# Ivrea mantle wedge, arc of the Western Alps, and kinematic evolution of the Alps-Apennines orogenic system

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**Abstract**

The construction of five crustal-scale profiles across the Western Alps and the Ivrea mantle wedge integrates up-to-date geological and geophysical information and reveals important along strike changes in the overall structure of the crust of the Western Alpine arc. Tectonic analysis of the profiles, together with a review of the existing literature allows for proposing the following multistage evolution of the arc of the Western Alps: (1) Exhumation of the mantle beneath the Ivrea Zone to shallow crustal depths during Mesozoic is a prerequisite for the formation of a strong Ivrea mantle wedge whose strength exceeds that of surrounding mostly quartz-bearing units, and consequently allows for indentation of the Ivrea mantle wedge and eastward back-thrusting of the western Alps during Alpine orogeny. (2) A first early stage (pre-35 Ma) of the West-Alpine orogenic evolution is characterized by top-

NNW thrusting in sinistral transpression causing at least some 260km displacement of internal Western Alps and E-W-striking Alps farther east, together with the Adria micro-plate, towards N to NNW with respect to stable Europe. (3) The second stage (35-25 Ma), further accentuating the arc, is associated with top-WNW thrusting in the external zones of the central portion of the arc and is related to the lateral indentation of the Ivrea mantle slice towards WNW by some 100-150km. (4) The final stage of arc formation (25-0 Ma) is associated with orogeny in the Apennines leading to oroclinal bending in the southernmost Western Alps in connection with the 50° counterclockwise rotation of the Corsica-Sardinia block and the Ligurian Alps. Analysis of existing literature data on the Alps-Apennines transition zone reveals that substantial parts of the Northern Apennines formerly suffered Alpine-type shortening associated with an E-dipping Alpine subduction zone and were backthrusted to the NE during Apenninic orogeny that commences in the Oligocene.

## A. Introduction

The Western Alps represent a prime example of a very arcuate mountain belt curving around from an E-W strike with dominant top-N nappe stacking in the Eastern Alps to a N-S strike in the northern Western Alps and finally into an E-W strike with top-S nappe stacking in the Ligurian Alps straddling the Mediterranean coast (Fig. 1). The question after the mechanisms of arc formation is an old and much debated issue in many orogens (Marshak 1988) involving, in the case of the Alps, elements of oroclinal bending of an initially straight orogen (Carey 1955, van der Voo 2004), induced by rigid body rotation of the hinterland (Vialon et al. 1989) and/or by W-directed indentation of the Adria microplate (Laubscher 1971; Schmid

& Kissling 2000). Transpressional movements associated with large top NNW displacements of portions of the inner zone of the Western Alps that occur at a small angle to the present-day N-S strike of the Western Alps were also invoked for being responsible for explaining the curvature from the Eastern Alps into the N-S striking Western Alps (Ricou & Siddans 1986). There have also been attempts to explain diverging kinematic indicators along the arc of the Western Alps ranging from top-N Western Switzerland to top-SW in southern France without taking resort to a multi-stage evolution allowing for strike-slip movements and/or large rotations (Platt et al. 1989; their group 3 kinematic indicators formed in the 40-15 Ma time interval). However, as pointed out by Butler (1986; his Fig. 2), strongly divergent thrust transport directions over large distances inevitably result in excessive orogen-parallel stretching.

In the Western Alps many previous workers (e.g. Laubscher 1971; Butler 1983) recognized the crucial role of the Ivrea mantle wedge residing at relatively shallow crustal levels and locally reaching the surface. Novel geophysical methods such as teleseismic tomography aiming at the structure of the earth's mantle (Lippitsch et al. 2003) and crustal tomography based on local earthquake data (Diehl et al. 2009), together with older information based on controlled source seismology (Blundell et al. 1992), led to substantial recent improvements in revealing the deep structure of the arc of the Western Alps. Th new geophysical data set of unprecedented quality calls for integration with a wealth of geological data collected over many decades to elucidate the formation of the arc of the Western Alps that also demands to analyze the Alps-Apennines transition area located in the Ligurian Alps and the Northern Apennines (Molli et al. 2010).

This contribution integrates the new results of crustal tomography and improved knowledge on Moho depth (Spada et al. 2013) with a geological interpretation of five crustal-scale transects across the Western Alps. Thereby we focus on the role of the Ivrea mantle wedge. Another important aim of our geological-geophysical interpretations is to put new constraints on the question as to how much continental crust was subducted in the Alps and what the variations in pre-orogenic crustal thickness might have been (see discussions by, e.g., Ménard & Molnar 1991; Butler et al. 2013; Mohn et al. 2014).

Two transects are revised versions of previously published profiles (Schmid et al. (2004) while three of them were newly constructed. Special emphasis is given to improved mapping of the outlines of the Ivrea mantle wedge based on geophysical data. This is then followed by an analysis of the complex multi-stage evolution of the arc of the Western Alps and an attempt to map the associated displacement history of geological units of the Western Alps-Northern Apennines system using plate reconstruction methodology (GPlates free software (Boyden et al., 2011) embedded in the global plate kinematic reconstruction of Seton et al. 2012). Finally, we provide a review-type discussion of the Alps-Apennines transition area and discuss controversial issues concerning questions of multi-stage vs. single subduction systems in the Apennines.

## B. Crustal-scale profiles across the Western Alps and the Ivrea mantle wedge integrating geological and geophysical information

To constrain the geometry of the Ivrea mantle wedge, and as a basis for discussing its role during the formation of the arc of the Western Alps, we constructed a series of crustal scale profiles along the arc of the Western Alps (Fig. 1).

These profiles integrate geological data with recent geophysical data focusing on the deep structure. The construction of the profiles is based on a series of assumptions, the most important one being that advocated by Schmid & Kissling (2000), namely that the European lower crust is decoupled from the upper crust and goes into subduction together with the underlying mantle for considerable distance below the Ivrea mantle wedge. This invokes a basal detachment of the external thrust sheets at the interface between upper and lower crust at a depth of 20-30km. While this assumption is compatible with the velocity structure (see below) and recent geophysical findings of Zhao et al. (2015) it is by no means generally accepted (see discussion in Butler 2013). However, there are considerable uncertainties about the exact pre-orogenic thickness of the European and Briançonnais continental crust in more internal parts of the profiles and this is one reason why we restrain from area balancing over the entire length of our transects. The main reason is that transport directions changed with time during the formation of the arc of the Western Alps, making 2D area balance along the transects impossible.

The geological literature used for constructing the individual profiles varies and hence will be cited when presenting the profiles. However, we used one and the same set of geophysical data for constraining the deep structure in all the profiles. Firstly, we used the results of a tomographic high-resolution 3-D P-wave model of the Alpine crust obtained by Diehl et al. (2009). This 3-D velocity model was used for constructing parts of the sections of Fig. 2 that present contours of equal P-wave velocity superimposed with a geological interpretation. When comparing the velocity contours with the geological interpretation it has to be kept in mind, however, that these contours have a low spatial resolution since they are based on assigning a given velocity to a sizable cell volume of 20x20 km (horizontal) and 15km (vertical).

Secondly, we complemented these tomographic data by geophysical data better constraining Moho-depth. Where available we took Moho depths from Spada et al. (2013; their Fig. 11; see also Wagner et al. 2012 regarding the methodology) who combined results of controlled-source seismology with local receiver functions Moho depth maps and local earthquake tomography-derived Moho maps.

Comparison between Moho depth compiled by Spada et al. (2013), characterized by a P-wave velocity jump, and the velocity model of Diehl et al. (2009) that defines Moho depth in terms a high velocity gradient in the range of 6.5–8.0 km $s^{-1}$, noticeable over large parts of the profiles, shows that on average the Moho as defined by Spada et al. (2013) best corresponds with the velocity contour of 7.25 km $s^{-1}$ in areas characterized by a subhorizontal or only weakly inclined Moho (Diehl et al. 2009). In order to define the strongly inclined boundaries of the Ivrea mantle wedge at shallower depth <30km, not well constrained by the compilation of Spada et al. (2013), we followed the 7.0 km $s^{-1}$ contour in all profiles.

In the profiles ECORS-CROP, NFP20-West and Ticino rather thin slices of the Ivrea wedge and/or the lower crust of the Adria microplate, geologically speaking the Ivrea Zone (Handy & Zingg 1991), reach the surface. In these profiles the outlines of the Ivrea mantle wedge are too thin to be resolved by the P-wave velocity model of Diehl et al. (2009) given its limited spatial resolution. The geometry of the very thin and strongly dismembered Ivrea mantle wedge and Adriatic lower crust is only tentatively drawn, being only constrained by the surface exposures.

Another problem one faces when determining the shape of the Ivrea mantle wedge is that the velocities >7.25 km $s^{-1}$ expected for the Ivrea mantle rocks that are part of the Adriatic microplate and the adjacent European mantle become indistinguishable from velocities expected for potential remnants of the subduction channel separating European and Ivrea (=Adriatic) mantle marking the present-day plate boundary at great depth. At a depth of 30-45 km and temperatures around 400°C- 500°C, we expect to reach blueschist facies conditions, the exact depth depending on temperature (Oberhänsli et al. 2004). Based on thermal modeling (Bousquet et al. 1997) the expectation of temperatures between 450° and 500°C at a depth of 30km is realistic. The P-wave velocity of blueschist rocks at 400°C and 1 GPa is around 7.25 km $s^{-1}$ on average according to Fujimoto et al. (2010). From a depth of around 50km onward, again depending on temperature, we expect eclogite facies conditions associated with even higher velocities around 8.0-8.5 km $s^{-1}$ (Worthington et al. 2013). Hence, the contours of velocities >7.25 km $s^{-1}$ are expected to cross over from the European to the Ivrea mantle rocks traversing the subduction channel formed by high-P metamorphic rocks. Such crossover is actually seen in all the profiles of Fig 2, also for velocity contours at 7.0, 6.5 and occasionally at 6.5km $s^{-1}$. This is because velocities expected for the subducted European lower crust (around 6.5 km $s^{-1}$; Ye et al. 1995) are similar to those expected for the stack of nappes that is northerly adjacent to the Ivrea mantle wedge, nappes that are only partially overprinted by high-P metamorphism (e.g. Henry et al. 1993). All this, together with the limited resolution of the 3-D velocity model, puts severe limits to a straightforward correlation of velocities with geological-tectonic entities, particularly at greater depths.

### **1. Argentera profile (Fig. 2a)**

The external SW part of this profile is characterized by minor shortening of around 21km in Mesozoic sediments (Chaînes Subalpines) of the European foreland (Lickorish & Ford 1998), associated with rather modest late-stage up-thrusting of the

Argentera external massif (Malaroda et al. 1970) during the Neogene. In this external part of the profile the European Moho coincides with a strong velocity gradient between 6.5 and 7.5 $s^{-1}$ in the velocity model of Diehl et al. (2009).

The still unfolded Chaînes Subalpines were thrust by sedimentary units attributed to the Subbriançonnais, together with non-metamorphic Helminthoid flysch detached from the Piedmont-Liguria paleogeographical domain above ophiolite-bearing mélanges (Autapie and Parpaillon nappes; Merle & Brun 1984). Final emplacement of these higher nappes occurred during the Late Eocene (Kerckhove 1969), i.e. before thrusting along the late Alpine (Early Oligocene) out-of-sequence thrust called "Penninic Front" and Neogene shortening in the Chaînes Subalpines. Since the Autapie and Parpaillon nappes lack metamorphic overprint they are considered as units derived from a part of the Piedmont-Liguria Ocean that always remained in an upper plate position. They form relics of the front of the overriding oceanic upper plate (Piedmont-Liguria Upper Plate units of Fig. 2a) that started to thrust the oceanic lower plate (Piedmont-Liguria Lower Plate units of Fig. 2a) from the late Cretaceous onwards, reaching the front of the Alps at the end of the Eocene, together with the underlying Subbriançonnais units.

There is a rather abrupt lateral change of velocities at a given depth across the boundary between external and internal Alps. This change coincides with the downward projection of the southern end of a very steeply NNE dipping late Alpine out-of-sequence thrust well documented in the Pelvoux section and known as the Penninic Front (Ceriani et al. 2001; Lardeaux et al. 2006; Loprieno et al. 2011; see description of the Pelvoux section). The term "Penninic Front", synonymous with the term "Briançonnais Frontal Thrust" coined by Tricart (1986) who first recognized the importance of this Early Oligocene out-of-sequence thrust in marking the boundary between internal and external parts of the Western Alps, is a misnomer in that earlier emplaced Penninic units are occasionally also found in front of this out-of-sequence thrust (Ubaye-Embrunais of profile in Fig. 2; Préalpes Romandes of Fig. 2d; see also Fig. 1). We use the term Penninic Front nevertheless because it became deeply entrenched in the literature by now. This top WMW Penninic Frontal Thrust turns into the sinistral Acceglio-Cuneo Line (Fig. 1) and is characterized by a sinistrally transpressive component as it swings around into the WNW-ESE strike of the Argentera-Cuneo line whose western end is traversed by the Argentera profile. The sinistral Acceglio-Cuneo Line allows for the WNW-directed indentation of the Ivrea mantle wedge along its southern boundary located near Cuneo (Trullenque 2005; Ceriani et al. 2001, see their Fig. 1). Laubscher (1971) first proposed the existence of a sinistral strike slip zone (his "Ligurian shear zone") allowing for westward indentation of the Ivrea mantle wedge for kinematic reasons (see also Laubscher 1991). Based on the interpretation of fission track data (Fügenschuh & Schmid 2003) and radiometric dating of mylonites (Simon-Labric et al. 2009) the thrust marking the Penninic Front (Roselend thrust of Loprieno et al. 2011) was active during the Early Oligocene (34-30 Ma). By downward extrapolation the steep dip of this crustal scale transpressive fault projects (Malaroda et al. 1970) into parallelism with the steeply overturned 6.5 $s^{-1}$ velocity contour that marks the SW edge of a high-velocity area with > 6.5 $s^{-1}$ located within the high-pressure Briançonnais units at a depth of >10km. This high velocity area is interpreted to consist of the high-pressure Briançonnais units that include the Dora Maira "massif" exposed at the surface at the eastern margin of the Internal Alps (Henry et al. 1993). We are aware that not all exposed parts of the Dora Maira massif constituting the high pressure Briançonnais show blueschist facies or eclogitic parageneses (Bousquet et al. 2012a). However, the high velocities at depth require that most of the volume taken up by the Briançonnais at depth must have undergone high-pressure metamorphism.

The shallow crustal parts of the metamorphic core of the Alps in this profile are characterized by large-scale back-folds that formed between 35-30 Ma (Bucher et al. 2004) and which affected pre-existing nappe contacts between: (1) the high-pressure Briançonnais units, (2) overlying so-called Acceglio and/or Pre-Piémontais units (Lemoine & Tricart 1986; Caby 1996), representing the former transition of the Briançonnais micro-continent into the Piedmont-Liguria Ocean, (3) Piedmont-Liguria high pressure units and, finally, (4) non-metamorphic Piedmont-Liguria units that are only preserved along strike in the area of the Chenaillet near Briançon (Manatschal et al. 2011). We attribute the Chenaillet ophiolites to that part of the Piedmont-Liguria Unit that occupied an upper plate position in the internal Western Alps (Schwartz et al. 2007) where, in our interpretation, the Austroalpine nappe system is totally lacking due to a westward lateral transition of the active margin of the Late Cretaceous to Paleogene Alpine subduction zone from the northern edge of the Adria continent into an intra-oceanic location (Molli, 2008; Handy et al. 2010). The geometry of the shallow crustal parts of the Internal Alps depicted in this profile closely follows the geometries shown by the works of Henry et al. (1993), Michard et al. (2004), Lardeaux et al. (2006), and Ford et al. (2006), the surface-near parts of the Penninic Front exposing the Subbriançonnais and non-metamorphic Briançonnais units was drawn after a series of detailed profiles given by Gidon (1972).

The 7.0 $s^{-1}$ velocity contour that runs parallel and near the Moho depicted by controlled-source seismology in the Adria microplate rises SW-ward to a depth of some 15 km in the NE part of the profile beneath the HP Briançonnais. It is taken as marking the outline of the top of the Ivrea mantle wedge. The inferred overall geometry demonstrates that crust and mantle of the Adriatic microplate became wedged into the Alpine edifice and underplated shallow crustal layers in the internal parts of the Alps. The dip of the Periadriatic line is constrained to a depth of some 30 km by connecting the southern prolongation of the Periadriatic Line (Fig. 1), delimiting the eastern edge of the root of the Piedmont-Liguria units near the eastern margin of the Dora Maira massif at the surface, with the western front of the Ivrea mantle wedge paralleling the 7.0 $s^{-1}$ contour. The exact geometry shown for the downward narrowing deepest parts of internal Alps, i.e. the subduction channel separating the European and Adriatic (micro) plates, is basically unconstrained by geophysical data and the details shown are merely conceptual. The downward bending of the 7.0 and 7.5 s-1 velocity contours across the suspected subduction channel is unable to depict the geometry of the subduction channel, due to the high velocities expected for the rocks within the subduction channel undergoing high-pressure metamorphism, velocities expected to be similar to those of the European lower crust, and at greater depth, similar to those of the adjacent mantle rocks.

