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Style of Alpine tectonic deformation in the Castellane fold-and-thrust belt (SW Alps, France): Insights from balanced cross-sections

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ABSTRACT

This study proposes a reappraisal of the role of the basement tectonics in the structuration of the Alpine foreland, across the Castellane fold-and-thrust belt located in southwestern Alps. We construct three 30 km length N–S balanced cross-sections across the entire fold-and-thrust belt, in order to quantify the amount of horizontal shortening due to the Pyrenean and Alpine deformations. We then assess the role of the basement inherited structures during the compressional phases which resulted in the exhumation of the Argentera–Mercantour External Crystalline Massif and the Barrot Dome. The construction of these balanced cross-sections suggests a dominant thick-skinned deformation style, which includes the reactivation of inherited Permian and Early Cretaceous basement normal faults. We reconstructed three restoration stages: the oldest one highlights the geometry of Permian and Early Cretaceous extensional structures; the intermediate one after the Late Cretaceous evidences a first compressional episode linked to the Pyrenean compression; and the latest one in the Oligocene shows the first Alpine folds resulting from the southward thrusting of the internal units over the external ones along of the Penninic Frontal Thrust. Balanced cross-sections suggest very moderate crystalline basement–sedimentary cover decoupling. On the western and central cross-sections, the estimated amount of shortening ranges from 9.5 to 10 km (21%) whereas on the easternmost one shows ~5 km of shortening (9%). These shortening values are consistent with previously published estimates in the surrounding foreland subalpine chains. They highlight a decreasing value of Pyrenean shortening toward the east, while the Alpine shortening dominates and amplifies this first phase in a similar direction. We interpret this dominantly thick-skinned structural style as a possible consequence of the Neogene thermal weakening in the European passive margin above the Ligurian slab rollback.

1. Introduction

The foreland domains of convergent orogens are formed by fold-and-thrust belts, which comprise sedimentary layers overlying crystalline/hard-rock basements. Shortening in these domains is controlled by a regionally-dominated compressional strain state and occurs in different geodynamic contexts such as subduction accretionary wedges and collisional foreland thrust belts (e.g., Dahlen, 1990; Davis et al., 1983; Suppe, 1987). In foreland thrust belts, the deformation of the sedimentary cover is the outermost expression of deeper crustal shortening occurring inwards along intra-basement thrusts (e.g., Jackson, 1980). The width of the fold-and-thrust belt, i.e. the horizontal offset between the frontal foreland thrust and the first (i.e. outermost) intra-basement one, depends on the presence of a horizontal level of décollement that allows a mechanical decoupling between the upper and lower units. At first approximation, when this décollement exists and is efficient, foreland deformation is characterized by a thin-skin tectonic style (e.g., Affolter and Gratier, 2004). This model has for instance been generalized for the front of the Alps (e.g., Phillipe et al., 1998). On the other hand, when the décollement is inefficient or inexistent, the deformed cover remains globally coupled to its basement. This so-called thick-skin deformation style is particularly visible in the case of the Pyrenean compressional phase in Provence, where the cover remains partly coupled to its basement (Espurt et al., 2012; Roure and Colletta, 1996), or in the Alpine External Crystalline Massif (ECM) (Bellahsen et al., 2012; Burkhard and Sommaruga, 1998). In detail, the occurrence of thick- or thin-skin deformation styles is still an open question and in some cases, especially on ancient and previously thinned continental margins, both can occur simultaneously (e.g., Jackson, 1980). At the 1st order however, the mechanical resistance of the lithosphere seems to play a first-order role in the style of shortening, thin-skin styles being prominent in old and rigid cratonic lithospheres (Mouthereau et al., 2013). Inherited basement faults and their associated sedimentary basins can also favor a thick-skin tectonic style, especially when the (lower) crust is hot and ductile (Bellahsen et al., 2012; Nilfouroushan et al., 2013). Discriminating between thin- and thick-skin tectonic styles is therefore not straightforward, but is necessary to address important questions such as: how is a foreland thrust belt structured, and what are the crustal-scale settings which control the

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deformation of the crystalline part of the crust and its related sedimentary cover during collisional shortening?

This questioning requires a better understanding of the role of essential parameters like the degree of basement–cover coupling resulting from the effectiveness of the décollement, and the importance of inherited crustal faults and basins that may have structured the continent before the onset of compressional deformation. Indeed, the existence of preexisting basement geometry may strongly influence the style of deformation (Bellanger et al., 2014; Boutoux et al., 2014; Nilfouroushan et al., 2013). For example, in the central Apennines, the structural style has been proposed to be thin-skinned (Ghisetti et al., 1993), which implies a very large amount of shortening (about 50%). However, taking into account the reactivation of basement structures inherited from a former extensional tectonic phase (Tozer et al., 2002) implies that calculated shortening with this thick-skin tectonic model is significantly lower (about 20%) than with the thin-skin model. In a given context, the style of tectonic deformation (thin-skin or thick-skin) implies different crustal mechanical behaviors and timings of crustal and cover shortening (Lacombe and Mouthereau, 2002).

