). For the Alps-Apennines junction area (Voltri Massif HP units), Vignaroli et al. (2010) proposed the two-stage exhumation process based on an integrated structural, petrological, geochronological/thermochronological study (U—Pb SHRIMP dating on a zircon rim and titanite, ^{40}Ar—^{39}Ar on phengite, apatite fission tracks). The two stages, evolving from synorogenic to post-orogenic setting, are: 1) the peak metamorphism to eclogitic/blueschist facies conditions with the early synorogenic exhumation (Eocene, 50–35 Ma) occurred within the subduction zone (Fig. 11c, d), and 2) subsequent post-orogenic exhumation starting with the main retrogressive event to greenschists facies conditions (Eocene-Oligocene boundary ~33 Ma) followed by cooling/exhumation events at upper crustal level (Late Oligocene-Miocene) occurred during crustal thinning linked to a regional geodynamic reorganization of the Mediterranean region which was concurrent to the opening of the Ligure-Provençal Basin since Oligocene (post-30 Ma) (Fig. 11a, b).

Therefore, the "exhumed" crustal block (Vp = 6.4–6.6 km/s, ρ = 2.88 g/cm^3) around the area of the Monte Doria seamount/rotation pole could be represented by serpentinized ultramafic rocks (peridotites, ophiolites) similar to those of the Voltri Massif and Alpine Corsica, which experienced pre-Oligocene tectonic unroofing stage corresponding to activation of regional-scale extensional detachment tectonics (Molli et al., 2010). These ultramafic and exhumed rocks are covered by a thin layer (Vp = 4.5–5.5 km/s, ρ = 2.50 g/cm^3) represented by basalts, as evidenced by sampling and submersible dives of the top of the Monte Doria seamount; the lava flows are of tholeiitic composition (basalts of potassic and acid composition) dated 11–12 Ma (Sosson and Guennoc, 1997; Réhault et al., 1998; Guennoc et al., 1995, 1998; Rollet et al., 2002).

7.2.2. Remnants of the Northern Apennine Oligocene-Middle Miocene foreland basin system

Toward the NE, the offshore-onshore transition domain contains also the remnants of the Oligocene-Middle Miocene foreland basin system of the Northern Apennines which are better preserved in the onshore portion (between km 200 and 270 in Figs. 9, 10). At the surface this portion of the transition domain includes most of imbricate stacks of the Northern Apennines' wedge including basement units overlain by Mesozoic carbonate rocks, Oligocene-Miocene foredeep siliciclastic sediments and Ligurian allochthonous units with their Epiligurian succession passively translated on top (Fig. 9). All these tectonic units crosscut by late Miocene to Recent extensional faults which contribute to the thinning of this zone and related to the Tyrrhenian extension.

It is within this offshore-onshore transition domain that occurs the HVDB (P-wave velocity of 7.2–7.4 km/s and density of 3.05 g/cm^3), already detected by Makris et al. (1999) which was described in details in previous chapters. This body arises doubts on its origin and it provides new insights on the evolution of the Alpine-Apenninic orogenic system.

7.3. The origin of the High-Velocity/high-Density Body (HVDB)

The presence of a high-velocity lower crust, referred to a layer at the base of the continental or transitional crust characterized by P-wave velocities significantly higher than average crustal velocities (> 7.0 km/

s, typically up to 7.8 km/s), provides clues about the tectonic history of the region and high-velocity lower crust formation mechanisms. The latter includes the magmatic underplating where mafic or ultramafic melts reach base of the crust, metamorphic densification of the lower crust due to eclogitization, origination of mafic cumulates during crustal differentiation and the terrain accretion processes of the emplacement of oceanic crust or continental collisional fragments during subduction (Eakin et al., 2014; Schulte-Pelkum et al., 2017; Chang et al., 2024).

In diverging plate settings, the occurrence of the layer with Vp >7.2 km/s in the lower crust (at the crustal base) are rather typical for the magma-rich (or volcanic rifted) margins, the majority of passive continental margins (Franke, 2013), which are characterized by thick wedges (up to 15 km) of volcanic flows. Volcanic rifted margins are often associated with large igneous provinces, where magmatism occurs in geologically short time spans of ~2 Ma (Roberts and Bally, 2012). Such volcanic margins are known in the North Atlantic in Norwegian–-Greenland Sea (Voss and Jokat, 2007; White et al., 2008; Mjelde et al., 2005), and in the South Atlantic margins (Dragoi-Stavar and Hall, 2009) where characterize OCT zone.