The Oligocene age of formation of the steeply dipping root of the Piedmont-Liguria Ocean, associated with rock uplift associated with exhumation (Fox et al. 2016) of these high-pressure units along the front of the adjacent Adria microplate can be inferred from the cooling history of the Dora Maira high pressure unit evidencing rapid cooling from 550°C to some 260°C within an extremely short time interval between 32 and 30 Ma according to the pressure-time path constrained by petrological data in combination with cooling ages derived from isotopic dating and fission track analysis (Gebauer et al. 1997; Rubatto & Herrmann 2001; Berger & Bousquet 2008). The earlier stages of this exhumation are possibly related to exhumation within the subduction channel, triggered by the buoyancy of continental rocks at mantle depth (Chemenda et al. 1995). These time constraints indicate that both the formation of the steeply hinterland-dipping Penninic Front and the wedging of the Adria mantle and crust associated with backthrusting of the internal Alps over the Tertiary Piedmont Basin are contemporaneous. Both appear to be associated with the WNW-directed indentation of the Ivrea mantle during the earliest Oligocene (34-30Ma).

The shallow NE crustal parts of the profile are covered by the Oligo-Miocene of the Tertiary Piedmont Basin and younger Plio-Pleistocene sediments. The construction of the profile is based on subsurface data acquired by AGIP (Pieri & Cropi 1981, Cassano et al. 1986, Schumacher & Laubscher 1996) that only vaguely constrain the geometry depicted. The 5.5 to 7.0 velocity contours, as well as the Moho depth indicated by controlled-source seismology, indicate that within the Adria microplate Moho, lower crust and basement-sediment interface are E-dipping.

In the immediate vicinity of the Periadriatic line the thickness of the Tertiary Piedmont Basin along the profile is considerable. Carrapa & Garcia-Castellanos (2005) showed that the subsidence of the Tertiary Piedmont Basin is controlled by far-field compression and thrust loading rather than by extensional tectonics. Hence, the early stages of thrust loading are probably due to the Early Oligocene backthrust of the metamorphic units of the Western Alps onto the Adria microplate.

However, at the eastern end of the profile the Tertiary Piedmont Basin overlies "Apenninic" units (Fig. 2a). Compressional tectonics continued into the Miocene when the Ligurian Alps became involved in N-directed thrusting, together with Apenninic units exposed in the northern Monferrato hill (Fig. 1) west of Torino (Piana 2000), first during the so-called Paleo-Apenninic phase (Schumacher & Laubscher 1996) sealed by a Mid-Miocene unconformity. Later, Neo-Apenninic thrusting formed the morphologically visible thrust front north of the Torino and Monferrato hills. It is this N-directed Apenninic thrusting over the Adria microplate that involved parts of the former Alps referred to as "Apenninic" units in the profile of Fig. 2a (see later discussion on the Alps-Apennines transition). Differential N-directed thrusting of joint Apenninic and Ligurian Alpine units calls for a sinistral strike slip zone that separates the joint Apennines-Ligurian Alps unit moving to the north from those parts of the Piedmont-Liguria units that are immediately adjacent to the Dora Maira backthrust, such as depicted near the eastern end of the section (Fig. 1a). The exact 3D geometry of this complicated Early Miocene to recent scenario still remains largely unresolved and is known as the problem of the "Ligurian knot" (Laubscher et al. 1992).

## 2. Pelvoux profile (Fig. 2b)

The western part of this profile crosses the Vercors massif (Chaînes Subalpines) and the more internal Belledonne–Grandes Rousses–Oisans external massifs and their Mesozoic cover (Philippe et al. 1998; Dumont et al. 2008) (Fig. 1). Profile balancing by Bellahsen et al. (2012), taking into account the effects of normal faulting preceding Alpine orogeny (e.g. de Graciansky et al. 1989) related to the opening of Alpine Tethys, the total amount of Miocene shortening in these external parts of the Alps (Dauphinois) amounts to some 28km.

In the external parts of the profile the P-wave velocity contours within the European lithosphere again exhibit a higher velocity gradient between 6.5 and 7.5 $s^{-1}$ near the Moho. However, the 6.0 $s^{-1}$ contour rises to an unusually shallow depth of <10km twice, a feature that is considered robust and points to the presence of high-velocity material at shallow depth. The more external amongst the two anomalies is found underneath the Belledonne and Grandes Rousses massifs, an area that is known for the occurrence of the pre-Alpine Chamrousse ophiolitic complex (Guillot & Ménot 2009), associated with ophiolites and eclogites being part of a former Variscan suture zone. This suture zone evolved into a major SW-NE striking and ca. 1500 km long dextral wrenching zone in Late Carboniferous times, marking the SE margin of the European Variscides with Gondwana. This broad mega shear zone is known as the External Massifs Shear Zone (Guillot & Ménot 2009) or East Variscan Shear Zone (Padovano et al. 2012). The interpretation of this anomaly in terms of high-velocity material constituting this steeply inclined mega shear zone is admittedly speculative. The origin of the second more internal anomaly is unknown but possibly also related to velocity inhomogeneities related to pre-Alpine structures.

At the surface the Penninic Front is exposed as a spectacular thrust zone behind the Pelvoux massif and in the footwall of the Massif du Combeynot, dipping with ca. 35° to the E and associated with top-WNW out-of-sequence thrusting (Butler 1992; Ceriani et al. 2001; Ceriani & Schmid 2004, Trullenque et al. 2006). The linear extrapolation of this dip observed at the surface to depth is unconstrained by currently available geophysical data along this profile. Fig. 2b opts for an extrapolation of the Penninic Front to great depth such as to join the subduction channel at the base of the Ivrea mantle wedge, consistent with what was observed along the Argentera profile described above. The 6.5 and 7.0 $s^{-1}$ contours that followed the lower crust of the European lithosphere in the external Western Alps start to rise to shallower levels at an angle of around 45° after having crossed the Penninic Front, indicating that westward they enter high-pressure Briançonnais units.

The shallow crustal parts of the metamorphic core of the Alps are drawn according to data published in Barféty et al (1995), Caby (1996), Ganne et al. (2006), and Lardeaux et al. (2006). In the Briançonnais units sedimentation continued until the Midle (internal Briançonnais) to Late Eocene (external Briançonnais) (Jaillard 1999) and hence predates thrusting at the Early Oligocene Penninic Front. They very much resemble those found along the Argentera profile. Again, almost the entire section across the inner Western Alps, except the very frontal part, is characterized by late-stage backfolding of pre-existing nappe contacts. The geometry of the top of the Ivrea mantle wedge is constrained by the 7.0 $s^{-1}$ velocity contour and controlled seismology data. Exact location and width of the subduction channel remain unknown, but an inverse velocity gradient of the 7.5 $s^{-1}$ contour associated with a down-bending of the same contour line points to the existence of a subduction channel below the Ivrea mantle wedge. A similar, although not identical geometry of a geological transect located near the Pelvoux profile was proposed by Zhao et al. (2015), who also postulated subduction of European continental lower crust all the way below the Ivrea mantle wedge, based on their interpretation of seismic imaging by P-wave receiver function analysis. In comparison, their model proposes a much shallower depth for the top of the Ivrea mantle wedge (at around 10km) and a much wider (about 30km) subduction channel below the Ivrea mantle wedge, features that are not compatible with the 3-D velocity model of Diehl et al. (2009) and the Moho depth as compiled by Spada et al. (2013).

The eastern parts of the profile are very similar to what was described for the Argentera section. They cross the northern parts of the Dora Maira massif and exhibit the same overturned root of the ophiolitic series of the Piedmont Lower Plate unit, juxtaposed with the front of the Adria microplate including the Ivrea mantle wedge along a backthrust onto the Tertiary Piedmont Basin, formed in the earliest Oligocene, i.e. contemporaneous with the formation of the Penninic Front. At its easternmost end the profile also reaches the Apenninic units thrust northward and perpendicular to the trace of the profile towards the Torino Hills in the Late Miocene.

## 3. ECORS-CROP profile (Fig. 2c)

This transect follows the trace of the geophysical campaign launched in 1985 by the French-Italian Étude Continentale et Océanique par Reflexion et Refraction Sismique - Progetto Strategico Crosta Profonda  (ECORS-CROP) geophysical campaign in the Western Alps involving high-resolution deep seismic sounding Roure et al. (1996). The geological interpretation of this profile is a modified version of the profile previously published by Schmid & Kissling (2000).

The external Alps exhibit 27km shortening, an amount very similar to that along the Pelvoux profile, calculated on the basis an area balance within an originally 17km thick upper crust, assuming decoupling at the interface with the lower European crust. This shortening is of Miocene age and postdates thrusting at the Penninic Front (Fügenschuh & Schmid 2003). Note that Butler (1983) postulated very substantially higher amounts of shortening (>50km) since his line balance assumes detachment along a floor thrust located only 1km below base Triassic. The 6.5 km $s^{-1}$ velocity contour beneath the external massifs, in the case of this profile the Belledonne Massif, indicates the presence of a steep tabular high-velocity body in the lower part of the upper crust that we again (see description Pelvoux profile) tentatively attribute to the East Variscan Shear Zone of Late Variscan age (Padovano et al. 2012). The velocities within the rest of the European upper crust are, apart from this anomaly, < 6.5 km $s^{-1}$ all the way back to the very internal parts of the Alps below the Gran Paradiso massif, built up the high pressure Briançonnais unit. This new finding about the P-wave velocity in the upper crust of the external part of the

Western Alps (Diehl et al. 2009) is incompatible with the presence of an allochthonous wedge of lower crust underneath the external Western Alps as postulated by Schmid & Kissling (2000) on the basis of gravity arguments (Bayer et al. 1989) and W-dipping seismic reflectors (Thouvenot et al. 1996) at the front of that putative wedge. Below the Gran Paradiso massif the 6.5 km $s^{-1}$ contour steeply raises to a depth of only some 10 km, across the Penninic Front and into the area of the high-pressure part of the Briançonnais units, indicating a high-pressure metamorphic overprint also at great depth.

At the surface the profile crosses the type locality of the top-WNW Penninic Front ("Roselend Thrust" of Loprieno et al. 2011). This thrust, together with a 10 km wide belt of units attributed to the Valaisan paleogeographic domain, containing, amongst other lithologies, black schists interlayered with mafic sills and low-T eclogites (Bousquet et al. 2002, 2012a; Loprieno et al. 2011) forms a planar band of strong reflectors that can be traced down to a depth of some 15 km along the ECORS-CROP reflection seismic profile (Thouvenot et al. 1996). Following Schmid & Kissling (2000), and in analogy with the Pelvoux profile discussed above, we kinematically link the Penninic Front with the base of the Adria mantle wedge.

The near-surface geometry of the upper crustal structures within the Briançonnais units, namely the non-metamorphic Zone Houllière and the High-Pressure Briançonnais units (Ruitor, Zona Interna and Gran Paradiso nappes), as well as the geometries of the Piedmont-Liguria Units (Zermatt-Saas and Combin Zones) are drawn following Schmid & Kissling (2000) and Bucher et al. (2003, 2004). They are characterized by large-scale post-nappe backthrusts and backfolds formed between 35 and 31 Ma, i.e. contemporaneous with thrusting along the Penninic Front. High-pressure metamorphism related to subduction was dated at 48-47 Ma in the

Briançonnais units, overprinted by greenschist facies conditions related to nappe stacking at around 43-39 Ma (Villa et al. 2014). Pressures associated with the high-pressure event increase eastward and reach eclogite facies conditions in parts of the Gran Paradiso internal massif, characterized by P-wave velocities around 6.0 to almost 7.0 km $s^{-1}$ at greater depth according to the velocity model of Diehl et al. (2009).

Compared to the Briançonnais units high-pressure metamorphism within the more internal Zermatt-Saas-Fee unit (Piedmont-Liguria Lower Plate of Fig. 2c) and within small continental slices at its top indicates an earlier onset of a long and protracted high P overprint (62-42 Ma) according to age dating (e.g. Dal Piaz et al. 2001; Skora et al. 2015; Weber et al. 2015; Fassmer et al. 2016). The most internal parts of the ECORS-CROP profile cross the Sesia Zone, which together with the Dent Blanche Klippe, was classically considered as an "Austroalpine" unit. This is because the protoliths forming the Sesia Zone are undoubtedly derived from the Adria microplate in a paleogeographical sense (Compagnoni et al. 1977). However, in a plate tectonic context, the Sesia Zone that reached eclogite facies conditions with up to 2 GPa pressures in the period 85-60 Ma (Regis et al. 2014) and consisting of thin crustal sheets (Giuntoli & Engi 2016) is interpreted as having been accreted as a part of the former Lower Plate to the Upper Plate Units (i.e., the Canavese, Ivrea Zone and Southern Alps) during Late Cretaceous to early Cenozoic orogeny in the Western Alps, associated with the subduction of the Piedmont-Liguria ocean (Babist et al. 2006, Handy et al. 2010). The situation is different in the Austroalpine nappes proper where high-pressure metamorphism is significantly older and related to Cretaceous intra-continental subduction that evolved within the future Austroalpine units that formed the upper plate during Cenozoic orogeny (Schmid et al. 2004). In summary, the Sesia Zone is regarded as a lower plate unit following Froitzheim et al. (1996; their Fig. 2) and Pleuger et al. (2007; their Fig. 14) who proposed that the Sesia Zone represents a continental fragment floating within the Piedmont-Liguria oceanic domain that rifted and drifted away from Adria during the opening of Alpine Tethys in the Mesozoic. Two important geological observations support this finding. Firstly, the southern edge of the mostly continental Sesia high-pressure unit, immediately adjacent to the Insubric Line, contains a high-pressure slice containing omphacite-bearing mantle peridotites, which are testimony of a former oceanic tract SE of the Sesia Zone that underwent high-pressure metamorphism (Pognante 1989, Barnes et al. 2014). Secondly, the non-metamorphic Canavese Zone, immediately south of the Insubric Line and north of the Ivrea Zone contains, amongst other lithologies, outcrops of exhumed mantle rocks overlain by Jurassic age radiolarites (Beltrando et al. 2015a), which points to the existence of an oceanic tract between the largely continental Sesia high-pressure unit and the Ivrea Zone. It is only the Canavese Zone, together with the Ivrea Zone and the Southern Alps that form the upper plate of the Alpine orogen in the internal parts of this transect. As demonstrated in the pioneering work by Elter et al. (1966) the Canavese Zone has to be considered the root of the highest nappe exposed in the Préalpes Romandes, the ophiolite-bearing nappe des Gets (see profile of Fig. 2d; unit labeled as Piedmont-Liguria Upper Plate). Both these units represent partly oceanic series that once formed the presently largely eroded upper plate of the Alpine edifice.

In contrast to the more southerly-located profiles, very thin slices of the Ivrea mantle wedge, accompanied by lower crustal lithologies of the Ivrea Zone, reach the surface of the earth, slices that are too thin to have a notable effect on the velocity contours. The main part of the Ivrea mantle wedge is confined to a depth >10km as indicated by the 7.0 km $s^{-1}$ contour line. Along this profile the 7.25 km $s^{-1}$ contour line reveals an inverted velocity gradient, allowing for localizing the base of the Ivrea mantle wedge above the subduction channel.

The Adria lower and upper crustal units are affected by backthrusts representing the easternmost extension of the Southern Alps. Details are drawn according to the interpretation given by Roure et al. (1990). This thrusting, detected in the subsurface of the Po plain in the course of the ECORS-CROP reflection seismic campaign, is of pre-Burdigalian (pre-20.5 Ma) age and affects part of the Gonfolite Lombarda sediments, providing an age range for thrusting of Late Oligocene to Early Miocene. Note that this profile does not reach the front of the N-directed Apenninic thrusts, which stays south of the SE end of this section.

## 4. NFP-20 West profile (Fig. 2d)

This transect follows the trace of a Swiss geophysical campaign (NFP 20 "Deep Structure of Switzerland" whose results were published in Pfiffner et al. (1997). Earlier geological interpretation of the deep structure revealed by controlled source reflection and refraction seismology such as those by Escher et al. (1997) and Schmid et al. (2004) have been revised in the light of the 3D velocity model of Diehl et al. (2009).