In the Alps, the characterization of the style of deformation in the external domain is still an open question. The subalpine belts (Bornes, Bauges, Chartreuse, Vercors) are clearly thin-skinned belts in which shortening is linked to an efficient décollement located within the Liassic or Triassic layers (Affolter et al., 2008; Deville and Chauvière, 2000); deformation along this décollement was activated by displacement along basement thrusts inwards (see synthesis in Bellahsen et al., 2014). Further south, shortening is mainly accommodated along the Digne nappe thrust, which subsurface structure is often drawn assuming a thick-skinned style (Lickorish and Ford, 1998). The Digne nappe connects laterally southward to the Castellane fold-and-thrust belt (also called the Castellane Arc), which structure has been proposed to be thin-skinned. However, the Pyrenean structures further west, as well as their Alpine reactivation are thick-skinned (Espurt et al., 2012; Lacombe and Jolivet, 2005 and references therein). One may thus wonder whether the Castellane Arc is thin- or thick-skinned. Depending on these possible tectonic styles, the related cover deformation might root down into basement thrusts, or just be transferred on a decoupling interface, and the amount of estimated shortening might change by a factor of two, or more (e.g., Ghisetti et al., 1993).

In this study, we provide new balanced and restored cross-sections located in the Castellane Arc between the Argentera–Mercantour External Crystalline Massif (ECM) and the Ligurian Sea. These cross-sections allow us to discuss the effectiveness of the basement–cover décollement, the resulting degree of coupling between both units, the possible inversion of inherited extensional structures formed from the Permian to the Lower Cretaceous (Lemoine et al., 1986), and the localization of the deformation during the Pyrenean and Miocene Alpine compressions.

2. Geological setting

The Alpine orogenic belt has developed between the Apulian and Eurasian plates following the closure of the Ligurian part of Tethys Ocean (e.g., Dumont et al., 2012; Lemoine et al., 1981, and references therein). In the Western Alps, oceanic subduction beneath Apulia took place until the Middle to Late Eocene and was followed by continental collision in the Oligocene (Duchêne et al., 1997; Lanari et al., 2014; Lemoine et al., 2000; Simon-Labric et al., 2009).

The External Alpine domain is made up of the deformed sedimentary cover and basement of the former European margin of the Neotethys Ocean previously thinned by continental rifting during the Permian (e.g., Lapierre et al., 1999; Ziegler and Stampfli, 2001) and the Lower to Middle Jurassic (Lemoine et al., 2000) extensional phases. The folded sedimentary cover crops out in the External subalpine chains (from the Jura to the area of Nice), and rests upon European Hercynian crystalline basement and Permian basins, which are visible in the western arc of ECMs. In southwestern Alps, the Mesozoic cover has undergone several compressional tectonic events that occurred from the Late Cretaceous to the Quaternary in response to the Pyrenean and Alpine compressions (Lemoine et al., 2000).

The Castellane fold-and-thrust belt is located between the Argentera–Mercantour ECM, the Valensole foreland basin and the Provence foreland (Fig. 1). In this area, the last Alpine collisional stage was preceded by the opening of the Ligurian basin between Corsica–Sardinia and Europe from the Late Oligocene to the Middle Miocene (e.g., Gattacceca et al., 2007; Rollet et al., 2002) due to back-arc opening above Tethyan slab roll-back (Jolivet and Faccenna, 2000). This area is characterized by a southward propagation of folds and thrusts affecting the Meso-Cenozoic sedimentary cover in response to continuous N–S compression (e.g., Giannerini et al., 2011; Ritz, 1992; Sanchez et al., 2010, 2011a; and references therein). The Mesozoic sequence is complete from the Triassic to the Cretaceous (Fig. 2); the Lower Triassic formations are mechanically coupled to the basement and represent a marker layer to infer its presence close to the surface. The Jurassic sedimentary pile is represented by thick (600 m) calcareous deposits and is generally thought to be mechanically decoupled from the Lower Triassic by a décollement level in gypsum and cellular dolomites of the Upper Triassic (Keuper). The Cretaceous is marked by marls and calcareous marls with large thickness variations (600 m to 1000 m), which are principally controlled by the activity of normal faults during the Lower Cretaceous. The Tertiary clastic basins are unconformably deposited onto the Upper Cretaceous marly sequences, witnessing an emersion of the Late Mesozoic sedimentary formations between the end of the Late Cretaceous and the Lower Eocene (made of reef limestones). This Meso-Cenozoic sedimentary pile rests unconformably on a Paleozoic basement made of crystalline rocks inherited from the Hercynian orogeny, and of Carboniferous and Permian basins.