Instead, the magma-poor margins are characterized by a smaller volume of magma and mechanism of its production, with high-velocity lower crust (Franke, 2013; Minshull, 2009). The best-known examples of magma-poor margins comprise the Iberia-Newfoundland margin (Boillot et al., 1995; Peron-Pinvidic and Manatschal, 2009; Reston, 2007), Brasil-Angola margin (Aslanian et al., 2009; Contrucci et al., 2004), and South-China Sea (Zhou and Yao, 2009; Yang et al., 2018). The important factors for their formation are assumed to be polyphase faulting, depth-dependent stretching resulting in exhumation of mantle rocks, and evolution of detachment faults in combination with the nature of the lower crust (Franke, 2013; Péron-Pinvidic and Manatschal, 2009; Whitmarsh et al., 2001; Reston, 2007).

In our case study, the new transect is located to the north of the majority of known magmatic bodies in Northern Tyrrhenian Sea (Fig. 1) (Bartole et al., 1991; Lustrino et al., 2011; Carminati et al., 2012) and their emplacements ranges between 15.3 and 2.2 Ma (Fig. 4 in Bartole et al., 1991). The Monte Doria Seamount, near the transect and recording volcanic events dated to 11–12 Ma (Guennoc et al., 1995, 1998; Rollet et al., 2002) and partly coeval to the Tyrrhenian magmatic bodies, is here considered more related the presence of the transform fault (Fig. 9). Therefore, the studied lithospheric transect is comparable to a magma-poor passive margin related to the Ligure-Provençal and Tyrrhenian extensions. In this respect, the HVDB can be associated to crustal scale detachment related to the thinning of the lithosphere active since Oligocene.

In subduction-collision setting of the transition zones, the high-velocity lower crust was generally formed by two primary processes – magmatic underplating evident in the Manila trench (Lester et al., 2013; Eakin et al., 2014) and structural underplating. The latter envisages that during accretion-subduction, material from the down going plate is scraped off and underplated to the bottom of the overriding plate's accretionary prism. This process is actively occurring in the southern Taiwan incipient mountain belt where thinned continental crust is being added to the accretionary wedge (Simoes et al., 2007; Chang et al., 2024) and could be occurred in the Western Alps, where the lower crust of the European plate subducts below the accretionary wedge and the overriding Adria lithospheric mantle (Malusà et al., 2021; Solarino et al., 2024; Nouibat et al., 2022). This situation of the Western Alps could be rather similar to Central Alps (Montemagni et al., 2026) and also to Ligurian Sea-Po Basin transect. In fact, the offshore-onshore transition domain shows certain similarities with the recently acquired CIFALPS2 and CIFALPS2-Lig profiles (Malusà et al., 2021, 2024; Solarino et al., 2024); both profiles evidenced lower crust-sub-Moho structures with a low-velocity bodies in a similar structural position of the HVDB in the Ligurian Sea-Po Basin transect. Though the former bodies are below the Adria Moho, they are considered excellent evidence of existence the fossil subduction zone of the Western Alps (Malusà et al., 2021, 2024; Solarino et al., 2024) as revealed also from the S-wave velocity model from ambient-noise tomography (Nouibat et al., 2022), complemented by P-wave local seismic tomography (Solarino et al., 2018, 2024) independently validated by receiver function analysis (Paul et al., 2022). These works unravel complex seismic velocity pattern consistent with the subducted European lithosphere. It was shown that the Dora-Maira U(HP) dome lays above partly serpentinized peridotites above the eclogitized lower crust of the (subducted) European plate (Solarino et al., 2018). Closer to our Ligurian Sea-Po Basin transect, in the Ligurian Alps, an exhumed mantle wedge and a former subduction channel was detected at (much) shallower levels beneath the Voltri (U)HP group and above a shallow-dipping lower-plate Moho (the question of the two Moho discussed by Solarino et al., 2024; Malusà et al., 2024). These studies argued that styles of subduction-channel and mantle wedge exhumation depend on increase of upper plate divergent motion from the Western Alps toward the Ligurian Alps (Solarino et al., 2024; Malusà et al., 2024). In addition, the deep structure of our transect has similarities with the seismic models obtained on two deep seismic sounding profiles crossing the Ligurian Sea and Tyrrhenian Sea (Solarino and Cassinis, 2007). They reveal two Moho interfaces: one below the Ligurian Sea at 18–20 km depth with Vp = 7.8–8.2 km/s plunging in the NE direction under the Moho of the Northern Apennines continental crust at a depth of 32–35 km with Vp = 7.5–8.1 km/s (Figs. 7 and 8 in Solarino and Cassinis, 2007). Moving north in the Central Alps, slices of subducted European crust are foreseen around 50 km depth beneath the Adria plate and partly exhumed due to orogen parallel extensional shearing (Montemagni et al., 2026). Therefore, following these analogies we may consider the HVDB as a remnant of the Alpine (Late Cretaceous-Eocene) subduction structures to the SE of those evidenced in the Central, Western and Ligurian Alps and NW of those evidenced in the Tyrrhenian Sea (Fig. 11c, d).