Along this profile the effects of the foreland propagation of Alpine shortening induced the decollement of the entire Molasse basin along Triassic evaporite horizons, which led to the formation of the Jura Fold-and-Thrust Belt. The Alpine thrust front jumped by some 100km to the north after about 12 Ma ago; this decollement is associated with some 26km of shortening and roots in the external massifs (Aiguille Rouge and M. Blanc massifs) of the Alps (Burkhard & Sommaruga 1998). However, the total amount of shortening affecting the upper European crust, which along this transect also includes the lowermost Penninic nappes behind the Penninic Front, amounts to some 50km according to an area balance along this profile, again assuming decoupling at the interface between upper and lower crust. This is almost twice as much as deduced for the ECORS-CROP profile but, in this case, also includes all the shortening that occurred since 35 Ma ago. The deep structure of the External Alps along this profile is characterized very marked undulations of the velocity contours between 7.5 and 6.0 km $s^{-1}$. Some of these undulations may again have been caused by velocity gradients related to pre-Alpine structures. However, one of these undulations, an antiformal structure located underneath the external massifs, appears to also affect the Moho; the closely spaced 7.5, 7.0 and 6.5 km $s^{-1}$ velocity contours all consistently indicate an antiformal structure affecting the Moho and lower crust. A local up doming of the Moho by about 5km correlates well with earlier observations in reflectivity pattern along the NFP20West near-vertical reflection profile (Pfiffner et al. 1997) that at the time were not further interpreted. As this Moho geometry has now been consistently observed by two independent seismic data sets, we assume this antiformal Moho-structure to be real and possibly representing a pre-Alpine feature.

For constructing the near-surface structures in the most external parts of the profile we mainly used Sommaruga (1997) for the Jura Mountains and Sommaruga et al. (2012) for the Molasse basin. In the case of the External Massifs, the Helvetic nappes and the Préalpes Romandes we based our interpretation of data provided by Burkhard (1988), Marchant (1993), Steck et al. (1997), Escher et al. (1997) and Burkhard & Sommaruga (1998).

The Penninic Front along this profile coincides with the south dipping Rhone-

Simplon line, a wide dextral shear zone exhibiting initially ductile deformation starting to be active at around 35 Ma (Steck & Hunziker 1994; Schmid & Kissling 2000), which graded into brittle deformation from the Miocene onwards (Cardello & Mancktelow 2015). The Simplon normal fault located immediately east of this transect (Mancktelow 1992; Grasemann & Mancktelow 1993) represents a tensile bridge that accommodates orogen-parallel extension during the backthrusting over the Ivrea mantle wedge towards ESE and associated top WNW thrusting at the Penninic Front of the Western Alps (Keller et al. 2006). The downward projection of the Rhone-Simplon line to some 15km follows that proposed by Escher et al. (1997). At a >15km depth any interpretation of the geometry of this fault and adjacent units is only loosely constrained by migrated reflection seismic data (Marchant 1993) and strike-parallel lateral projection of near surface structures. Our interpretation differs from that of Escher et al. (1997) at a depth > 35km because we conceive the Rhone-Simplon line as delimiting the northern border that acted as a lateral ramp for that part of the Alpine upper crust that moved towards WNW together with the Ivrea mantle wedge during its indentation.

The internal units of the Alps SE of the Rhone-Simplon line are characterized by spectacular backfolds within the Briançonnais units whose geometry down to a depth of about 20km is constrained by reflection seismology (Marchant 1993; Steck et al. 1997) and orogen-parallel axial projections of surface structures (Escher et al. 1997; Keller & Schmid 2001; Kramer 2002). At a depth >20km the P-wave velocity contours of Diehl et al. (2009) served as a guideline for constraining the geometry. The 6.0 km $s^{-1}$contour indicates a large-scale velocity inversion that follows the SE limb of the Monte Rosa backfold down to a depth of some 30km. There the 6.0 km $s^{-1}$ contour swings back into a region characterized by a normal velocity gradient and runs parallel to the 6.5 and 7.0 km $s^{-1}$ velocity contours that are typical for the European lower crust. As a consequence we limit the area characterized by such backfolding to the uppermost 35km of the crust.

Farther to the SE the 6.0 and 6.5 km $s^{-1}$ contours rise to a shallow depth of only some 10 km and 15km, respectively. This rise occurs within the Sesia Unit that is known to exhibit eclogite facies conditions in its most internal parts only (Bousquet et al. 2012a). Because this eclogite facies part of the Sesia Zone is immediately adjacent to the Ivrea mantle rocks, the limit between the Adria mantle wedge and the under-plated Sesia zone is ill constrained. The near-surface structures of the Southern Alps are drawn after Cassano et al. (1986).

## 5. Ticino profile (Fig. 2e)

The Ticino profile runs parallel to the reflection seismic lines C1, C2 and C3 of NFP 20 project (Pfiffner & Heitzmann 1997); the southernmost parts of this profile are located near lines S4 and S5 of the same project (Schumacher 1997). At depth, the profile crosses the NE-most part of the Ivrea mantle wedge, very near its termination (see Fig. 3). Note also that this profile crosses a pronounced along-strike axial culmination (the core of the Lepontine dome; Berger et al. 2005; Berger & Mercolli 2006; Steck et al. 2013) at a location where the distance between external massifs and Periadriatic line is near a minimum (Fig. 1), going hand in hand with a maximum amount of shortening and exhumation. Since this transect is located east of the Rhone-Simplon line, dextral transpression allowing for top WNW thrusting along the Penninic Front is taken up by the Periadriatic line only.

According to an area balance, again assuming decoupling at the interface between upper and lower crust, the total amount of shortening affecting the upper

European crust (including the Penninic nappes below the Valaisan and Briançonnais units) since around 35 Ma increases by as much as some 115 km when compared to the 50km determined along the NFP20-West traverse. Rosenberg & Kissling (2013) measured some 71 km shortening along a profile located between our NFP20-West and Ticino profiles (their Simplon profile). This indicates a steady increase in the amount of post-collisional (post 35 Ma) shortening within the accreted European lower plate towards northeast into the Ticino transect.

In the external part of the profile the 6.0 km $s^{-1}$ contour rises to shallow depth underneath the frontal part of the external massifs, hinting at the existence of a pre-Alpine structure (External Massifs Shear Zone of Guillot & Ménot 2009?) also along this profile. The near-surface features in the foreland of the Molasse Basin are drawn according to Sommaruga et al. (2012), the geometry of the external Aar and Gastern massifs is modified after Müller (1938) and Pfiffner (2011), the structures in the Helvetic nappes, overlying Penninic nappes and the Gotthard nappe (formerly considered a "massif") are drawn after Menkveld (1995) and Pfiffner et al. (2010).

The internal parts of the transect, depicting the nappe pile within and above the Lepontine dome, are characterized by a large-scale backfold associated with the Northern Steep Belt located within the southern border of the Subpenninic Gotthard nappe (Schmid et al. 2004; Berger et al. 2005), separated by a flat-lying nappe pile from the Southern Steep Belt as defined by Milnes (1974), who was first in realizing the importance of post-nappe recumbent mega-folds affecting a previously existing nappe stack consisting of Briançonnais, Penninic and Subpenninic units. The formation of the Southern Steep Belt, the southern limb of a second more southerly located backfold, is closely associated with backthrusting of the Alpine nappe stack over the Ivrea mantle wedge and overlying crust in the area of the Southern Alps and along the Periadriatic Line that initiated between 35 and 30 Ma in a dextrally transpressive scenario (Schmid et al.1987 & 1989; Steck & Hunziker 1994). Note that Miocene thrusting in the Southern Alps is younger, i.e. post-20Ma (Schönborn 1992). Based on the offset of the Alpine grade of metamorphism across the Periadriatic Line, the vertical component of dextral transpression amounts to some 20km along the trace of this profile. The amount of the simultaneously occurring horizontal displacement across the transpressive Periadriatic Line remains unknown due to the lack of markers; estimates vary between 100km (Schmid & Kissling 2000) and 240km (Handy et al. 2010). Details concerning the shallower parts of this central portion of the profile are drawn based on structural information by Milnes (1974), Huber et al. (1989), Leu (1987), Berger et al. (2005), Steck et al. (2013) and own observations. Reflection seismic data along line C1 and C2 (Pfiffner & Heitzmann 1997) were used for placing nappe contacts at the base of the Antigorio nappe and the Gotthard nappe, respectively, and for estimating the depth of the top of the European lower crust.

At a depth >20km the NW edge of the Ivrea mantle wedge is defined by the 7.0 km $s^{-1}$ contour. At a depth of <20km the Periadriatic fault, whose dip is evaluated on the basis of surface observations and a series of reflection seismic profiles (Schmid & Kissling 2000; their Fig. 5) delimits the NW edge of the Ivrea mantle wedge, including the Ivrea Zone that represents the adjacent Adriatic lower crust. The Ivrea Zone, including some ultramafic bodies at its northern rim, has been rotated along a near-horizontal axis into a vertical orientation due to Alpine deformation under low-grade metamorphic conditions (Handy 1987; Schmid et al. 1987; Zingg et al. 1990). The Ivrea Zone is well known for representing a map-view section across lower continental crust (Fountain 1976). It is likely that the topmost parts of the Ivrea Zone and mantle became highly tectonized under semi-brittle conditions, as is schematically shown in Fig. 2e, allowing for this rotation.

The 6.5 km $s^{-1}$ contour rises to a shallow depth of around 10 km above the Ivrea mantle wedge also in this section, crossing from the European lower crust across the subduction channel over the presumed top of the Ivrea mantle wedge. The northern limit of the Ivrea mantle wedge and the adjacent Ivrea Zone are constrained by downward extrapolation of the Periadriatic Line to a depth of 20km; at greater depth the Periadriatic line delimits the northern limit of the Ivrea mantle wedge.

The profile across the upper crust of the Southern Alps is partly based on the geological interpretation of reflection seismic data gathered during the NFP-20 and CROP campaigns (Schumacher 1997; Schumacher et al. 1997; lines C3-south, S4, S5, S6, S7& CROP 88-1) and partly on the subsurface information provided by Cassano et al. (1986). Furthermore, we based our interpretation also on the work of Bertotti (1991) and Schönborn (1992). The total amount of shortening within the Southalpine retro-belt amounts to 74km according to section balancing in the area crossed by this transect (Schumacher et al. 1997). A large, although not well constrained part (between 24km and 43km along a section east of Lake Como according to Schönborn 1992) of this shortening pre-dates the Late Eocene. This leaves between 31 and 50 km of shortening that occurred in the Middle to Late Miocene. The lower bound of this shortening estimate is more realistic in view of the overall geometry of the transect depicted in Fig. 2e. The reason for the upward excursion of the 6.5 km $s^{-1}$ contour inside the Adriatic upper crust remains unknown; possibly it is related to heterogeneities formed during Variscan tectonics.

## 6. Horizontal section (Fig. 3)

To discuss the overall shape of the Ivrea mantle wedge and lateral variations in its geometry the five profiles described above are complemented with a horizontal section across the 3-D P-wave tomographic model of Diehl et al. (2009) at a depth of 20km (Figure 3). This horizontal section shows (1) a curved area of positive P-wave anomalies >10% nicely mimicking the well-known positive Bouguer anomaly associated with the Ivrea mantle wedge (Kissling 1984; Serva 2005), (2) an extension of the Ivrea mantle wedge across the Periadriatic Fault visible north of Locarno (in reality the wedge underlies the N-dipping Periadriatic Fault, a backthrust in its hangingwall see Fig. 2e), and, (3) a possible dextral offset of the central axis of the Ivrea anomaly between profiles ECORS-CROP and NFP-20 West that leaves no trace at the surface of the earth when consulting the geological maps of the area.

Interestingly, the central axis of the high-velocity area is located east of the Periadriatic Fault in the north and gradually changes over to lying west of the Periadriatic line in the south. This indicates increasing amounts of underthrusting of the mantle wedge beneath the Internal Western Alps towards south, culminating in what is seen in the Argentera section (Fig. 2a): a Periadriatic Fault that acted as a huge backthrust and severe modification of the suture between Europe and Adria (lower plate Piedmont-Liguria ophiolites) in the form of a huge backfold with the high-pressure Briançonnais Units in its core (Fig. 2a).

## C. Multi-stage evolution of the arc of the Western Alps

The following discussion on the evolution of the arc of the Western Alps integrates the findings described above with a review of existing knowledge on the multi-stage evolution of the Western Alps concerning kinematic evolution and time constraints derived from field data. In view of the considerable difficulties in exactly dating metamorphic events using isotopic ages, particularly in the case of HP metamorphism (see review by Berger & Bousquet (2008), we also heavily rely on biostratigraphical constraints. We also present tectonic reconstructions using Gplates reconstruction software (Boyden et al. 2011) that have already been applied to parts of the Western Mediterranean and eastern realm adjacent of the Alps-Apennines orogenic system (van Hinsbergen & Schmid 2012; van Hinsbergen et al 2014a; Advokaat et al. 2014).

## 1. Exhumation of Ivrea Zone and underlying lithospheric mantle in the Mesozoic

The Ivrea Zone is well known for exposing a map-view section across lower continental crust of the Southern Alps that formerly was part of the Adriatic microcontinent (e.g. Fountain 1976, Handy 1987, Handy & Zingg 1991). It includes a series of small peridotite lenses at its northern rim, due to later tilting in the context of Alpine orogeny. Restoration of the pre-Alpine setting via backrotation of Ivrea Zone and adjacent parts of the Southern Alps adjacent to the Periadriatic Line reveals that the Ivrea Zone was part of the distal passive continental margin that formed in connection with Tethyan rifting culminating at around 180-170 Ma ago (e.g. Bertotti et al. 1993). The Ivrea Zone and underlying Ivrea mantle wedge were located near the western margin of the distal continental margin of the Adria microplate that had a strike (Weissert & Bernoulli 1985) almost parallel to the present-day NNE-SSW strike of the Ivrea mantle wedge (Fig. 3). Based on a combination of P-T estimates of pre-Alpine parageneses and radiometric dating it was inferred that crustal thickness was locally reduced to 10km or less in the area of the Ivrea Zone and adjacent upper crustal units (Handy et al. 1999). Hence the present-day slope of the Moho, rising from some 40km to about 20km revealed by the Moho-map of Spada et al. (2013), and, rising to some 15km in Figs. 2c&d, is interpreted as inherited from passive margin formation in Mesozoic times. The Canavese area, coinciding with the ocean-continent boundary, from which the Sesia extensional allochthon separated during rifting (Froitzheim et al. 1996; Beltrando et al. 2015a,b), is only some 50km away from the Ivrea Zone.

The shallow depth of the top of the Ivrea mantle wedge inherited from Mesozoic rifting has important consequences regarding rheology and strength of the Ivrea mantle wedge during Alpine orogeny. The brittle to crystal plastic transition of peridotite is estimated to be at around 600° according to most workers (e.g. Druiventak et al. 2011). Temperatures in the depth range of the Ivrea mantle wedge (15-40km) are between 400° and 650°C in the central area of the Alpine orogen according to the steady state kinematic model of Bousquet et al. (1997). Hence, most of the Ivrea mantle wedge is expected to be controlled by frictional strength around 1000 MPa at 15 km depth in a compressional regime under dry conditions, reaching excessive values at greater depth (Byerlee 1978). At its front the Ivrea mantle wedge is surrounded by upper crustal nappes (Fig. 2) whose rheology is likely to be controlled by viscous creep of large quartz-bearing volumes of rock. Extrapolation of flow stress data to geological strain rates, combined with paleopiezometry using recrystallized grain size, predict flow stresses between of 10-100 MPa within the same 400° and 650°C temperature range (Stipp et al. 2002). This indicates that the strength of the Ivrea mantle wedge exceeds that of quartz-bearing units at its front by at least one order of magnitude. This suggests that indentation of the Ivrea mantle wedge into the Alpine nappe stack probably played an important role in the formation of the arc of the Western Alps, as will be further discussed below.

## 2. Top-NNW thrusting during sinistral transpression (Stage 1, pre-35 Ma)

The orientation of stretching lineations in combination with associated sense of shear criteria have been applied to deduce tectonic transport directions in highly sheared rocks from the Alps (e.g. Malavieille et al. 1984; Platt et al. 1989). Along the arc of the Western Alps the orientation of the lineations is dominantly transverse to the strike of the orogen. The inferred tectonic transport smoothly changes from top NW in the area of Monte Rosa to top WSW in the area of the Dora Maira area (Malavieille et al. 1984). This apparently simple arrangement, if taken to indicate sense of movement, leads to obvious difficulties in geometrical and plate kinematic terms (Platt et al. 1989). The major difficulties in interpreting kinematic data based on stretching lineation analysis is that (1) stretching lineations may indicate the orientations of finite strain (e.g. Ramsay 1980) rather than shearing during a particular stage of a complex orogenic evolution and (2) that orientations of older lineations may have been reoriented by younger deformations (e.g. Walcott and White, 1998; Keller & Schmid 2001; van Hinsbergen and Schmid, 2012). Hence, only careful local studies addressing the polyphase structural evolution of particular areas can reveal a more complex kinematic evolution. The results of such studies are compiled in Fig. 4.