A balanced N–S cross-section through the Castellane area has been proposed assuming a thin-skinned tectonic style (Laurent et al., 2000; Fig. 3). The orientation of this cross-section is based on the assumption that there is no out-of-plane shortening. Actually, between the Ligurian Sea and the Dôme du Barrot, the direction of shortening linked to the exhumation of the External crystalline Argentera–Mercantour Massif from 26 Ma to present (Bigot-Cormier et al., 2006; Sanchez et al., 2011a, 2011b) is north–south (Bauve et al., 2014; Ritz, 1992). The propagation of E–W to NE–SW compression related to the Penninic Front propagation during Oligocene times largely affects the internal zones and the Nice Arc domain, which displays clear interference of E–W and N–S fold structures (Bauve et al., 2012; Giannerini et al., 2011). Yet, these interference folds are absent from the Castellane fold-and-thrust belt, since the front of E–W compressional deformation is located east of it, along the Var Valley (e.g., Schreiber et al., 2011). According to Laurent et al. (2000), the shortening in the sedimentary cover is accommodated by thrust-related anticlines organized in a ramp-and-flat system, which sole into the mechanically weak interface between the Upper Triassic units and the basement (or sometimes into Early Triassic units which are coupled to the basement).

3. Balanced cross-sections and restorations

In this study, we have made three ~40 km-long balanced cross-sections oriented N–S between the Argentera–Mercantour ECM and the Ligurian Sea (Fig. 4). The orientation of these cross-sections is in agreement with the global shortening orientation during the exhumation of the Argentera–Mercantour Massif from the Miocene to the Quaternary in the Castellane fold-and-thrust belt (Ritz, 1992; Sanchez et al., 2011a, 2011b). We used sedimentary thicknesses, strata dip and geometry according to the corresponding geological maps and notices (1:50,000 geological maps of Grasse–Cannes, Roquesteron and Puget-Théniers, Bonifay et al., 1970; Faure-Muret and Fallot, 1957; Goguel, 1980, respectively) and punctual field observations.