The previous considerations put forward two hypotheses for the origin of the HVDB: 1) remnant of Alpine subduction (Late Cretaceous – Eocene); 2) mantle shear zone related to back-arc basin formation (Oligocene-Present).

7.3.1. *The HVDB as remnant of Alpine obliquely subducted slab*

The subduction of the European plate under the Adriatic plate since Late Cretaceous (Fig. 11d) manifested itself in the lithosphere of the Western Alps by the plunging of the European plate lower crust into the upper mantle and the formation of an accretionary prism in the upper crust and exhumation zone with HP/LT mantle rocks that now are exposed as ophiolites of Penninic zone (Kissling, 1993; Nicolas et al., 1990; Scafidi et al., 2009; Malusà et al., 2015, 2021; Schmid et al., 1989, 2017). The above cited profiles CIFALPS and CIFALPS2-Lg revealed a low-velocity layer *E*- and NE dipping below the present-day Moho and possibly related to remnants of former subducted European lower crust below the Adria crust (Malusà et al., 2015, 2021; Solarino et al., 2024; Nouibat et al., 2022; Paul et al., 2022). In our revisited Ligurian Sea - Po Basin transect, the noted HVDB is dipping NE (Fig. 9) as the lower crust-upper mantle structure on the CIFALPS and CIFALPS2-Lg mentioned above. The following arguments could support similar interpretation for both the revisited transect and the transects across Western and Ligurian Alps. The revisited transect Ligurian Sea - Po Basin is located in close proximity to the junction area between the SW portion of the Western-Ligurian Alps and the northern sector of the Northern Apennines. The exposed UHP/LT rocks of the Voltri massif appear to prosecute into the Monte Doria Seamount and further south into the Alpine Corsica (See subchapter 7.2.1 Remnants of the Alps (Alpine Corsica) and exhumed units of HP rocks of the suture zone). Deep structure of the junction zone is characterized by its extreme complexity due to successive geodynamic environments and deep-seated processes in this S-shaped (knot) transition zone which can be considered the (dismembered) suture zone created by the collision of the European and Adria Plates. During Eocene-Oligocene, this suture zone became involved into the retro-wedge of the Alps extending from the Southern Alps to the Apennines

and to the Betics (Réhault et al., 1984; Doglioni, 1994; Gueguen et al., 1998; Faccenna et al., 2001; Carminati and Doglioni, 2012; van Hinsbergen et al., 2020; Schmid et al., 2017; Jolivet et al., 2021): the Alpine-Apenninic orogenic system began (Fig. 11c). The retrowedge build up is coeval to a significant plate movement reorganization which is testified by the NNW movement of the Adria plate faster than European plate (Corsica-Sardinia block) and resulted in the formation of a sinistral transpressive plate margin in the Western Alps (Malusà et al., 2011 among the others) and Apennines (Marroni et al., 2019) favoring exhumation of (U)HP/LT metamorphic rocks. Thus, in the schematic paleogeographic reconstructions, we show the possible position of the present day HVDB during the Alpine subduction (Figs. 11c) and during the retrowedge formation (Fig. 11c). In the latter, the northward movement of Adria of about 400 km (since Selandian 60 Ma to Rupelian 30 Ma in van Hinsbergen et al., 2020) allows to speculate that the present day lower crustal body was translated from the SE, was part of the Alpine subduction (Fig. 11d) and never exhumed till nowadays (Figs. 11a,b) except portions which might be exhumed within the Alpine metamorphic core (Fig. 11c). The estimated large amount of left-lateral translation and oblique convergence of Adria and Europe plates pose uncertainties about the original paleogeographic position of the HVDB. This also implies that the Alpine-Apenninic orogenic system might be formed by far travelled tectonic slices accreted during the oblique collision of the converging plates; HVDB is one of these slices.