A first generation of lineations indicating top north transport is best documented in areas of the Western Alps located behind, i.e. east, of the Penninic Front (areas 1, 2 and 3 in Fig. 4 and references therein). In these areas the orientations of this first generation of inferred transport directions is related to pre-35Ma nappe stacking. According to detailed analyses of structural superposition this first generation of lineations pre-dates thrusting along the Penninic Front initiating at around 35 Ma and is hence is of Eocene or even earlier age (e.g. Bucher et al. 2003; Ceriani & Schmid 2004; Loprieno et al. 2011). However the orientations of these first generation lineations scatter considerably between top-NW and top-NE due to later re-orientation. In the more internal parts of the internal Western Alps the preservation potential of the orientation of such early stretching lineations is low. In most places they are not discernable any more because of intense shearing during subsequent deformation during stage 2 (see below). In the more internal units of the Aosta Valley, for example, they are seen to scatter too much for inferring transport directions of this earliest phase (Bucher et al 2004). Only in some parts of the eclogitic lower plate Piedmont-Liguria units is a former top NNW eclogite facies lineation preserved (area 5 in Fig. 4; Philippot 1988, 1990). Interestingly, also parts of the External Western Alps (Pelvoux massif and its cover) are known to have been affected by stage 1 to N to NW-directed deformation of pre-Priabonian age (Dumont et al. 2011).

Large-scale structural considerations (Schmid & Kissling 2000; Handy et al. 2010) and plate kinematic constraints (Fig. 5) indicate that thrusting and plate convergence were top N to NNW before 35 Ma. This implies a scenario of sinistral transpression in the Western Alps during the early stages of orogeny when the upper plate tectonic units below which the Alpine nappe stack formed (Adria lithosphere; oceanic in case of the Western Alps, continental in case of the Eastern Alps) moved towards N to NNW, presumably past that part of the European plate presently located in SE France, as first proposed by Ricou & Siddans (1986). Pre-35Ma plate convergence was associated with nappe stacking and early stages of exhumation of parts of nappes buried to high-P (e.g. Schmid et al. 1996, their Fig. 8). A point located at the northern tip of Adria (Monti Lessini) moved some 260km to the NNW in respect to stable Europe between 55 and 35 Ma go (1.3 cm/y) according to the reconstruction shown in Fig. 5d-f, which assumes that Adria was part of Africa at this time, and used the Europe-Africa plate circuit of Seton et al. (2012). This is more than the 195km (0.98 cm/y) of shortening calculated by Schmid & Kissling (2000) for the same time interval but considerably less than the estimate of Handy et al. (2010) who inferred 465km shortening for the interval 35-67Ma (1.45cm/y), both these estimates being solely based on interpretations of the geological structure and metamorphic history of the Alps. According to Handy et al. (2010) top-N thrusting started as early as some 84Ma ago, consistent with the plate circuit estimates for the onset of convergence based on Seton et al. (2012), and first led to high-pressure metamorphism in the Sesia Zone (Manzotti et al. 2014), predating high-pressure metamorphism within the lower plate units derived from the Piedmont-Liguria Ocean that started at around 60Ma ago (Berger & Bousquet 2008).

In summary, peak pressures within the Western Alps were reached in a scenario of sinistral transpression along an original approximately N-S striking segment of the western Alps (Fig. 5d&e) during this first stage. Most of the exhumation of the eclogites to moderate crustal levels very probably also took place during this first stage of the evolution of the arc of the Western Alps.

## 3. Top-WNW thrusting and WNW-directed indentation of the Ivrea mantle slice (Stage 2, 35-25 Ma)

A second population of stretching lineations is associated with deformation that post-dates stage 1 top-NNW thrusting according to structural overprinting criteria. The top-WNW lineations formed during this stage can reliably be attributed directly to thrusting along the Penninic Front in the case of locations 1,3,4 and 7 in Fig. 4 (references cited therein). There the inferred thrusting directions vary between azimuth 260° and 345° with a mean at around azimuth 300°. The measurements at locations 2 and 5 were taken in more internal positions with respect to the Penninic Front while those at location 6 are from the footwall of the Penninic Front (lineations in the “schistes à bloc formation” immediately below the base of the Embrunais-Ubaye nappe stack; Trullenque 2005). Given the similar spread in orientations shearing during stage 2 top-WNW thrusting is likely also for locations 2, 5 and 6.

Several independent sets of data provide dating constraints. $^{40}Ar$-$^{39}Ar$ dating of phengites that formed syn-kinematically with shearing of basement rocks of the Pelvoux massif in shear zones along the Penninic Front or in its immediate footwall vary between 35 and 30 Ma with one younger age at 27Ma (Simon-Labric et al. 2009; Dumont et al. 2012). This dating confirms earlier evidence for the onset of top WNW thrusting at around 32 Ma based on fission track dating (Fügenschuh & Schmid 2003). Since the Penninic Front is associated with dextral shearing along the Rhone-Simplon line (Fig. 1) and dextral shearing in the Insubric mylonites (part of PL of Fig. 1 around Locarno) the dating of such dextral shearing provides further timing constraints. The “ductile Simplon shear zone” of Steck et al. (2015), predating late stages of normal faulting at the Simplon normal fault (Mancktelow 1992), was active between 34 and 25 Ma ago as inferred from the ages of the mantle-derived andesite and tonalite (32–29 Ma) and synkinematic crustal aplite and pegmatite intrusions in the southern steep belt between Domodossola and Locarno (Steck et al.

2013). Dextral shearing in the mylonite belt of the Insubric line, a segment of the Periadriatic Line (PL in Fig.1) initiated at around 32 Ma ago based on radiometric age data from the eastern margin of the Bergell pluton (32-30 Ma; von Blanckenburg, 1992). The deep-seated Bergell intrusion including its subsequent fast exhumation is closely related to a combination of backthrusting and dextral strike-slip movements accommodated by the greenschist mylonites of the Insubric line (Schmid et al., 1989; Berger et al., 1996). The top-WNW senses of shear in the “schistes à bloc” formation (Kerckhove 1969) at location 6 of Fig. 4, affected by top-WNW shearing during a first phase of deformation (Trullenque 2005), provides a stratigraphic age constraint; this chaotic mélange overlies the Priabonian-age flysch de Champsaur (Bürgisser & Ford 1998), and hence, its age of formation is considered to be latest Priabonian to earliest Oligocene (around 34 Ma). A similar age constraint is provided for the sinistral senses of shear observed at the western termination of the Argentera-Cuneo Line at location 7 of Fig. 4 that is kinematically connected with the Penninic Front (Trullenque 2005); these mylonites are found in immediate tectonic contact with the grès d’Annot of Late Eocene to Early Oligocene age that terminate the evolution of the Eocene Northern Alpine foreland basin in the southern Subalpine chains (Ford & Lickorish 2004).

The end of stage 2 shearing is not equally well constrained but certainly pre-dates the activity of the left-lateral Giudicarie transpressive belt starting to offset the Periadriatic line at around 23-21 Ma ago (Scharf et al. 2013). An end of stage 2 at around 25 Ma ago is compatible with the onset of the Paleo-Apenninic phase (Schumacher & Laubscher 1996) at about this time. Stage 2 certainly pre-dates the onset of rapid counterclockwise rotation of the Corsica-Sardinia block at 21 Ma ago (Gattacecca et al. 2007) that we link to stage 3 oroclinal bending discussed in the next chapter.

At the large scale thrusting at the Penninic Front is interpreted to be kinematically connected to dextral strike slip along the Rhone Simplon and Periadriatic Lines on the one hand, and to sinistral strike slip along the Argentera-Cuneo Line on the other hand. The orogen-perpendicular lineations in the Western Alps compiled by Malavieille et al. (1984) are probably related to stage 2 top-WNW shearing and associated backfolding that is pervasive within entire volume of rocks between Penninic Front and the Periadriatic Line between Ivrea and Cuneo; this shearing formed new lineations and/or reoriented older ones (Fig. 1). It is proposed that this pervasive shearing, linked to WNW-directed indentation of the Ivrea mantle wedge, first proposed by Laubscher (1971), also took place during stage 2 (35 and 25 Ma). The amount of top WNW displacement along the Penninic Front is not well constrained. Schmid & Kissling (2000) estimated WNW-directed displacement of the Ivrea Zone to amount to around 100km, (Handy et al. 2010) proposed some 200km. The reconstruction in Fig. 5c yields 93km and 150km for the SW and NE ends of the Ivrea mantle wedge, respectively.

The reconstruction shown in Fig. 5c depicts displacement vectors that diverge between Western Alps and Eastern Alps and hence are bound to induce orogen-parallel extension in the Alps that is known to also have started at around 35 Ma ago (Pleuger et al. 2008; Beltrando et al. 2010). The backfolding and backthrusting, which is so typical for the Western Alps (see profiles of Fig. 2 and accompanying text) and which affected a pre-existing nappe pile, also occurred during stage 2 (e.g. Bucher et al. 2003). Hence, it is also associated with the indentation of the Ivrea mantle wedge that underthrusted much of the internal zones of the Alps (see Fig 2a-c). The spectacular backthrust over the Cenozoic cover of the Tertiary Piedmont

Basin shown at the eastern end the profiles shown in Fig. 2 a & b also formed during stage 2.

## **4. Oroclinal bending and top SW thrusting in the southern part of the Western Alps (Stage 3, 25-0 Ma)**

The external Alps south of the Pelvoux massif are well known for late-stage top-SW thrusting (e.g. Fry 1989, Lickorish & Ford 1998, Bürgisser & Ford 1998) that is not seen farther to the north along the Western Alpine arc. By studying overprinting relationships south of the Pelvoux massif (area at location 4 in Fig. 4) Trullenque (2005) discovered that this top-SW thrusting actually post-dates stage 2 top WNW thrusting. Lickorish & Ford (1998) have shown that 10.5 km out of a total of around 21 km shortening, associated with minor thrusting rooting in the basement of the Argentera Massif, occurred in Late Miocene to Pliocene times. Ritz (1992), based on a paleostress analysis in the Digne, Castellane and Nice thrust systems, found that during Middle Miocene to recent times the $\sigma_1$ trajectories varied from NE-SW in the Digne area, gradually turning into a N-S orientation in the Castellane arc and eventually to a NW-SE direction in the case of the Nice thrust sheet. This same radial arrangement was found for the p-axes derived from stress inversion of earthquake focal mechanism data by Delacou et al. (2004, their Fig. 5). This radial pattern is very likely related to rotation around a pivot located east of the Argentera massif in the area of the "Ligurian knot" (Laubscher et al. 1992), i.e. in an area where the 3 Moho's beneath Alps, Apennines and northern Tyrrhenian Sea meet (Spada et al. 2013).

Three largely independent lines of evidence support the idea that oroclinal bending in the southernmost Western Alps is to be seen in connection with the structural observations mentioned above. Firstly, Collombet et al. (2002) reported counterclockwise rotations about a subvertical axis in respect to stable Europe that increase from 68° in the Ubaye region (area between profiles 1 and 2 of Figure 1) to 117° in the Ligurian Alps SE of Cuneo (Fig. 1). The data were collected in sedimentary rocks of the Briançonnais tectonic unit. The age constrains for magnetization are rather loose; the authors propose that the rocks were remagnetized after the backthrusting phase (our stage 2) and therefore to be of Late Oligocene age. Secondly, all the major tectonic elements of the internal Alps derived from the Briançonnais and Piedmont-Liguria paleogeographical domains can be followed into western Liguria and all the way to Genova and slightly beyond (Fig. 1). In the Ligurian Alps all the pre-Oligocene Alpine structures, sealed by the Tertiary Piedmont Basin, are top-S in present-day coordinates. Since the counterclockwise rotation of stable Adria only amounts to 10±10° (van Hinsbergen et al., 2014b) or 20° (Marton et al. 2011) at least a part of the 49° differential rotation of Liguria in respect to the Ubaye region (117°-68°= 49°), namely 39±10°, is probably related to oroclinal bending during stage 3. It is likely that oroclinal bending during stage 3 was partially driven by eastward trench retreat of the Apennines slab. However, the rotation in the Ubaye region itself, located north of the Argentera-Cuneo Line is probably also due to stage 2 indentation of Adria. Thirdly, a 50° counterclockwise rotation phase of the Corsica-Sardinia block occurred between 20.5 and 16 Ma ago (Gattacceca et al. 2007) around a pivot near the N tip of Corsica, as Corsica and Sardinia drifted away from their pre-Miocene location south of southern France (Advokaat et al. 2014); this virtually demands substantial counterclockwise rotation in Liguria associated with oroclinal bending. Note that such counterclockwise rotation is directly demonstrated by the ~45-50° anticlockwise rotation of mostly Oligocene sediments of the Tertiary Piedmont Basin sealing pre-Oligocene Alpine structures evidenced by paleomagnetic data (Maffione et al. 2008).

Oroclinal bending in the southernmost Western Alps during stage 3 is evidently associated with orogeny in the Apennines. Slab rollback of the Adria plate towards N in the northernmost Apennines is the underlying driving force for both oroclinal bending and the rotation of Corsica and Sardinia. The discussion of the Alps-Apennines transition, discussed in the next chapter, will shed more light to the Miocene history of the Ligurian Alps that cannot be understood without discussing orogeny in the Apennines and its spatial and temporal relation to orogeny in the Alps.

## D. The Alps-Apennines transition in space and time

### 1. Tertiary Piedmont Basin and Epiligurian basins

Late Eocene to Neogene sedimentary basins found in the interior of the Western Alpine arc, the Tertiary Piedmont Basin, and similar basins found on top of the External Ligurides of the Northern Apennines, referred to as "Epiligurian" basins, play a crucial role for understanding the spatial and temporal transitions between the west- and north-verging Alps and the east- and north-verging Apennines (Fig. 6).

The southern limit of the outcropping pre-Pliocene parts of the Tertiary Piedmont Basin (Fig. 1) transgresses the high-pressure lower plate units of the Ligurian Alps (i.e. the high pressure Briançonnais and Piedmont-Liguria units), as well as adjacent upper plate units of the Alps derived from the Piedmont-Liguria Ocean found immediately east of Genova (i.e. the Antola nappe). As first pointed by Elter & Pertusati (1973) the Antola nappe, predominantly consisting of Helminthoid

Flysch, is best considered as a former part of the Alps; they wrote: “La falda del Monte Antola e le altre formazioni liguri comprese nel triangolo Genova-Val Staffora-Levanto sono da considerarsi appartenenti alle Alpi in quanta sono saldate a queste dai terreni trasgressivi del Bacino terziario Piemontese fin dall'Eocene”). While earlier works suggested that this basin initiated by subsidence in a scenario of crustal stretching (Mutti et al. 1995) there seems to be some consensus in more recent publications that the Oligocene to Lower Miocene (Aquitanian) sediments of this basin were laid down in a compressive to transpressive setting (e.g. Carrapa & Garcia-Castellanos 2005; Mosca et al. 2010).

As shown in the profiles of Fig. 2 a & b the westernmost parts of the Tertiary Piedmont Basin were installed onto the upper crust of the Adria plate, which in turn overlies the Ivrea mantle wedge. It formed the foredeep of the backthrusted and -folded Western Alps during the Oligocene, which later became buried under post-Messinian cover. Such syn-sedimentary backthrusting to the east to southeast is seen in the seismic sections of lines 3 and 4 of Mosca et al. (2010, their Fig. 4). Thrusting is sealed by the Burdigalian unconformity that is ubiquitous in seismic sections of the subsurface of the Po plain (e.g. Schumacher & Laubscher 1996 and references therein). Such pre-Burdigalian thrusting is also seen along the front of the Southern Alps in western Lombardy where it affects sediments of Oligocene to Lower Miocene age, the Gonfolite Lombarda Group *sensu lato* (e.g. Bernoulli et al. 1989, 1993) that are part of the Southalpine foredeep (Garzanti & Malusa 2008) and laterally connect with the Tertiary Piedmont Basin (Mosca et al. 2010). Subsidence in the Piemonte region continued in Burdigalian and younger times, when this basin started to become involved in N-directed thrusting in connection with the Apenninic orogeny. The Torino and northern Monferrato hills immediately east of Torino (Fig.