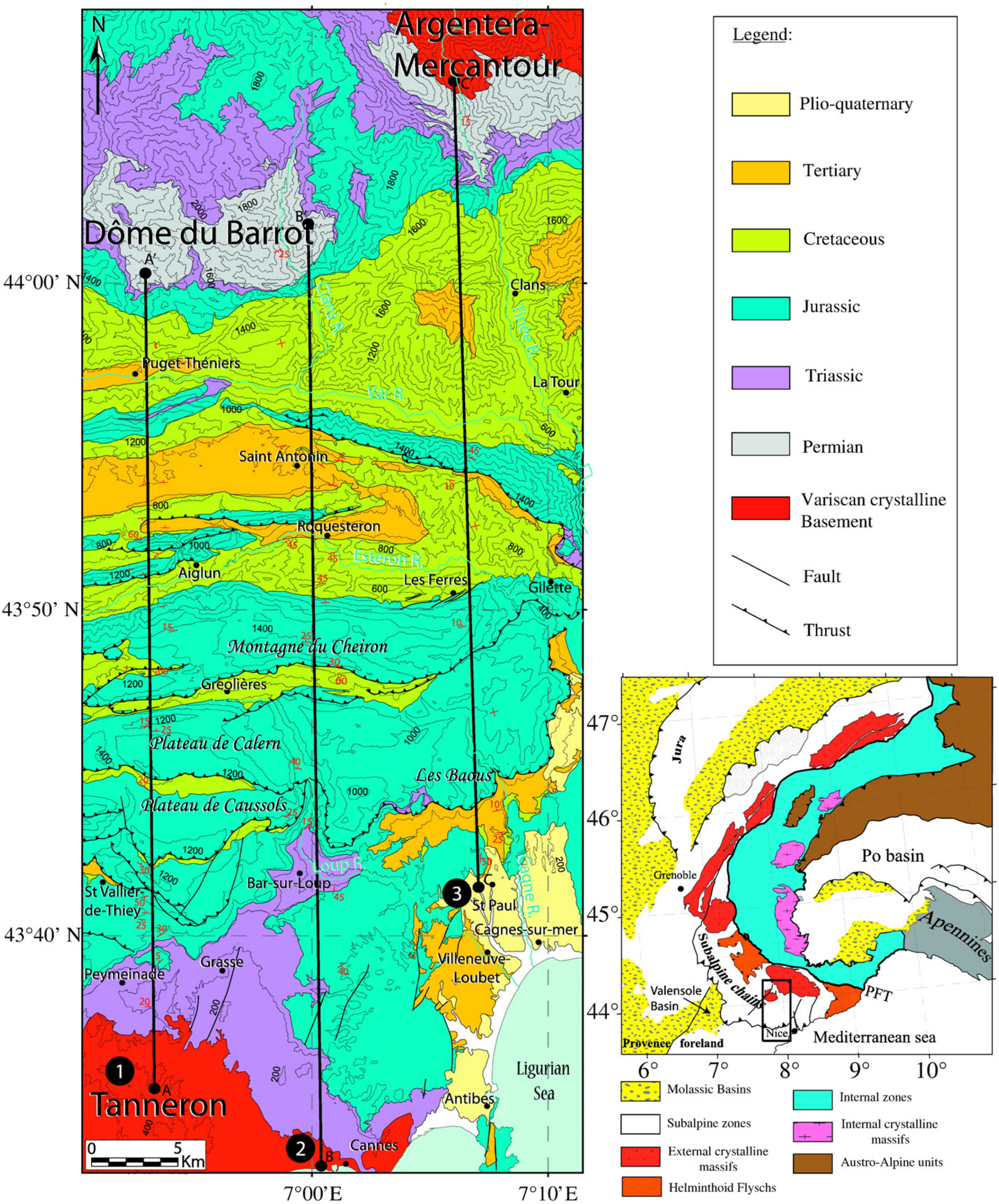

Fig. 1. Geological map of the eastern part of the Castellane fold-and-thrust belt between the Argentera and Maures–Tanneron crystalline massifs (drawn after the 1:50,000 geological maps of Grasse–Cannes, Roquesteron and Puget-Théniers by Bonifay et al., 1970; Faure-Muret and Fallot, 1957; Goguel, 1980, respectively, and new measurements). Inset shows the location of the study area in the framework of the Alps mountain belt. Solid lines indicate the location of the three N–S cross-sections presented in this study.

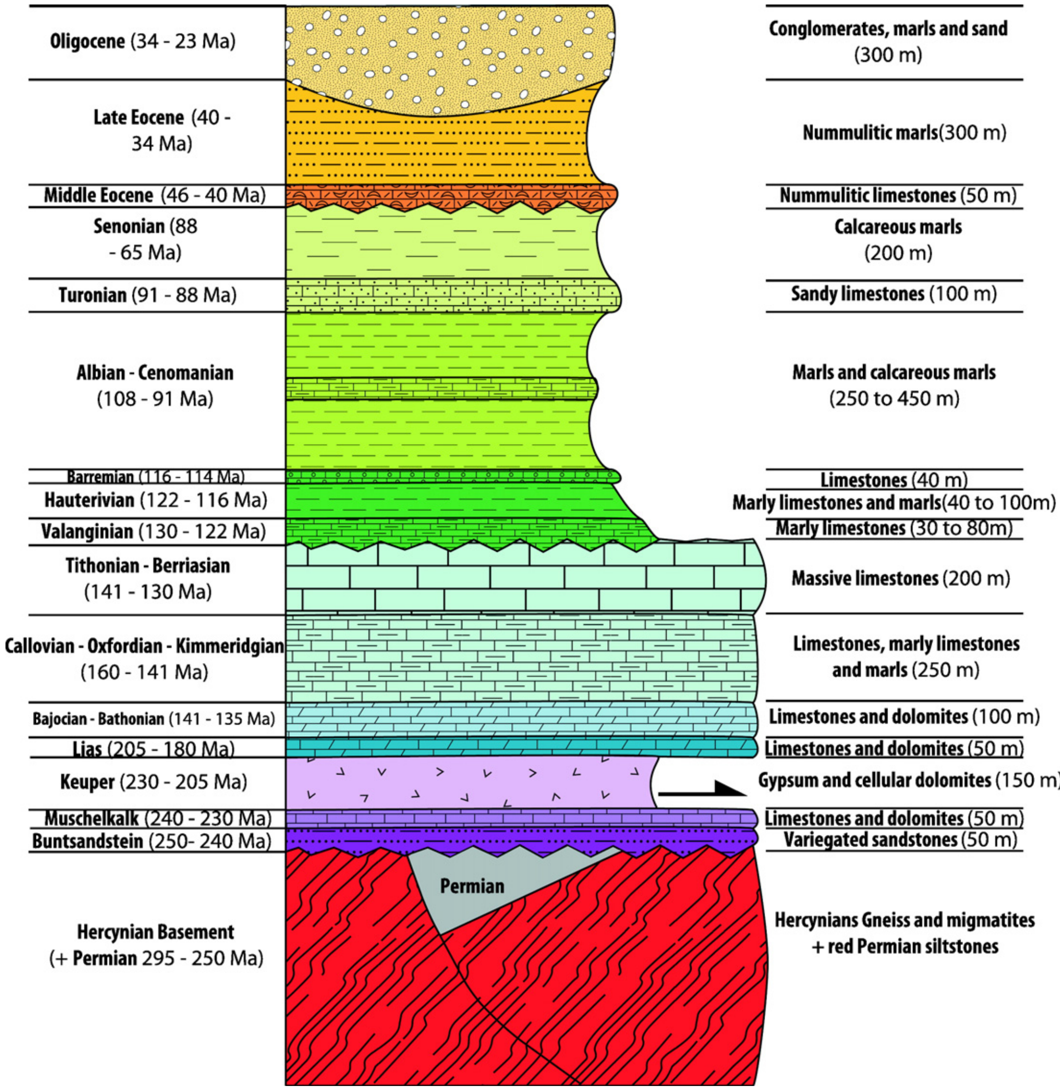

Fig. 2. Synthetic stratigraphic log of basement and post-Hercynian sediments in the study area (drawn after the 1:50,000 geological maps of Grasse–Cannes, Roquesteron and Puget-Théniers, Bonifay et al., 1970; Faure-Muret and Fallot, 1957; Goguel, 1980, respectively). Vertical scale is about 2 km, but the total thickness of Mesozoic–Cenozoic cover varies from 1 to 3 km.