For the above reconstructions, we take into account that the lithosphere of the Alps preserves other deep bodies which were revealed by deep sounding experiments of ECORS Project (Blundell et al., 1992; Scrocca et al., 2003) and integrated with surface geological data (Roure et al., 1996; Schmid and Kissling, 2000; Dal Piaz et al., 2003; Reddy et al., 2003; Schmid et al., 2004; Ford et al., 2006; De Graciansky et al., 2011; Schmid et al., 2017; Montemagni et al., 2026). These deep bodies, geophysically constrained and geologically envisaged but not yet fully understood, are structurally located around the suture zone of the collided European and Africa/Adria plates and, in some cases, paleogeographically located closer to the Adria passive margin. Portions of these bodies might be exhumed and cropping out in Sesia Lanzo zone or Western Austroalpine (sensu Dal Piaz et al., 2003) or in the Central Alps (Montemagni et al., 2026). The HVDB, revealed in the revisited lithospheric transect (Figs. 9, 10), occupies a similar structural position of these deep Alpine bodies and, in this respect, is still a remnant of the slab involved in the Alpine subduction but derived from the Mesozoic passive margin of Adria (Austroalpine/Insubric?) or of Europe, underlain and amalgamated to Adria/Southern Alps as shown in the Late Cretaceous-Early Paleocene reconstruction (Fig. 11d). These bodies, not exhumed and revealed by geophysical investigations, and gravity modelling as in our case, are involved in post-Oligocene Apenninic orogenic wedge evolution, from Apenninic subduction initiation as retro-wedge of the Alps (Fig. 11c) to Apenninic orogenic wedge dismemberment by Tyrrhenian back-arc extension (Fig. 11a).

7.3.2. The HVDB as a mantle shear zone related to back-arc basin opening

The other explanation on the origin of this high-velocity/high-density body (HVDB) derives from focusing on the Apenninic and post-collisional evolution. As mentioned various times, back-arc extension process was related with the opening of the Ligure-Provençal basin and Tyrrhenian Sea (Fig. 11a); the latter affecting the Northern Apennines with large-scale tensional tectonics producing low-angle and then high-angle normal faults within the NW-SE striking grabens (Lunigiana and Val di Vara) and eastward roto-translation of the Italian Peninsula moving away from Corsica-Sardinia block (Molli et al., 2018). This extensional tectonics thins most of the upper crust and could be the cause of this high-velocity/high-density body (HVDB) in the lower crust. But, during the last 30 Ma, changes in temperature and pressure conditions could have been changed; a heating event would soften the HVDB and move upward the Moho discontinuity: the HVDB became part of the upper mantle. As already remarked, Makris et al. (1999) considered the HVDB as a sub-Moho body and, anyhow, it is located at the lower crust/upper mantle transition with its specific architectural, physical and rheological properties, that could accumulate strength and serve as a kind of lower crust/mantle shear zone. This allows us to put forward the idea of the HVDB as an analogue of a hydrated mantle shear zone accommodating the Ligure-Provençal and, more recently, Tyrrhenian related extension as well as transform fault. Unlike brittle fractures in the crust, shear zones in the mantle involve slow, ductile flow of the rock under high pressure and temperature. Mantle shear zones are often detected on seismic reflection profiling images as shallow-dipping seismic reflectors in the lithospheric mantle interpreted as the continuation of normal faults and strike-slip faults in the upper mantle (Vauchez et al., 2012; Reston, 1990). There are many examples from British Caledonides, North Sea and Slave craton of Canada of Moho offsets and sub-crustal reflections aligned with the trace of crustal faults suggesting that these faults penetrate the upper mantle (Reston, 1990; Vauchez et al., 2012) as well as from exhumed mantle peridotites (Hoogerduijn Strating et al., 1993, Vissers et al., 1995; Dijkstra et al., 2004; Vauchez et al., 2012). The upper mantle shear zones preserved in the ophiolite Voltri Massif (Erro-Tobbio peridotite in Vissers et al., 1995) is an example in the area adjacent to our transect.