1) are also considered as a part of the Tertiary Piedmont Basin by most authors although they would better be considered a part of the Epiligurian basins (see below). They morphologically mark the northern front of the Apenninic orogen that formed in Late Miocene to Pliocene times. These two northernmost parts of the Tertiary Piedmont Basin transgressed an “Alpine-type” basement in case of the Torino hills (Piana 2000) and a part of the External Ligurides in the case of the northern Monferrato hills (Laubscher 1971; Mosca et al. 2010; Festa & Codegone 2013). Together with the rest of the Tertiary Piedmont Basin that transgressed Alpine structures they became thrusted northwards as wedge-top basins, much like the so called Epiligurian basins of the External Ligurides, described below, along-strike farther to the east and southeast.

The term Epiligurian basins is used for wedge-top basins overlying the External Ligurides of the northern Apennines (Fig. 6) whose age-range and sedimentary characteristics are similar to those of the Tertiary Piedmont Basin (Molli et al. 2010 and references therein). The sedimentary sequence starts with the deposition of the deep marine Middle to Upper Eocene Monte Piano marls above a regional unconformity that indicates some pre-Apenninic deformation of parts of the External Ligurides that represent a fossil ocean-continent transition (Marroni et al. 2001) within the Adria plate that was part of the upper plate of the Alpine orogen. The Early Oligocene Ranzano formation was laid down above a second unconformity and received detritus from the metamorphic units of the Central Alps (Lepontine dome) that started to become exhumed north of the Periadriatic line (Garzanti & Malusa 2008). In the Early Oligocene the External Ligurides were part of the Adriatic foredeep in respect to the Southern Alps that formed due to “retroshearing” (Beaumont et al 1994) within the Alpine orogen (Schmid et al. 1996). Subsidence and siliciclastic sedimentation in the Epiligurian basins continued during the Late Oligocene when these basins and the underlying units became involved in the Apenninic orogeny during its earliest stages. From ca. 23 Ma onward the Epiligurian basins evolved into wedge-top basins as the External Ligurides became underplated by the accretion of Subligurian and Tuscanide units within the evolving Apenninic orogen (Molli et al. 2010 and references therein). Very recently Piazza et al. (2016) provided evidence for intense syn-depositional sinistral transpression during Oligocene to Early Miocene times sealed by a Burdigalian unconformity within one of the Epiligurian basins. This sinistral transpression is contemporaneous with sinistral strike slip along the E-W striking Villalvernia-Varzi line (Elter & Pertusati 1973) that defines the boundary between the Antola nappe often considered as a part of the Alps in the south and the External Ligurides considered as a part of the Apennines in the north (stippled line in Fig. 1 north of Genova). This sinistral strike slip movement was active in Late Oligocene to Early Miocene times as well (Di Giulio & Galabati 1995; Schumacher & Laubscher 1996). Laubscher (1991) considered the Villalvernia-Varzi line as accommodating the WNW-directed indentation of the Ivrea mantle wedge previously discussed (stage 2, during the evolution of the arc of the Western Alps). Although we consider the direct link between the sinistral Argentera-Cuneo line and the Villalvernia-Varzi line proposed by Laubscher (1991) as unlikely (see Figs. 1 & 6) sinistral strike slip motion along the Villalvernia-Varzi line and sinistral transpression in the Epiligurian basins were probably kinematically connected with stage 2 WNW-directed indentation of the Ivrea mantle wedge during the Oligocene.

In summary, the Tertiary Piedmont Basin seals very substantial pre-Late Eocene Alpine structures. The Epiligurian basins are of a similar type and age but they were deposited above an angular unconformity onto a unit that never underwent substantial Alpine deformation, namely the External Ligurides of the Northern Apennines (Fig. 6). Later, i.e. from about 23 Ma onwards, the latter represented wedge-top basins riding on the External Ligurides, which thrusted over more external units of the Northern Apennines towards NE (Molli 2008; Molli et al. 2010, and references therein). In essence, the stratigraphic base of these basins post-dates the pre-Late Eocene top-NNW thrusting during stage 1 sinistral transpression in the Western Alps described earlier while it pre-dates the onset of top-NE thrusting during the Apenninic orogeny, which is contemporaneous with stage 3 oroclinal bending and top SW thrusting in the southern part of the Western Alps. It follows from this that substantial parts of these basins probably represented syn-orogenic deposits during stage 2 top-WNW thrusting and indentation of the Ivrea mantle wedge into the Western Alps at 35-25 Ma ago.

## 2. The Alps-Apennines transition in map view

An interpretative map sketch of the Alps-Apennines transition is presented in Fig. 6. Considerable parts of this area are either buried by deposits of the Tertiary Piedmont Basin and very young Cenozoic cover of the Po plain or by the waters of the Ligurian and North Tyrrhenian Sea; hence this compilation in Fig. 6 remains speculative in many respects.

The continuation of the undeformed or little deformed Alpine foreland has to be located in Variscan Corsica that became thrusted by Alpine allochthonous units until the Late Eocene (Malavieille et al. 1998; Molli 2008). Note, however, that the present-day NW-SE strike of the Late Eocene Alpine front across the Ligurian Sea and Corsica restores to a N-S strike after the retro-deformation of 47° post-Eocene rotation of the Ligurian Alps (Maffione et al. 2008; see Figs. 5d&e) and 50° rotation of Corsica-Sardinia (Gattacceca et al. 2007). The Alpine allochthons in Corsica consist of, from bottom to top, the high-pressure units of the Tenda "massif" representing formerly subducted and subsequently exhumed parts of Variscan Corsica that remains devoid of an Alpine overprinted in the west, the high-pressure units of the Piedmont-Liguria Ocean and finally overlying non-metamorphic ophiolitic and continental nappes (Balagne and Nebbio nappes). The two lower units clearly represent lower plate units of the Alps that have to be correlated with the Briançonnais and Piedmont-Liguria units of the Western Alps, respectively. Although the origin of the uppermost non-metamorphic units is controversial (see Marroni & Pandolfi 2003 and references therein) we follow Malavieille et al. (1998) who argue for an origin of these highest units from the ocean-continent transition at the eastern margin of the Piedmont-Liguria Ocean. This makes them comparable to the non-metamorphic upper plate units of the Western Alps mapped in Figs. 1 & 6 (Préalpes Romandes, Embrunais-Ubaye and Western Liguria Helminthoid flysch). Oligocene-Miocene extension and transtension localized in the Gulf of Lion and the Ligurian Sea strongly affected the entire Corsican belt (Fig. 6), reducing its crustal thickness to normal thickness. In map view the width of the Alpine belt in Corsica is drastically reduced in comparison with the Western Alpine orogenic belt. We interpret this reduction in width to be partly related to the switch in subduction polarity between Alps and Apennines that led to a new west-dipping "Apenninic" subduction zone that must have evolved offshore eastern Corsica since latest Eocene or earliest Oligocene times (Fig. 6; Doglioni et al. 1998; Carminati et al. 2004; Molli & Malavieille 2011).

This new plate boundary was located east of the Corsica basin (Mauffret et al.

1999; see also CROP-03 and M-12A seismic sections in Finetti et al. 2001 and Finetti 2005). It evolved after cessation of Alpine, southeastward subduction in the Oligocene, and was followed by the two-stage opening of the Gulf of Lion back-arc basin. In a first, Oligocene stage, at around 30 Ma Corsica-Sardinia rifted off the Provence margin of southern France (Séranne, 1999). This was followed by the 21-16 Ma Corsica-Sardinia rotation around a pole located somewhere between the northern tip of Corsica and the Ligurian coast (Advokaat et al. 2014). The new plate boundary probably had a ~N-S-direction, slightly oblique to the NNW-SSE present-day strike of the Alpine orogen in Corsica (Fig. 6). In respect to the Apenninic orogeny both lower and upper plate units of Alpine Corsica occupied an upper plate position, remaining largely unaffected by Apenninic shortening, while easterly and northerly adjacent parts of the former Alps were heavily reworked when backthrusted onto the Adria continental margin and subsequently stretched as the Adria plate below the North Apennines rapidly rolled back eastward (Grandjacquet & Haccard 1977; Doglioni et al 1998; Rosenbaum & Lister 2004; Spakman & Wortel 2004; Marroni et al. 2010).

The northern margin of the lower plate metamorphic core of the Ligurian Alps (AFSZ in Fig. 1) is not exposed being buried beneath the Cenozoic cover of the Tertiary Piedmont Basin and younger cover (Fig.1). In Fig. 6 it is mapped according to subsurface seismic data (Mosca et al. 2010). Along a transect located west of Genova (profile A in Mosca et al. 2010) the metamorphic units of the Ligurian Alps were thrusted northward by about 50 km over the Adria continental autochthonous during the Apenninic orogeny in Miocene-Pliocene times. This implies that the western continuation of the older (pre-25Ma) Argentera-Cuneo Line, delimiting the southern boundary of the indenting Ivrea mantle wedge and overlying internal

Western nappe stack during the stage 2 top-WNW thrusting in the arc of the Western Alps, must be cut off by a N-S striking sinistral shear zone (Figs. 1 & 6). This putative shear zone, which is only schematically drawn in Fig. 6, delimits the western edge of the Ligurian Alps affected by these 50 km of Apenninic top N thrusting from a narrow strip of the Southern Alps retro-belt in the west.

Eastward, i.e. north of Genova, the northern limit of the metamorphic core of the Ligurian Alps bends into a N-S strike and is exposed at the surface forming the well-known Sestri-Voltaggio line (or zone; Cortesogno & Haccard 1984) that is often taken as representing the Alps-Apennines junction (e.g. Castellarin 2001). Although reworked in later Neogene times (Crispini et al. 2009) this line basically represented a Late Paleocene to Early Eocene normal fault (Hoogerduijn Strating 1994) sealed by the deposits of the Tertiary Piedmont Basin. This normal fault juxtaposes the high-pressure rocks of the Voltri Group in the west (1.5-2.5 GPa), attributed to the Alpine lower plate, with low metamorphic grade units to the east (Sestri-Voltaggio Zone, various Ligurian flysch units and finally the Antola nappe; see Fig. 8d in Molli et al. 2010), whose grade of metamorphism progressively decreases eastward and up-section (see Fig. 2 in Capponi et al. 2009). It finally reaches the diagenetic zone in terms of illite crystallinity (Ellero et al. 2001) over a short horizontal distance of only some 5-7 km. Following Elter & Pertusati (1973) who considered the Antola nappe as an integral part of the former Alps that became backthrusted during Alpine orogeny, we attribute the units east of the Sestri-Voltaggio normal fault to the Alpine upper plate (Fig. 6).

The unknown Alpine-type basement of the Torino hills buried under the deposits of the Tertiary Piedmont Basin and the Antola nappe were mapped together as parts of a strip of former Alpine upper plate units mostly buried under Cenozoic deposits in Fig. 6. During the Apenninic orogeny, together with the metamorphic core of the Ligurian Alps, they thrusted the External Liguride units of the Monferrato hills along the Rio Freddo Fault Zone (Piana 2000) striking NW-SE between the Torino and Monferrato hills. The Rio Freddo Fault Zone links up with the E-W-striking Villalvernia-Varzi line (Fig. 1), a Late Oligocene to Early Miocene thrust, active during the deposition of a part of the Tertiary Piedmont basin fill, with a strong sinistral strike slip component (Laubscher et al. 1992; Schumacher & Laubscher 1996). This sinistral transpressional fault is responsible for laterally transporting the External Ligurides exposed in the Monferrato hills in front of the Antola nappe and the Torino hills Alpine basement. At its eastern termination the Villalvernia-Varzi line bends into a N-S-strike as it transforms into an Apenninic top NNE thrust of the Antola nappe over the external Ligurides in the north and over the Internal Ligurides further south (Fig. 6). The polyphase history of the Antola nappe, with early and Alpine top-NW thrusting and folding, followed by top-NNE Apenninic thrusting over the Ligurides, is well documented (e.g. Marroni et al. 1999; 2002a; Levi et al. 2006). Offshore the continuation of this thrust is only tentatively drawn in Fig. 6; southward it is interpreted to eventually abut the northern continuation of the west-dipping upper/lower plate boundary of the Apennines orogen.

The next eastward unit of the Apennines, the Internal Ligurides, has also to be considered as having represented an integral part of the former Alps. It occupied an upper plate position before is became backthrusted to the NNE during the Apenninic orogeny. The internal Ligurides are characterized by the presence of classical ophiolites and an Upper Jurassic to Lower Cretaceous sedimentary cover (cherts, Calpionella limestone and Palombini shales) associated with Upper Cretaceous–Paleocene turbiditic sequences (Marroni & Pandolfi 1996) that represent the pelagic cover of the ophiolitic basement of the Piedmont-Liguria ocean (Decandia & Elter 1972) whose stratigraphy is virtually indistinguishable from that of the ophiolites of the Alps *sensu stricto* (Bernoulli et al. 2003). Moreover, since the pioneering work of Grandjacquet & Haccard (1977) several authors evidenced classical structural overprinting patterns at all scales, clearly showing that the early tectonic history of the Internal Ligurides was characterized by a top W to NW Alpine tectonic fingerprint, associated with low-grade metamorphism reaching some 300°C (Pertusati & Horrenberger 1975; Van Wamel 1987; Marroni & Pandolfi 1996; Ellero et al. 2001).

The External Ligurides are far-travelled and detached units that are confined to the northernmost Apennines between Torino and Bologna (Fig. 6). They overlie thin slices of Subligurian units and the Cervarola unit representing flysch deposits of the Adriatic foredeep outcropping in two small windows (Molli et al. 2010). The External Ligurides dominantly consist of detached Cretaceous-Paleocene flysch sequences (Helminthoid flysch) overlying basal mélange-type complexes (Marroni et al. 1998, 2001, 2002b). Two main subgroups of units can be recognized: those associated with ophiolites and with ophiolite derived debris (Molli 1996), and others without ophiolites and associated with fragments of Mesozoic sedimentary sequences and conglomerates with continental Adria affinity often remarkably similar to those found in the Alpine upper plate units of the Préalpes Romandes (Elter et al. 1966). Hence, the External Ligurides are regarded as representing the former ocean-continent transition within the Adria plate. Very commonly the stratigraphic base of the Oligocene part of the Epiligurian basins only exhibits a modest angular unconformity up to few tens of degrees (Piazza et al. 2016). Due to the syn-depositional sinistral transpression during Oligocene to Early Miocene times sealed by a Burdigalian unconformity mentioned earlier, the base of the Burdigalian and younger deposits of the Epiligurian succession locally overlies overturned older sequences of Late Cretaceous to Early Miocene successions (Piazza et al. 2016). In summary, the External Ligurides represent an ocean-continent transition area within the Adria upper plate of the Alps that was only locally overprinted by substantial late Alpine deformations of Oligocene to Early Miocene age, largely synchronous with stage 2 deformation in the Western Alps.

A substantial Alpine overprint has not been documented in the southerly adjacent Tuscan Ligurides, representing a Piedmont-Liguria ophiolitic sequence identical to that observed in the Internal Ligurides or the Alps (Principi et al. 2004). Due to massive extension locally leading to core complex formation (Brunet et al. 2000), Tuscany exposes deep portions of the Apenninic orogen that locally reached blueschist facies conditions with pressures up to some 1.3 GPa (Rossetti et al. 2002), mostly affecting the underlying Tuscan nappe that represents the lower plate unit during Apenninic orogeny. Locally, also the immediately overlying Tuscan Ligurides (Jolivet et al. 1998; Brunet et al. 2000; Rossetti et al. 2002) became involved in subduction during the Apenninic orogeny.  Hence, it is likely that the Tuscan Ligurides only became involved in W-directed Apenninic subduction and suffered penetrative deformation within a W-dipping subduction channel. A significant ophiolitic complex that is part of the Tuscan Ligurides is exposed on Gorgona Island (Orti et al. 2002). Metasediments associated with these ophiolites suffered blueschist facies metamorphism reaching about 1.5 GPa at around 25 Ma (Rossetti et al. 2001). Gorgona island is located far offshore Tuscany (Fig. 6) and very close to the northern continuation of the new plate boundary that formed during the earliest stages of Apenninic orogeny associated with W-directed subduction and subsequent roll back of the mantle slab below the North Apennines (Spakman & Wortel 2004). W-directed subduction initiated east of the Corsica basin that represents a >8.5 km thick rift basin filled with undeformed Late Oligocene to Burdigalian sediments (Mauffret et al. 1999; Finetti 2005) deposited as the Adriatic slab started to roll back, causing extension in the Corsica basin located in the Apenninic upper plate. The range of radiometric ages obtained for the high-pressure metamorphism in the Tuscan units is also significantly younger (26-20 Ma; Rossetti et al. 2002) when compared with the radiometric ages obtained in the ophiolitic units of Corsica where Cretaceous to latest Eocene ages (Vitale Brovarone & Herwartz 2013, and references therein) have been obtained. In summary, the Tuscan Ligurides are very likely to only have been affected by Apenninic orogeny, associated with W-directed subduction in latest Oligocene to Miocene times.