3.1. General overview

3.1.1. Cross-section 1 (Tanneron–Dôme du Barrot)

On the northern part of the cross-section, the large synclines of La Rochette and Puget-Théniers are filled by a complete 1500 m-thick Mesozoic to Tertiary sedimentary sequence separated by a narrow anticline. The core of this anticline is made of Triassic series. The anticline itself has developed on an evaporate diapir which was active from the Lias to the Late Cretaceous, possibly as the result of the Liassic extension above a basement normal fault (Jackson and Vendeville, 1994). Thin-

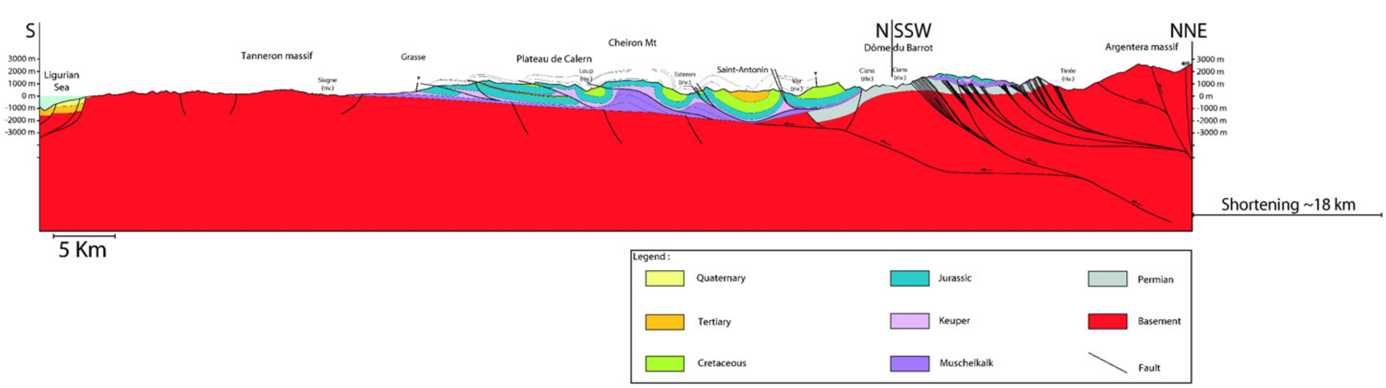

Fig. 3. Balanced cross-section from the Maures–Tanneron to the Argentera Massif with a thin-skin tectonic style (redrawn after Laurent et al., 2000). Pinpoints used to construct the balanced cross section are indicated.