Then, in mantle shear zones, hydrated (serpentinized) rocks play a significant role in subduction zones and oceanic transform faults. These shear zones, often associated with serpentinization, act as fluid pathways, influencing earthquake activity, fluid flux, and the overall strength of these tectonic boundaries. In the revisited transect Ligurian Sea – Po Basin the supposed mantle shear zone associated with velocities of 7.2–7.4 km/s could indicate a moderately hydrated mantle composition (~30% serpentinization) that are not much different from unaltered mafic crustal rocks (Hyndman and Peacock, 2003). Thus, the Moho-related shear zone on the Ligurian Sea – Po Basin transect could be a continuation of the NW-SE transform fault that bounds the Monte Doria Seamount area (Figs. 1, 9 and 10). This fault is a steep fault that flattens at the present day Moho (Fig.9). Gravity modelling does show a sharp boundary along the transform fault zone where extended crust and upper mantle low-velocity/density layer are in contact (Fig. 9). The transform fault zone is evident down to a depth of about 24 km where it likely flattens (Fig. 10 a, b); deeper discontinuities in the lithospheric mantle are not revealed by our gravity modelling but not excluded (whitened area in Fig. 10b, c).

Indeed, this lower crust/upper mantle shear zone could be interpreted also as a basal shear zone or basal detachment related to low-angle, crustal-scale faults that formed during continental rifting (Manatschal, 2004; Lavier and Manatschal, 2006; Péron-Pinvidic and Manatschal, 2009). This detachment facilitates the separation of the upper and lower crust from the upper mantle, leading to formation of significant crustal blocks and, eventually, passive continental margins. These basal detachments are characterized by ductile deformations in the shear zone, often associated with mylonites, and it is typically capped by brittle fault rocks.

Finally, as additional support to this alternative interpretation of this lower crustal/upper mantle body (HVDB), the study of the continental margins of Iberia and the ancient margin exposed in the Alps have shown that the rifted margins and the position of the final break-up are controlled by inherited heterogeneities within the lithosphere and that during the final break-up, the rheology of the extending lithosphere is controlled by the thermal structure related to the rise of the asthenosphere and by serpentinization process (Manatschal, 2004). The basal detachment zones are formed by a system of conjugate faults that accumulate large amounts of extension and form a thin crust by exhumation of (upper) mantle material (Lavier and Manatschal, 2006). All the above lithospheric scale shear zone related to the passive margin formation could be represented by the HVDB.

Nonetheless, it can not be excluded that this mantle shear zone with envisaged hydrated mantle and related to Ligure-Provençal and Tyrrhenian extension and transform fault was reusing or superposed on the

remnants of inherited obliquely subducted slab of the Alpine subduction system: lithospheric inheritance likely favored localization of shear zones and asthenospheric flows.

7.4. Low-velocity low-density layers in the mantle lithosphere

Rifting occurred in the area of the modern Ligurian Sea (since middle Miocene - ~20–15 Ma) led to formation of the modern Western Mediterranean Basin and southwards opening of the Tyrrhenian Sea under the influence of asthenospheric flows. The latters, recorded by the shear-wave tomography of Italy (Brandmayr et al., 2010), are associated with the low-velocity layers below the Moho in the offshore south-western portion of the transect and in the upper mantle below the major part of the transect at the depth ≥ 120 km. Brandmayr et al. (2010) interpreted these layers as thin soft mantle lithosphere layer and top of the asthenosphere or LAB (Lithosphere-Astenosphere Boundary) (Fig. 5). Corresponding zones of low density (up to 3.20–3.25 g/cm^3) are present in the mantle lithosphere of our density model (Figs. 9, 10). The distribution of the high terrestrial heat flow zone above the offshore extended continental crust strictly corresponds to the shallow heterogeneity confirming their recent origin and caused by overheated and ductile (soft) portions of the mantle lithosphere. The Low-velocity and low-density layer determined at the depth of 120–140 km are associated with the LAB below the Ligurian Sea – Po Basin transect. Looking at the regional context of continental rift evolution of Jolivet et al. (2025), the above described features of the upper mantle – asthenospheric layers in the study transect could be related to the short-lived magma-poor period in the European Cenozoic Rift System during the transition from the collision of Africa, Iberia and Eurasia (forming the Alps and Pyrenees) to the fast slab retreat forming the Mediterranean back-arc basins (Jolivet et al., 2025; Séranne, 1999; Ziegler and Dézes, 2007). This confirms that the Ligurian Sea-Po Plain transect preserves a younger imprint of the magma-poor rifting zones in the Ligurian Sea portion.