The Mesozoic Adriatic passive margin stratigraphy that accreted to the overriding European plate (Corsica) and the Ligurides during the Apenninic orogeny is exposed in the Tuscan nappe. The Chattian-Aquitanian Macigno turbidites (Di Giulio 1999) represent the youngest sediments involved. This 2.5-3 km thick turbiditic sequence is testimony of the deposition of siliciclastics derived from the Alps, particularly from the Lepontine dome (Garzanti & Malusa 2008). This dome shed an immense volume of metamorphic basement estimated at 17-20 x $10^3$ km$^3$ (Di Giulio 1999) as the Lepontine dome was uplifted by retro-thrusting and exhumed by erosion. This led to deposition of the Macigno pile of turbidites onto the still undeformed Adriatic foredeep during retro-folding and thrusting of the Alps north of the Periadriatic Line, predating most of the backthrusting of the Southern Alps in Lombardy (Schmid et al. 1996). Related exhumation, erosion and re-deposition onto the still undeformed Adriatic continental margin are largely synchronous with stage 2 top-WNW thrusting and WNW-directed indentation of the Ivrea mantle wedge (Stage 2, 35-25 Ma).

The more external Apennines thrust sheets involve flysch deposits of the foredeep only. The Cervarola thrust sheet is made up of Aquitanian to Early Langhian turbidites that still received detritus from the Alps. The more external Umbria-Marche thrust sheets involve the Marnoso Arenacea sequence of Serravallian to Tortonian age (Di Giulio 1999). The post-Messinian frontal imbricates are buried below younger deposits of the Po plain (Picotti & Pazzaglia 2008).

## 3. Lateral or temporal change in subduction polarity between Alps and Apennines? A discussion

Since the early days of Argand (1924) most authors (e.g. Mattauer et al. 1981; Durand-Delga 1984; Molli et al. 2006; Molli and Malavieille 2011; Handy et al. 2010) regarded the pre-Oligocene Corsican nappe stack as a continuation of the Alpine orogen. Some authors (e.g. Principi and Treves 1984; Lahondère, 1996; Jolivet et al. 1998; Brunet et al., 2000), however, regarded Alpine Corsica as an integral part of the Apennines, both having formed in connection with a W- dipping subduction zone since Cretaceous times. They considered Variscan Corsica as a backstop of the Apennines orogenic wedge onto which the deepest part of the Apennines accretionary complex, namely Alpine Corsica, was backthrusted. This view essentially transfers the area of the Alps-Apennines transition more to the north into the area around Genova, which is rather unlikely in the light of the available data on Alps and northern Apennines summarized in the previous chapters.

Debate also exists on whether the Alpine subduction zone with Eurasia in a lower plate position has continued beyond Corsica (e.g., Handy et al., 2010), or whether a lateral change in pre-Oligocene subduction polarity occurred in an area located somewhere between Corsica and Sardinia active since Late Cretaceous to Early Eocene times (e.g. Marchant & Stampfli 1997; Faccenna et al. 2003; Lacombe & Jolivet 2005; Lustrino et al. 2009; Argnani 2012; Advokaat et al., 2014; van Hinsbergen et al. 2014a). This transition is supposed to have occurred along a discrete trench-trench transform located SE of Corsica-Sardinia (e.g. Argnani 2012) or more diffusively, with partly overlapping trenches associated with opposite subduction polarity somewhere offshore east of Corsica (e.g. Lacombe & Jolivet 2005). An often quoted argument in favor of this view is the existence of Late Eocene magmatism in Sardinia (Lustrino et al. 2009) within but a tiny fraction (Calabona subvolcanics) of the dominantly latest Oligocene and younger igneous magmatic province of Sardinia, starting around 28 Ma and peaking during the 22–18 Myr time range. This younger magmatic province undoubtedly has to be interpreted as related to W-directed Apenninic subduction and roll back. However, the Late Eocene magmatism at the Calabona locality was also interpreted to represent the effect of the northwest-directed subduction system that started to consume ancient oceanic lithosphere of the Tethys Ocean. This subduction would have initiated in a time span prior to the Late Eocene that is long enough to culminate in 100-150 km of subduction required to generate an arc (Lustrino et al. 2009). Given the very slow Africa-Europe convergence rates in the pre-Late Eocene (only 250 km of convergence occurred between Africa and Eurasia prior to 35 Ma at the position of Sardinia, and even less farther to the west, van Hinsbergen et al. 2014a), this would suggest that an Alpine-vergent thrusting in Sardinia and farther to the west would have been a back-thrust above a long-lived Apennines-type subduction. However later, Lustrino et al. (2013) argued, based on geochemical arguments, that the Eocene manifestations of the subduction related magmatism are due to the partial melting of Hercynian lower crust rather than to the dehydration processes of subducting Alpine Tethys oceanic crust. They pessimistically concluded that “whatever subduction dip, convergence rate and depth of partial melting are chosen, it is not possible to obtain a geologically sound model able to explain the spatial–temporal relationships of the western Mediterranean subduction-related igneous districts”.

Since the pioneering work of Boccaletti et al. (1971) another group of authors (e.g. Guerrera et al. 1993; Doglioni et al. 1999; Michard 2006; Molli 2008; Handy et al. 2010) favored the idea of a temporal rather than lateral change in subduction polarity for the entire Apennines area (see discussion in Michard et al. 2006, their figure 9). The validity of this hypothesis is intimately linked to the question about kinematics and timing of thrusting, and, paleogeographic origin of the Calabrian-Peloritanian Belt that consists of high-grade Variscan basement nappes thrusted on top of Ligurian-type ophiolites (Schenk 1990; Vitale & Ciarcia 2013). It is also related to the question as to weather the Calabrian belt is part of an ALKAPECA block (Bouillin et al. 1986; Michard et al. 2006; Handy et al. 2010) rather than a part of Sardinia, i.e. Europe.

The data and their interpretation presented in the previous chapters favor a temporal rather than lateral change in subduction polarity at least for the Northern Apennines and their transition into the Alps. However, a more detailed discussion of the southern Apennines of Calabria is outside the scope of this contribution. The controversy around a lateral vs. temporal change in subduction polarity in Calabria needs be solved by future investigations in Calabria.

Although our data and interpretations support a temporal change in subduction polarity in the Northern Apennines realm and Corsica this interpretation faces a serious problem that is not yet solved. It is not easy to explain why this change occurred so shortly after the end of E-directed Alpine subduction in Corsica (about 35-37 Ma; Vitale Bavarone & Herwartz 2013). The switch in subduction polarity around Corsica is not precisely known but probably occurred in Latest Eocene to Early Oligocene times. According to kinematic and paleomagnetic reconstructions (Séranne 1999; Gattacecca et al. 2007; Advokaat et al. 2014) fast roll back of the East Ligurian oceanic crust and rotation of Corsica-Sardinia commenced at around 30Ma, followed by rotation of Sardinia between 21-16 Ma. This only leaves some 5-7 Ma for slab break and change of subduction polarity and sufficient depth of subduction in order to trigger roll back of the W-dipping slab. Apenninic W-directed subduction and fast roll-back was certainly triggered by the negative buoyancy of large parts of the Piedmont-Liguria Ocean that still stayed open when Alpine orogeny in the Ligurian Alps ended at the end of the Eocene (Handy et al. 2010). This points to a change from plate convergence-driven Alpine E- to SE-directed subduction in Western Alps and Corsica toward W-directed subduction almost entirely driven by slab roll back in the Northern Apennines.

## E. Summary and conclusions

The geological interpretation of the P-wave velocity model of Diehl et al. (2009) allows for depicting unprecedented details concerning the outlines of this Adriatic mantle wedge and integrates its outline into the overall structure of the Western Alps. Important along-strike changes are visualized in five radially arranged crustal-scale transects across the Alps (Fig. 2) and the tectonic maps of Figs. 1 & 6). The interpretation of these findings reveals an important difference between Eastern and Western Alps. While the upper plate of the Eastern Alps is defined by the Austroalpine nappe system consisting of crustal imbricates of continental material, the upper plate of the Western Alps is formed by non-metamorphic klippen of the Piedmont-Ligurian Oceanic crust and lithosphere (Molli 2008).

Based on this, and combined with a review of existing structural data, a 3-stage evolution is proposed for the formation of the arc of the Western Alps. A first stage (Late Cretaceous-35 Ma) is characterized by top-NNW thrusting in sinistral transpression causing at least some 260km displacement of the E-W-striking Central and Eastern Alps, together with the Adria micro-plate, towards N to NNW with respect to stable Europe. A second stage (35-25 Ma) is characterized by lateral indentation of Adria including the Ivrea mantle wedge at its leading edge towards the WNW. This culminated in some 100-150km of right-lateral strike-slip of the Ivrea mantle wedge, Western Alps and Adriatic micro-plate relative to the eastern Alps and Eurasia. This was accommodated in the Western Alpine nappe pile by top-WNW thrusting in the external zones of the central portion of the arc along the Penninic Frontal thrust combined with underthrusting of the Ivrea mantle wedge below the Western Alpine nappe pile. The third and final stage of arc formation (25-0Ma) was associated with orogeny in the Apennines and led to oroclinal bending in the southernmost Western Alps, simultaneously with the 50° counterclockwise rotation of the Corsica-Sardinia block. Particularly during the second stage of the evolution of the arc of the Western Alps the Ivrea mantle wedge, exhumed to shallow crustal depth already during Jurassic rifting at the passive continental margin of the Adria continent and representing a high strength wedge, played a crucial role by controlling the lateral ends of indentation near Cuneo and the Rhone-Simplon fault.

The third stage of the evolution of the arc of the Western Alps calls for an analysis of the Alps-Apennines transition zone. A pre-requisite for understanding this transition is the realization that those parts of the Piemont-Liguria oceanic domain that form the upper plate in the Western Alps remained open until the Late Eocene. A combination of the gravitational potential of a southwards enlarging domain of old and oceanic lithosphere, the arrival of the Corsican continent in the Alpine trench resisting subduction, and ongoing Africa-Europe convergence induced a change in subduction polarity from Alpine S to SE-directed subduction towards NW-directed subduction and roll back of the Apennines slab, pulled by the negative buoyancy of old oceanic lithosphere large parts of which remained unaffected by Alpine orogeny.

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**Figure legends**

Fig. 1:

Tectonic map of the arc of the Western Alps after Schmid et al. (2008) and Bousquet et al. (2012b) with indication of the traces of the profiles depicted in Fig. 2. The selected faults indicated in red are: RSL: Rhone-Simplon Line; PL: Periadriatic Line; PF: Penninic Front; ACL: Argentera-Cuneo Line; AFSV: front of the metamorphic core of the Alps and Sestri-Voltaggio line; VVL: Villalvernia-Varzi line. Abbreviations in black are: LA: Ligurian Alps; TH: Torino hills; MH: Monferrato hills; GB: Giudicarie transpressive belt.

Fig. 2:

Five selected geological-tectonic transects across the Western Alps. The sections are superimposed with and partly based on vertical cross sections across the final P-wave velocity model of Diehl et al. (2009) indicated by red dashed lines. The Moho (blue line) is taken from a combination of controlled source seismology, local earthquake tomography and receiver function analysis (Wagner et al. 2012; Spada et al. 2013). See text for data sources and further discussion and Fig. 1 for the traces of the profiles. a: Argentera transect; b: Pelvoux transect; c: ECORS-CROP transect; d: NFP 20-West transect; e: Ticino transect.

Fig. 3:

Horizontal section across 3-D P-wave tomographic model of Diehl et al. (2009) at a depth of 20km. Yellow triangles indicate seismic stations used for tomographic inversion, crosses indicate inversion grid nodes. Note that the velocity structure is shown as percentage change relative to the 1-D initial reference model rather than in terms the P-wave velocities shown in Fig. 2. Bold black contours include well-resolved regions covering most of the area depicted. Compare Fig. 1 for major tectonic units outlined: RSL: Rhone-Simplon Line; PL: Periadriatic Line; PF: Penninic Front; ACL: Argentera-Cuneo Line. Comparison with Fig. 2 shows that the entire area with P-wave anomaly >10%, as well as a part of the area with P-wave anomaly >5% between north of Locarno and Cuneo is within the Ivrea mantle wedge.

Fig. 4:

Three generations of stretching lineations and associated senses of shear around the arc of the Western Alps. Red arrow: stage 1 top N transport; blue arrows: stage 2 top WNW transport; green: stage 3 top SW transport; arrow pairs reflect the variation of the measured lineations; see text for further details. Numbered areas of measurements and literature sources are as follows: area 1, Valaisan and Penninic Front (Loprieno et al. 2011); area 2, Zone Houillère of the Briançonnais Unit (Bucher et al. 2004); area 3, Subbriançonnais and Penninic Front (Ceriani et al. 2001; Ceriani & Schmid 2004); area 4, Penninic Front (top NW) and basal decollement of Cenozoic flysch of the Dauphinois (top SW) (Bürgisser & Ford 1998; Trullenque 2005; Trullenque et al. 2006); area 5, eclogite facies lineations (top N) and greenschist facies lineations (top SW) in M. Viso ophiolites (Philippot 1988, 1990); area 6, Latest Eocene “schistes è blocs” at the base of the Embrunais-Ubaye nappes (Trullenque 2005); area 7, Penninic Front N of the Argentera massif (Trullenque 2005); area 7, thrust transport constraints from the Dauphinois of the Digne thrust system (Lickorish & Ford 1998, Ford et al. 2006).

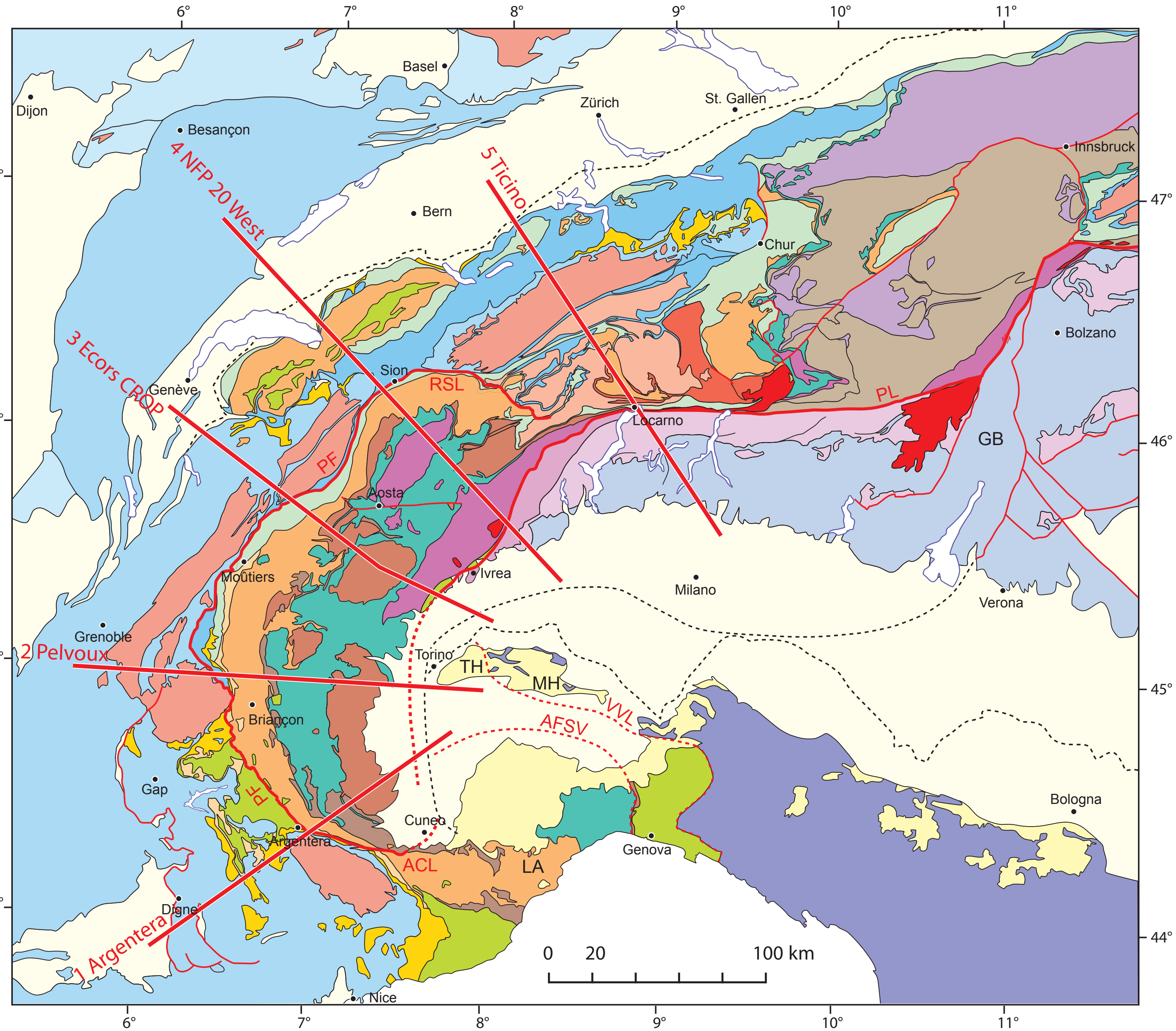
Fig. 5:

Kinematic restoration of the Alps-Apennines orogenic system using Gplates reconstruction software (e.g. van Hinsbergen et al. 2014a) holding Europe fixed; for easier orientation all time slices also show the present day position of the external massifs and Tauern window in the Alps, as well as the present-day position of northern Adria. Arrowheads indicate displacement from a given time slice to the next one, they are anchored at the position of a chosen a set of key locations at the chosen time slice. Red lines: coast lines and outlines of external massifs and Tauern Window in the Alps; blue lines: outlines of the Argentera, Pelvoux and Tambo-Suretta nappes; green area: horizontal section across the Ivrea mantle slice taken from Fig. 3. The present day position of numbered locations shown is as follows: 1: Sardinia; 2: Corsica; 3: Dora Maira massif (Briançonnais); 4: Gran Paradiso massif (Briançonnais); 5: Tambo-Suretta nappes (Briançonnais); 6: Oberammergau (front Austroalpine nappes); 7: Dimaro (eastern end Tonale Line); 8: Mauls (western end Pustertal Line; 9: Slovenske Konjice (eastern end Gailtal line); 10: Cuneo (S end of Ivrea wedge); 11: Brione (N end of Ivrea wedge); 12: Pavia (Adria microplate); 13: Monti Lessini (Adria microplate); 14: Lignano (Adria microplate); 15: W of Ancona (frontal thrust of Umbria-Marche Apennines); 16: S of Parma (frontal thrust of external Ligurides, Apennines); 17: offshore Baleares at 25 Ma ago (N front of oceanic Adria microplate = Piedmont-Liguria lower plate).