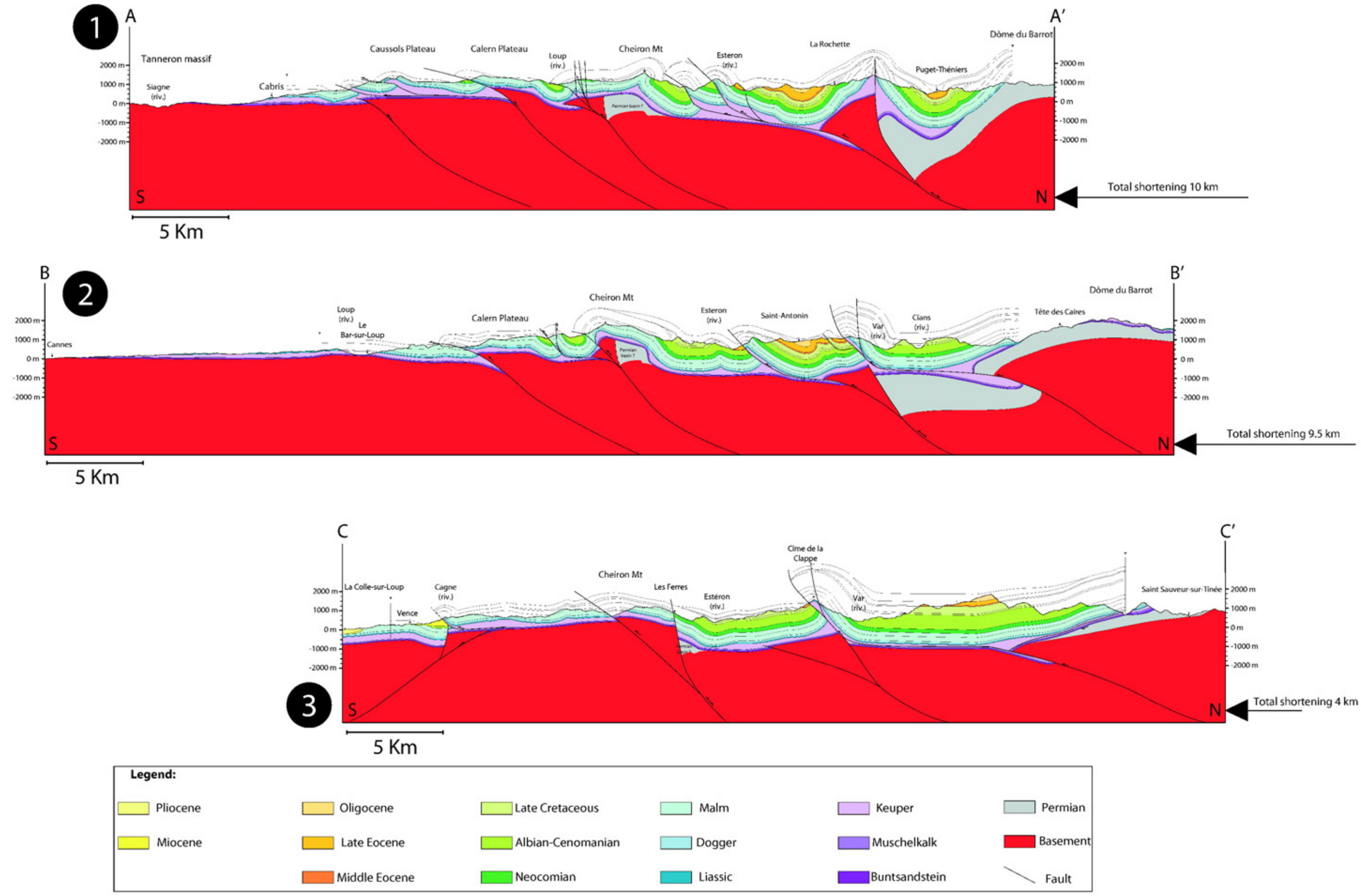

Fig. 4. Balanced cross-sections across the Arc of Castellane following a thick-skin assumption (this study, see Fig. 1 for location). Vertical:horizontal scale ratio is 1:1. Pinpoints used to construct the balanced cross section are indicated. See Fig. 2 for the color scale.

skinned thrusts south of La Rochette syncline connect down on a large north-dipping basement fault, which accommodates the Dôme du Barrot uplift. This basement thrust branches down on an ancient Permian normal fault, creating a basement short-cut beneath the northernmost anticline. The Cheiron Mountain, in the middle part of the cross-section, is the highest topographic peak south of the Argentera Crystalline Massif. The Cheiron Mountain is an overturned anticline with a south-dipping axial surface and is cored by Upper Jurassic limestones. Its high topography with respect to the surrounding areas can be explained by the inversion of a former Permian basin forming a box fold and a short-cut developed on a south-verging blind thrust beneath the Loup River. The exact basement geometry cannot be inferred directly from the surface geology, but the short wavelength of basement folding beneath the Cheiron is probably due to the inversion of a Permian basin, which should be more easily deformable than the underlying crystalline rocks. South of Cheiron, there is a succession of flat imbricated structures. The hanging walls are slightly deformed whereas the footwalls are marked by squeezed synclines with Cretaceous formations in the syncline fold bends. The deformation in the basement is accommodated by two thrusts with a moderate offset (100 to 500 m) beneath the Caussols and Calern plateaus, the latter accommodating the shortening of the two frontal peel thrusts (i.e., ~1000 m). The topography of this southern area is marked by flat-lying plateaus of massive limestones of Jurassic age, whereas Cretaceous formations are found in topographic depressions.

3.1.2. Cross-section 2 (Cannes–Dôme du Barrot)

To the north, a large Cretaceous to Tertiary basin is deformed by a series of asymmetric folds and by two thrusts, which branch northward in the décollement level connected to basement ramps. The innermost part of the depression beneath the Cians River is much shallower than beneath Puget-Théniers (cross-section 1, Fig. 4), suggesting that the horizontal shortening of the northern part of this cross-section is accommodated by a shallower, low-dipping basement thrust (instead of a basement fold). The Cheiron Mountain presents an asymmetric anticline developed on an inverted Permian basin and basement short-cut, like on the cross-section 1. It is worth noting that the topography is more important on this part of the Cheiron than on the other cross-sections, suggesting a larger amount of basement uplift. North of the Calern plateau, a narrow syncline cored by Upper Cretaceous formations is deformed by a flat-ramp system rooting down into the Cheiron basement thrust. Finally, the two frontal peel thrusts and the southernmost basement thrust visible on the western cross-section (beneath the Caussols plateau) do not exist, and the outermost structure is the flat-lying thrust affecting the sedimentary cover of the Calern plateau, itself underlain by a basement thrust. As in the westernmost cross-section, the cover is only slightly decoupled from the basement.