According to us, based on the data so far collected and integrated in the new gravity model along the newly compiled lithospheric transect Ligurian Sea-Po Basin, there are no compelling constraints which favour one of the two possible proposed interpretations on the origin of the HVDB remarked and discussed in this work. Focused tomographic images, possibly derived from recent AlpArray dataset (www.alparray.ethz.ch) (Kolínský et al., 2019; Wolf et al., 2021; Paffrath et al., 2021; Handy et al., 2021; Friederich et al., 2025), additional integrated transects with similar gravity/density modelling,lithospheric modelling and comparison with similar lithospheric setting of convergent collisional margins that are changing fast, such as those in SE Asia (Hall and Wilson, 2000; Morley, 2009; Spakman and Hall, 2010; Molli and Malavieille, 2011; Pownall et al., 2016; Liao et al., 2018; Morley and Wang, 2023), will may help to define the most proper interpretation of this deeply hidden lower crust/upper mantle body.

8. Summary and conclusions

● A new lithospheric transect was constructed from the Ligurian Sea through the northwestern portion of the Northern Apennines to the Po Basin as far as Verona province. It is 465 km long, SW-NE directed and integrates three composite seismic reflection profiles, magnetic and gravity anomalies, surface heat flow, refraction seismology, and surface wave tomography. The integrated composite transect was the base for constraining a new lithospheric density/gravity model.

● This new lithospheric Ligurian Sea-Po Basin transect crosses: 1) the extended Alpine chain in the Ligurian sea; 2) the complex stack of embricated thrust sheets of Northwestern Apennines, partly extended (approximately as far as the Apenninic watershed) and partly still shortening the pre-late Messinian foredeep depocenters which are buried underneath the Ligurian allochthonous nappes; 3) the late Messinian to recent foredeep of the Po Plain; 4) the foreland area of the Northern Apennines within the Southern Alps foothills and retrowedge of the Alps.

● The new integrated gravity modelling shows a suite of crustal types from a thin (18 km) extended continental crust in the Ligurian Sea to ~40 km-thick continental crust with a thick (up to 18 km) meta-sedimentary and sedimentary succession of the Po Basin; in between, the offshore-onshore transition domain.

● The offshore-onshore transition domain, as wide as ~150-km, is characterized by crustal thickness increasing northeastwards from 18 km to 35 km depth; it includes most of the Northern Apennines orogen. This transition domain is abruptly separated from the deep-water offshore oceanic domain of the Ligurian Sea by a narrow block confined by steep transform faults and marked by local magnetic anomaly. We associate this domain with the exhumed HP/LT alpine metamorphic rocks belonging to the suture zone of the collided Europe-Adria plates which crop out in the Alpine Corsica and in Voltri Massif; all of them being part of the junction area between the southern sector of Western-Ligurian Alps and northern portion of the Northern Apennines or the Ligurian Knot.

● Inside this offshore-onshore transition domain, a notable lithospheric feature is observed. It is a lower crust high-velocity/density body (HVDB) as thick as ~7-km that starts around 20 km depth and deepens and thins out northeastward; it ends at 38 km depth beneath the watershed of the Apennines.