Fig. 6:

Interpretative map sketch of the Alps-Apennines transition; see text for further discussion.

# Tectonic map of the Western Alps




## Cenozoic basins

- Cenozoic in general
- Cenozoic of Piedmontese basin

## Lower Plate Units of the Alps

European continental margin

- Cenozoic flysch (Dauphinois and Helvetic Nappes)
- European Mesozoic undeformed
- European Mesozoic deformed
- Allochthonous Mesozoic & Cenozoic (Helvetic Nappes)
- European basement (External Massifs)
- European upper crust of the Subpenninic Units
- High-pressure Subpenninic Units

Valaisan partly oceanic domain

- Valaisan

Briançonnais micro-continent

- Sub-Briançonnais
- Briançonnais
- High-pressure Briançonnais
- Pre-Piemontais Acceglio

Sesia-Dent Blanche Extensional Allochthons

- Sesia, Dent Blanche, Margna, Sella Units

Piedmont-Liguria Ocean

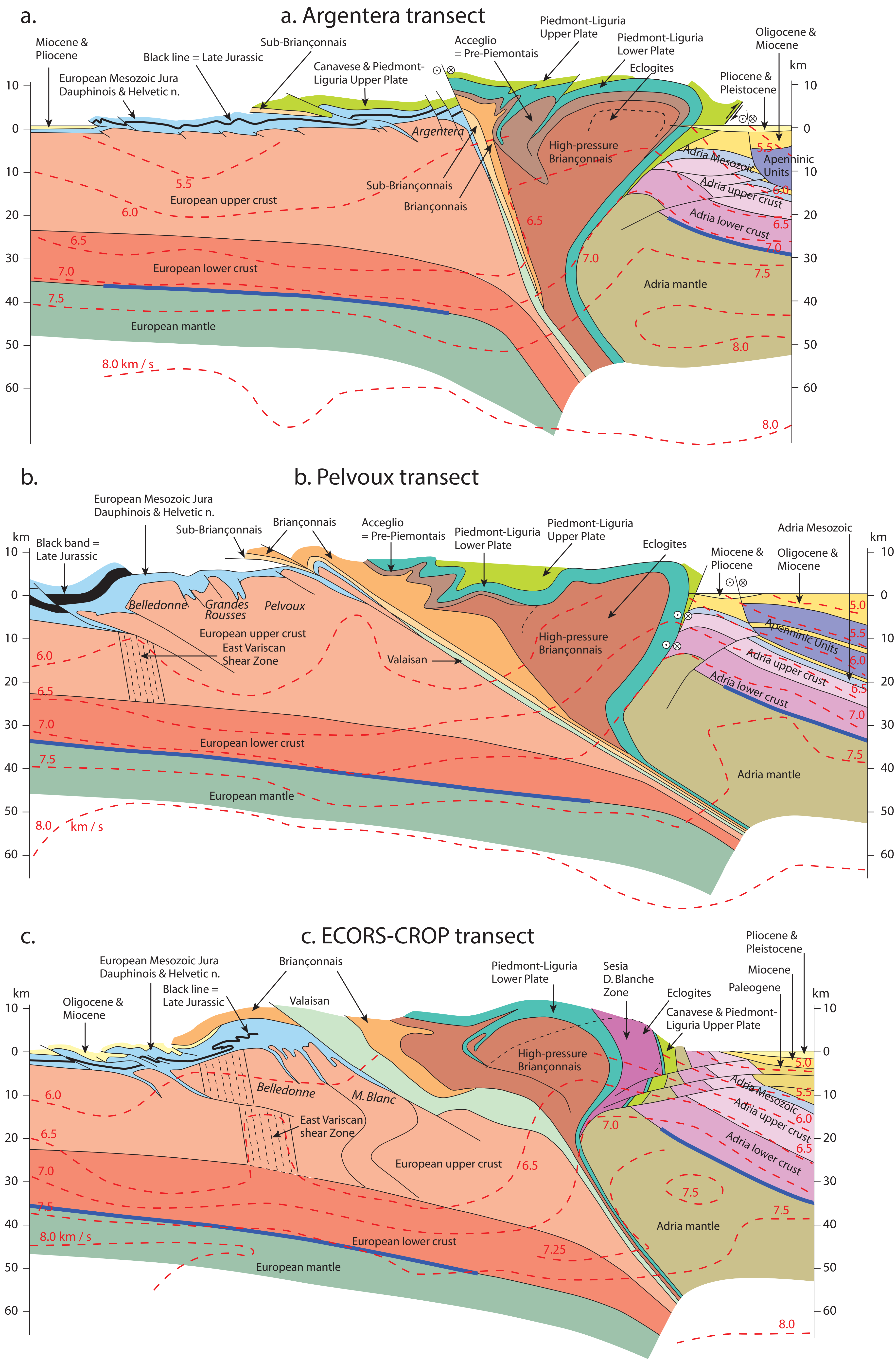
- Lower Plate Piedmont-Liguria

## Upper Plate Units of the Alps

- Upper Plate Piedmont-Liguria and distal continental margin of Adria
- Adria Permo-Mesozoic
- Adria upper crust
- Adria lower crust
- Austroalpine Permo-Mesozoic
- Austroalpine basement
- Cenozoic plutons

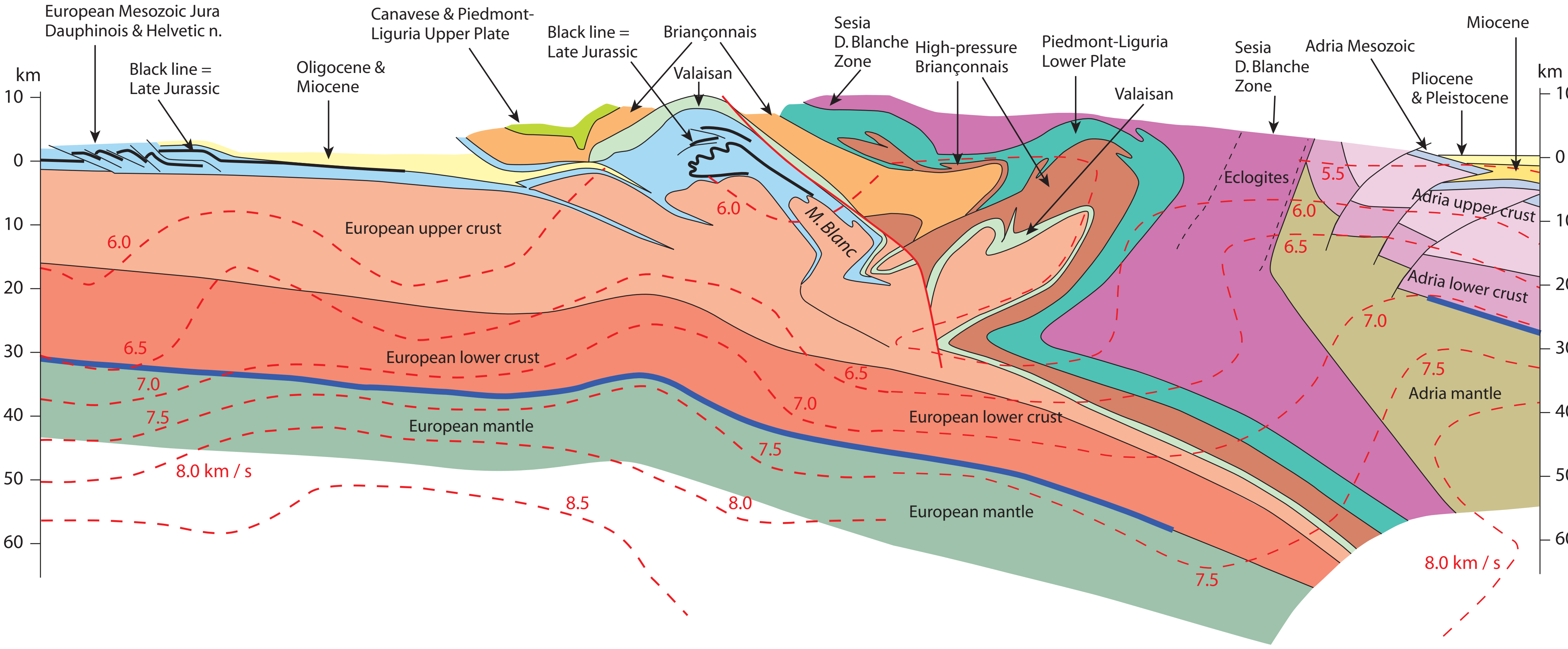
## Apennines

- Apenninic Units in general


a.
a. Argentera transect
Miocene & Pliocene
European Mesozoic Jura Dauphinois & Helvetic n.
Black line = Late Jurassic
Sub-Briançonnais
Canavese & Piedmont-Liguria Upper Plate
Acceglio = Pre-Piemontais
Piedmont-Liguria Upper Plate
Piedmont-Liguria Lower Plate
Eclogites
Oligocene & Miocene
Pliocene & Pleistocene
km
Argentera
High-pressure Briançonnais
Adria Mesozoic
Apenninic Units
Adria upper crust
Adria lower crust
Sub-Briançonnais
Briançonnais
European upper crust
European lower crust
European mantle
Adria mantle
8.0 km / s
b.
b. Pelvoux transect
European Mesozoic Jura Dauphinois & Helvetic n.
Sub-Briançonnais
Briançonnais
Acceglio = Pre-Piemontais
Piedmont-Liguria Lower Plate
Piedmont-Liguria Upper Plate
Eclogites
Adria Mesozoic
Miocene & Pliocene
Oligocene & Miocene
km
Black band = Late Jurassic
Belledonne
Grandes Rousses
Pelvoux
European upper crust
East Variscan Shear Zone
Valaisan
High-pressure Briançonnais
Apenninic Units
Adria upper crust
Adria lower crust
European lower crust
European mantle
Adria mantle
km / s
c.
c. ECORS-CROP transect
European Mesozoic Jura Dauphinois & Helvetic n.
Briançonnais
Piedmont-Liguria Lower Plate
Sesia D. Blanche Zone
Eclogites
Pliocene & Pleistocene
Miocene
Paleogene
Canavese & Piedmont-Liguria Upper Plate
Oligocene & Miocene
Black line = Late Jurassic
Valaisan
km
High-pressure Briançonnais
Belledonne
M. Blanc
East Variscan shear Zone
Adria Mesozoic
Adria upper crust
Adria lower crust
European upper crust
Adria mantle
8.0 km / s
European lower crust
European mantle

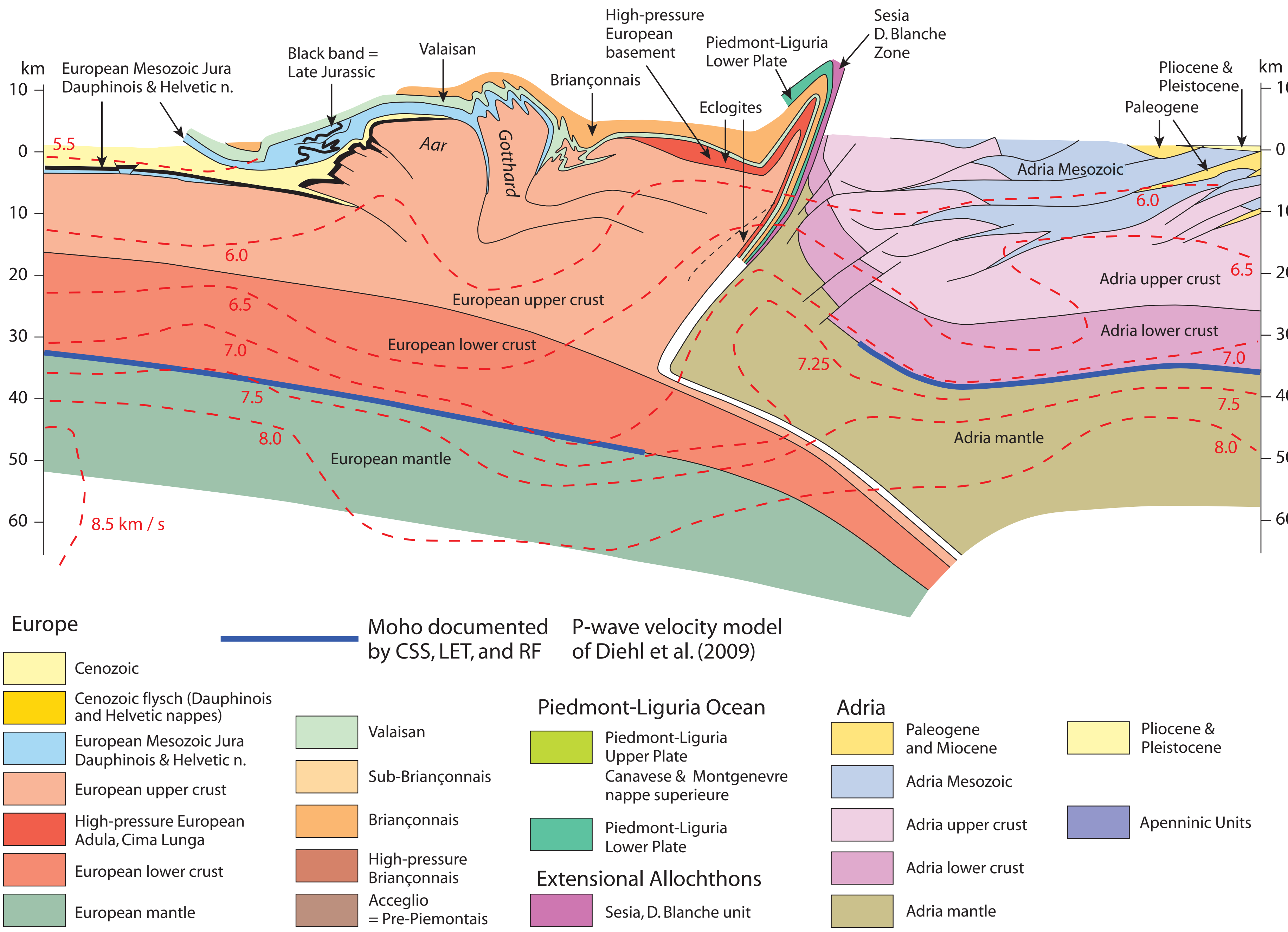
d.