3.1.3. Saint-Paul–Argentera cross-section 3

On this cross-section, the Dôme du Barrot Permian structure does not crop out (3, Fig. 1), which might be due to a thinner Permian cover than in the west, or to an insufficient amount of thrusting as to uplift and bend up the Permian strata. In the basin located north of the Cime de la Clappe thrust, the whole sedimentary pile (from Triassic to Oligocene) is present, and is almost totally flat-lying with no internal deformations except along the southern thrust ramp. Along this ramp, a Cretaceous normal fault is reactivated and puts in contact the Liassic and Eocene formations with a vertical offset of ≈1500 m. South of it and north of the Cheiron Mountain, another Cretaceous normal fault (located near the village of Les Ferres, Fig. 1) marks the southern limit of a deep basin wherein all the Cretaceous sequence is deposited. This fault may have a very limited reverse reactivation because of its very important dip (70°) and the normal offset is still visible today in the field. South of the Cheiron, a thrust makes the Upper Triassic series crop out at the surface and shortcut the normal fault of Les Ferres. Finally, the southern frontal thrust system is interpreted as an inverted south-dipping normal fault associated with a basement short-cut developed on a Miocene basin. This easternmost cross-section shows fewer compressional structures than on the two western ones.

3.2. Detailed description of crustal shortening accommodation

Shortening in the basement and cover is accommodated by different structures. The plateaus of Calern and Caussols are uplifted by thrusts which crosscut the basement, ending up by flat-lying thrusts in the sedimentary cover, with little basement–cover decoupling. Deformation of these outermost domains is larger in the western cross-section than in the central one, where only one basement thrust located beneath the Calern plateau is present. The high dip of this thrust suggests that it may be a reactivated normal fault. The Caussols plateau (cross-section 1) displays two cover-scale thrusts with a moderate offset (<1 km), and two high-angle thrusts in the basement, beneath the Caussols and Calern plateaus.

Significant basement–cover decoupling is observed beneath the large Saint Antonin and La Rochette synclines, where cover and basement ramps are connected by a large flat in the décollement level. Here again, the cumulated offset is larger on the western section (>2000 m) than on the central one (<1000 m).

The Cheiron Mountain is interpreted as a box anticline underlain by a basement fold developed on an inverted Permian basin, where the former north-dipping normal fault has been shortcut by a major south-verging thrust. The inferred Permian half-graben is intensely deformed beneath the northern limb of the Cheiron anticline on the central and western cross sections, whereas it is only moderately folded on the easternmost one. Near Les Ferres, on the cross-section 3, a normal fault is still visible at this place and seems to have been reactivated during the Cretaceous extensional episode. The existence of a deep basin filled with Cretaceous sediments visible on the three cross-sections north of the Cheiron Mountain suggests that the normal fault of Les Ferres (Fig. 4C), which is not outcropping on cross-sections 1 and 2, possibly gives place westward to a gentle flexure of the basement. Basement shortening is also accommodated by short-cuts visible south of the Dôme du Barrot structure and north of Vence, due to the deformation of tilted blocks developed on Permian and Miocene normal faults and basins, respectively. The northernmost short-cut, visible on the central and western cross-sections, accommodates large basement uplift. On the western cross-section, Upper Triassic rocks crop out in a pinched anticline located on a Triassic diapir above the trace of a Permian normal fault (Jackson and Vendeville, 1994), whereas on the central cross-section a ramp thrust is observed on the top of this fault. This short-cut is not visible on the eastern cross-section where the normal offset of the ancient fault is only partially inverted beneath the Cime de la Clappe ramp (Fig. 4). Finally, uplift of the Dôme du Barrot and Argentera massifs is taken up by a large basement thrusting on the eastern and central cross-sections, with more intense folding of Permian strata beneath the western one.

3.3. Restoration method and shortening estimates

The shortening estimates (Fig. 5) were made assuming a conservation of the layer length since there is no evidence for internal layer deformation, except in the Upper Triassic evaporitic formations. The time stages of these restored cross-sections were chosen to reflect the major tectonic events unraveled by regional unconformities, and to understand how structural inheritance may have controlled the compressional structure localization. For the two western cross-sections, we define three restoration stages, starting before the Oligocene in order to unfold the main Alpine deformations (Fig. 5A and B). The easternmost cross-section has a first restoration step in the Lower Miocene to highlight the choice made on the most frontal thrust and the importance of Lower Miocene extension in the stepping down of the margin to the south of the section (Fig. 5C). For all cross-sections, a second restoration step is made between the Late Eocene and the Lower Cretaceous in order to shed light on an extensional phase, which occurred during the Lower Cretaceous and caused significant variations in the sediment thickness. This phase was followed by a Late Cretaceous compressional event linked to the Pyrenean collision, which created asymmetric long wavelength folds. Those folds were then sealed by the Eocene deposits that rest unconformably on the underlying Upper Cretaceous sequences. Finally, the last restoration step was drawn for post-Lower Cretaceous times, and allows us to estimate the overall structure of this region before the occurrence of the first compressional episode.