● This HVDB could be related to Apennines hinterland extension caused by a hydrated (serpentinized) mantle shear zone accommodating the Ligure-Provençal and Tyrrhenian extension (Oligocene-Recent) associated to back-arc basin formation. But it is not excluded that it could represent an inherited remnant of the Alpine subduction (pre-Oligocene) welded/amalgamated to the Adria plate obliquely colliding with the Europe plate and partly remoulded by younger events including the back-arc basin opening which dismembers the Alpine-Apenninic orogenic system.

● The new lithospheric transect allows also to link tectonic processes to asthenospheric flows evidenced by S-wave tomography and verified by gravity modelling. Two low-velocity and low-density layers are evidenced in the subcrustal portion and in the mantle lithosphere. The first and shallowest one is a thin sub-Moho layer limited to the offshore south-western portion of the transect; it corresponds to the high heat flow zone confirming its relation to recent mantle lithosphere (re)activation just below the offshore oceanic domain of the Ligurian Sea with thinner and extended continental crust; it is associated with overheated and ductile (soft) portion of the upper mantle. The second and deepest one is at the depth of ≥120–140 km beneath the whole Ligurian Sea – Po Basin transect; it corresponds to the LAB (Lithosphere-Asthenosphere Boundary).

● The new integrated gravimetric modelling allows to provide a new onshore-offshore lithospheric transect which provides a unifying and updated geological model of the Northern Apennines and associated sedimentary basins. This updated model sheds new lights on the complexity of the Alpine-Apenninic orogenic system in relationship with the obliquely colliding Europe and Adria continental plates and provides an evolutionary model since late Cretaceous of a key area ("knot") of the Mediterranean Region.

CRediT authorship contribution statement

T. Yegorova: Writing – review & editing, Writing – original draft, Visualization, Validation, Supervision, Software, Methodology, Formal analysis, Data curation, Conceptualization. **L. Torelli:** Supervision, Data curation, Conceptualization. **A. Artoni:** Writing – review & editing, Writing – original draft, Visualization, Validation, Supervision, Software, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization. **O. Verpakhovska:** Writing – original draft, Visualization, Software, Methodology, Data curation. **A. Murovskaya:** Writing – review & editing, Methodology, Investigation, Formal analysis, Data curation. **G.**

Molli: Writing – review & editing, Writing – original draft, Visualization, Validation, Supervision, Methodology, Investigation, Formal analysis, Data curation. **A. Qadir:** Writing – review & editing, Visualization, Software. **N. Chizzini:** Writing – review & editing, Visualization, Validation, Methodology. **S. Cioce:** Writing – review & editing, Visualization. **F. Camurri:** Writing – review & editing, Validation, Software, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **F. Rizzini:** Writing – review & editing, Software, Investigation, Formal analysis. **F. Storti:** Writing – review & editing, Resources, Investigation, Funding acquisition.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgments

This work was jointly supported by the ARTONI_A_FIL (University of Parma Funding), BORSE_UCRAINA_ARTONI and BORSE_UCRAINA_STORTI (borse di ricerca per ricercatori ucraini Università di Parma-MUR). This work has benefited from the equipment and framework of the COMP-R Initiative, funded by the 'Departments of Excellence' program of the Italian Ministry for University and Research (MUR, 2023-2027). We would like to sincerely thank the Guest Editors Jimin Sun and Pietro Sternai for handling our manuscript and providing constructive comments. We are also very thankful to Piotr Krzywiec for his support and two anonymous reviewers for their throughout reviews and insightful suggestions which have significantly improved the quality of this paper.

The paper is the result of a resilient experience, a drop in the ocean as we usually call it, consequence of Russia-Ukraine war which we all wish stop very soon. This experience could be realized with the help of many people who are not named above: administrative staff of the University of Parma, of the Ministery of University and Research (MUR), of the Institute of Geophysics of National Academy of Sciences of Ukraine in Kiev, of the Journal's Editorial management. They are all thanked for being collaborative and for demonstrating that non-fighting community is the base for realizing projects in peace.

We dedicate this paper to **Prof. Luigi Torelli** who passed away during the preparation of this work. We are grateful to him and his collaborators who dedicated many efforts to unravel this "intriguing" area. His and their works gave new insights which help to reach the results here presented as well as to open new questions; as Prof. Luigi Torelli was used to say looking at the data and interpretations: "yes, but now the problem is .....".

Data availability

The original data generated in this study are provided in the text and are also available from the authors upon request.

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