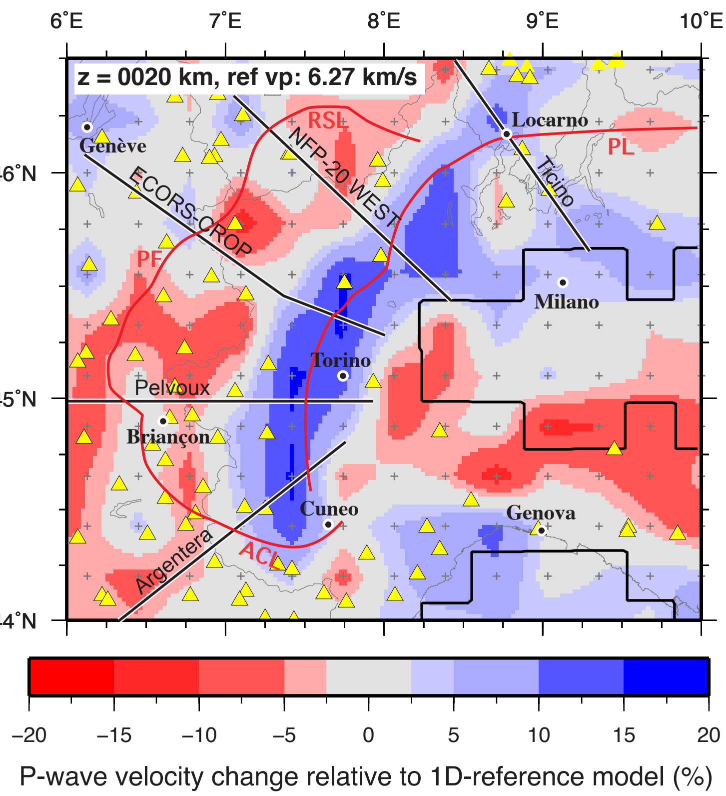
## d. NFP 20-West transect

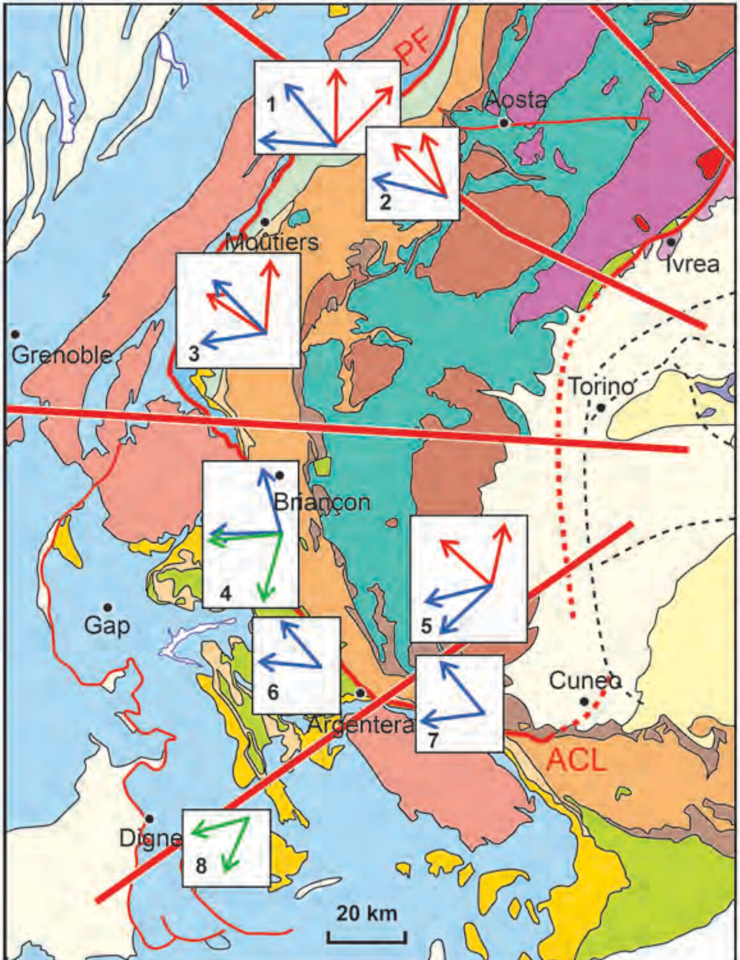



e.

## e. Ticino transect




z = 0020 km, ref vp: 6.27 km/s
6°E
7°E
8°E
9°E
10°E
46°N
45°N
44°N
Genève
Locarno
Milano
Torino
Briançon
Cuneo
Genova
RSL
PL
PF
ACL
ECORS-CROP
NFP-20 WEST
Ticino
Pelvoux
Argentera
−20
−15
−10
−5
0
5
10
15
20
P-wave velocity change relative to 1D-reference model (%)




1: Valaisan & Penninic Front (Loprieno et al. 2011)

2: Zone Houillère Briançonnais (Bucher et al. 2004)

3: Subbriançonnais & Penninic Front (Ceriani et al. 2001; Ceriani & Schmid 2004)

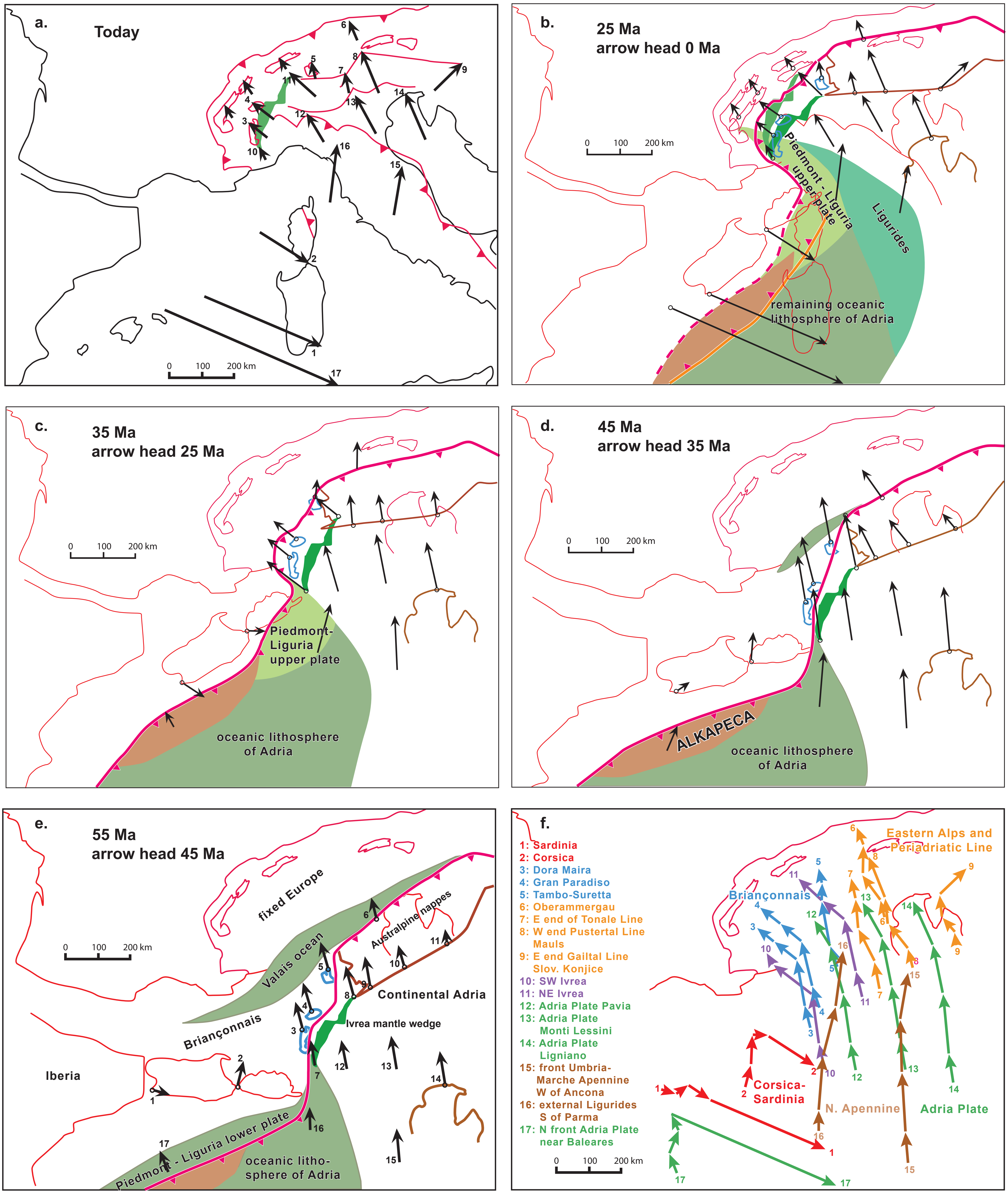
4: Top-NW Penninic Front & top SW basal decollement of the Cenozoic flysch of the Dauphinois (Bürgisser & Ford 1998; Trullenque 2005; Trullenque et al. 2006)

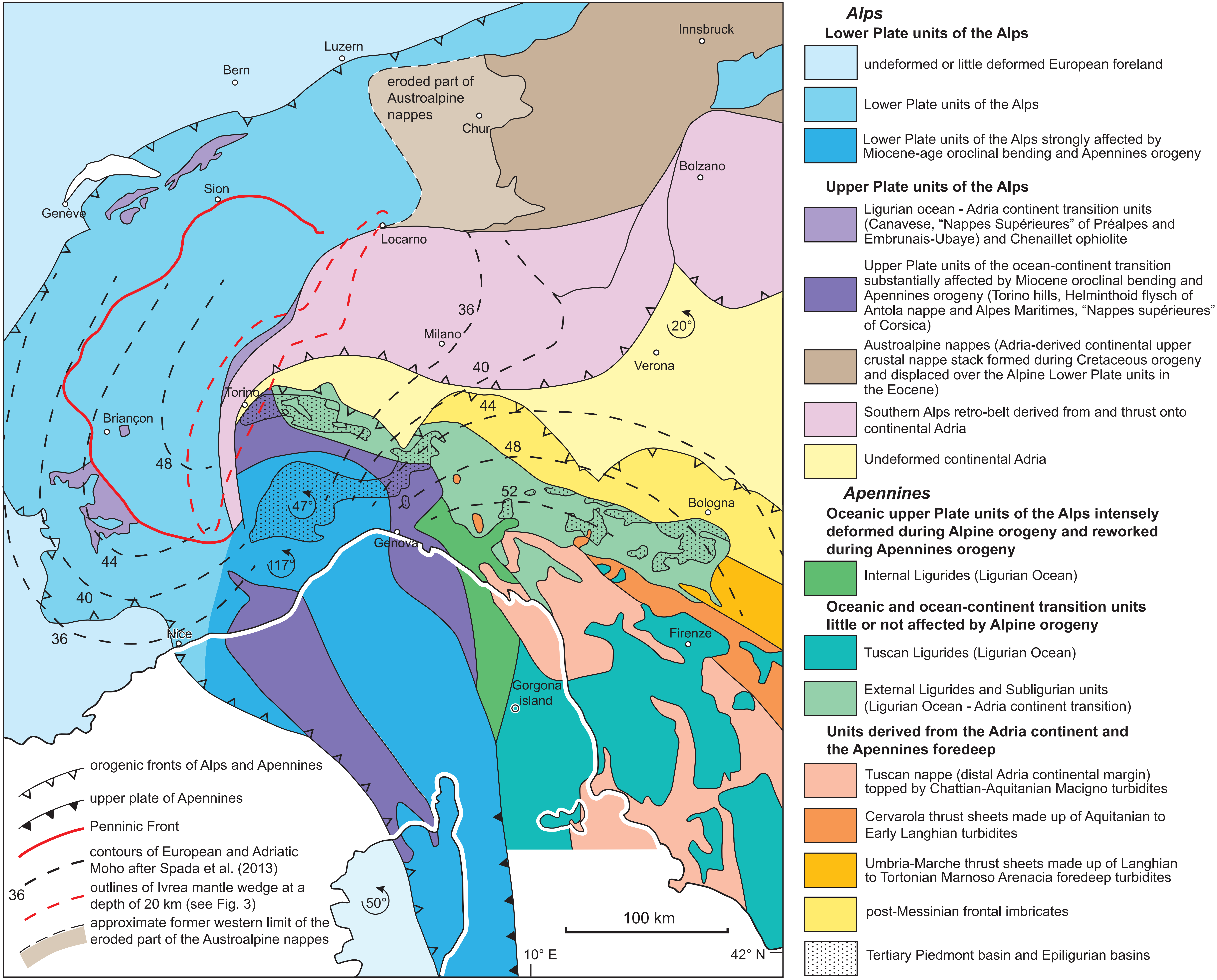
5: Eclogite facies top-N lineations and greenschist facies top-SW lineations M. Viso ophiolites (Philippot 1988, 1990)

6: “schistes à blocs” base Embrunais-Ubaye nappes (Trullenque 2006)

7: Penninic Front (Trullenque 2005)

8: Dauphinois of the Digne thrust system (Lickorish & Ford 1998; Ford et al. 2006)


a.
Today
0 100 200 km
b.
25 Ma
arrow head 0 Ma
Piedmont - Liguria upper plate
Ligurides
remaining oceanic lithosphere of Adria
c.
35 Ma
arrow head 25 Ma
Piedmont-Liguria upper plate
oceanic lithosphere of Adria
d.
45 Ma
arrow head 35 Ma
ALKAPECA
oceanic lithosphere of Adria
e.
55 Ma
arrow head 45 Ma
fixed Europe
Valais ocean
Austroalpine nappes
Continental Adria
Briançonnais
Ivrea mantle wedge
Iberia
Piedmont - Liguria lower plate
oceanic litho-sphere of Adria
f.
1: Sardinia
2: Corsica
3: Dora Maira
4: Gran Paradiso
5: Tambo-Suretta
6: Oberammergau
7: E end of Tonale Line
8: W end Pustertal Line Mauls
9: E end Gailtal Line Slov. Konjice
10: SW Ivrea
11: NE Ivrea
12: Adria Plate Pavia
13: Adria Plate Monti Lessini
14: Adria Plate Ligniano
15: front Umbria-Marche Apennine W of Ancona
16: external Ligurides S of Parma
17: N front Adria Plate near Baleares
Eastern Alps and Periadriatic Line
Briançonnais
Corsica-Sardinia
N. Apennine
Adria Plate



Innsbruck
Luzern
Bern
eroded part of Austroalpine nappes
Chur
Bolzano
Genève
Sion
Locarno
36
Milano
40
Verona
20°
Torino
44
Briançon
48
48
52
47°
Bologna
Genova
117°
44
40
Firenze
Nice
36
Gorgona island
50°
100 km
10° E
42° N
orogenic fronts of Alps and Apennines
upper plate of Apennines
Penninic Front
contours of European and Adriatic Moho after Spada et al. (2013)
36
outlines of Ivrea mantle wedge at a depth of 20 km (see Fig. 3)
approximate former western limit of the eroded part of the Austroalpine nappes
Alps
Lower Plate units of the Alps
undeformed or little deformed European foreland
Lower Plate units of the Alps
Lower Plate units of the Alps strongly affected by Miocene-age oroclinal bending and Apennines orogeny
Upper Plate units of the Alps
Ligurian ocean - Adria continent transition units (Canavese, "Nappes Supérieures" of Préalpes and Embrunais-Ubaye) and Chenaillet ophiolite
Upper Plate units of the ocean-continent transition substantially affected by Miocene oroclinal bending and Apennines orogeny (Torino hills, Helminthoid flysch of Antola nappe and Alpes Maritimes, "Nappes supérieures" of Corsica)
Austroalpine nappes (Adria-derived continental upper crustal nappe stack formed during Cretaceous orogeny and displaced over the Alpine Lower Plate units in the Eocene)
Southern Alps retro-belt derived from and thrust onto continental Adria
Undeformed continental Adria
Apennines
Oceanic upper Plate units of the Alps intensely deformed during Alpine orogeny and reworked during Apennines orogeny
Internal Ligurides (Ligurian Ocean)
Oceanic and ocean-continent transition units little or not affected by Alpine orogeny
Tuscan Ligurides (Ligurian Ocean)
External Ligurides and Subligurian units (Ligurian Ocean - Adria continent transition)
Units derived from the Adria continent and the Apennines foredeep
Tuscan nappe (distal Adria continental margin) topped by Chattian-Aquitanian Macigno turbidites
Cervarola thrust sheets made up of Aquitanian to Early Langhian turbidites
Umbria-Marche thrust sheets made up of Langhian to Tortonian Marnoso Arenacia foredeep turbidites
post-Messinian frontal imbricates
Tertiary Piedmont basin and Epiligurian basins