Prior to the onset of the Pyrenean compression, the crystalline basement progressively steps down toward the north due to several north-dipping normal faults (Fig. 5). The western and central cross-sections (1 and 2) are characterized by the presence of two main faults, one in the foreland with a moderate offset and another one in the hinterland, the latter being presumably the ancient fault that accommodated the subsidence of the 2000 m-thick Dôme du Barrot Permian basin (Fig. 5A and B). On the cross-section 2, the normal fault located in the hinterland accommodates important thickness variations of Early Cretaceous sediments, which indicate that it was active as a normal fault until the Cenomanian (Fig. 5B). The eastern cross-section displays two Cretaceous normal faults located in the central part of the domain, which accommodate a northward increase of Early to Middle Cretaceous sediment thickness (Fig. 5C).

The Late Cretaceous stage is characterized by a very small amount of shortening (200 to 600 m) in this formerly extended crust, with the development of wide folds in the sedimentary cover associated with a gentle buckling of Permian sediments. To explain this geometry, we propose that compression causes basin inversion and folding of Permian strata against the Dôme du Barrot normal fault. During this stage, the sedimentary cover is locally decoupled from the underlying basement and the ductile décollement layer flows beneath the cover anticline hinges.

Finally, the Oligocene stage, which corresponds to the onset of the Alpine compression (s.s.) is characterized by an increasing amount of compressional strain (200 to 1000 m), which causes a tightening of Late Cretaceous folds and a more intense basement buckling and faulting. Oligocene deformation is restricted to the northern part of the cross-sections, the foreland domain being almost undeformed. On the easternmost cross-section (Fig. 5C), the Lower Miocene stage is characterized by the development of a south-dipping normal fault to the south, in response to the Ligurian basin rifting stage.

We calculated total shortening values after the Oligocene–Lower Miocene in order to quantify the deformation due to the inversion and exhumation of the Argentera–Mercantour Massif during the Alpine episode. On the cross-section 1, the total estimated amount of shortening is about 10 km, with ~600 m inherited from the post-Cretaceous (i.e., Pyrenean) stage and 1000 m from the Oligocene (i.e. early Alpine) stage (Figs. 4 and 5A). On the cross-section 2, a total shortening of 9.5 km is decomposed as follows: 200 m during the Late Cretaceous and 600 m during the Oligocene (Figs. 4 and 5B). On these two cross-sections, the shortening ratio is therefore 21%.

Finally, on the easternmost cross-section the amount of shortening is only about 4 km (shortening ratio of 9%, Figs. 4 and 5C). Therefore, there is an important decrease in the amount of compressional strain from the central cross-section to the eastern one. At first sight, there are indeed fewer compressional structures to the east than to the west. The structural map (Fig. 1) shows only three reverse faults in the eastern part of the Castellane thrust-and-fold belt while there are eight or nine of these in the western part.

4. Discussion

4.1. Related geodynamic evolution

The balanced and restored cross-sections analyzed in the light of field data, together with the review of previously published studies in the southwestern Alps allow us to highlight four tectonic events in the Early and Late Cretaceous, in the Oligocene, and in the Early to the Late Miocene. The two first events are likely related to the extensional and compressional events linked to the "Pyrenean orogenic cycle": the Lower Cretaceous extensional tectonic event can be connected to the Pyrenean continental rifting to the southwest and to the Valais domain opening in the northern Alps (Handy et al., 2010; and references therein) although the Valais domain may be Jurassic in age (Beltrando et al., 2007; Manatschal, 2004). This Mesozoic extensional phase has resulted in a strong structuration of the basement in the future southwestern Alps, and has resulted in the Triassic décollement level offset by several hundreds of meters due to normal faulting. At least two of the Cretaceous normal faults are probably reactivated Permian normal faults, as shown on the cross-sections 1 and 3 (Fig. 4). This extension was followed during the Late Cretaceous by a compressional event, which formed long wavelength folds that were also mapped in surrounding areas such as the Sainte-Victoire Mountain in Provence (Espurt et al., 2012), and locally resulted in significant erosion of the Mesozoic cover (e.g., Ford, 1996).

The third and fourth events are related to the main Alpine compressional event in the southwestern Alps. During the Oligocene only moderated shortening is recorded and coeval with the deposition of foreland basins (e.g., Ford et al., 2006). The last and main deformation episode occurred during the Miocene together with the Argentera–Mercantour ECM exhumation, essentially in two stages (Sanchez et al., 2011a, 2011b): from 26 to 20 Ma ($^{40}Ar/^{39}Ar$ on phengite ages), and since 12.0 Ma (fission-track and (U–Th)–He apatite ages). This ECM is a structural wedge made of crystalline basement, which has been thrust underneath the sedimentary cover and over the foreland basement (Sanchez, 2010). Against this rigid back-stop, deformation propagated southward during the Miocene, and is well-expressed in syn-tectonic basins, such as the Roquebrune basin, east of the Nice Arc (Giannerini et al., 2011). The shortening estimate in the foreland and the geometrical reconstructions (Fig. 4) are in agreement with an uplift of 4 to 5 km on the frontal blind thrust bounding the Argentera–Mercantour ECM. It

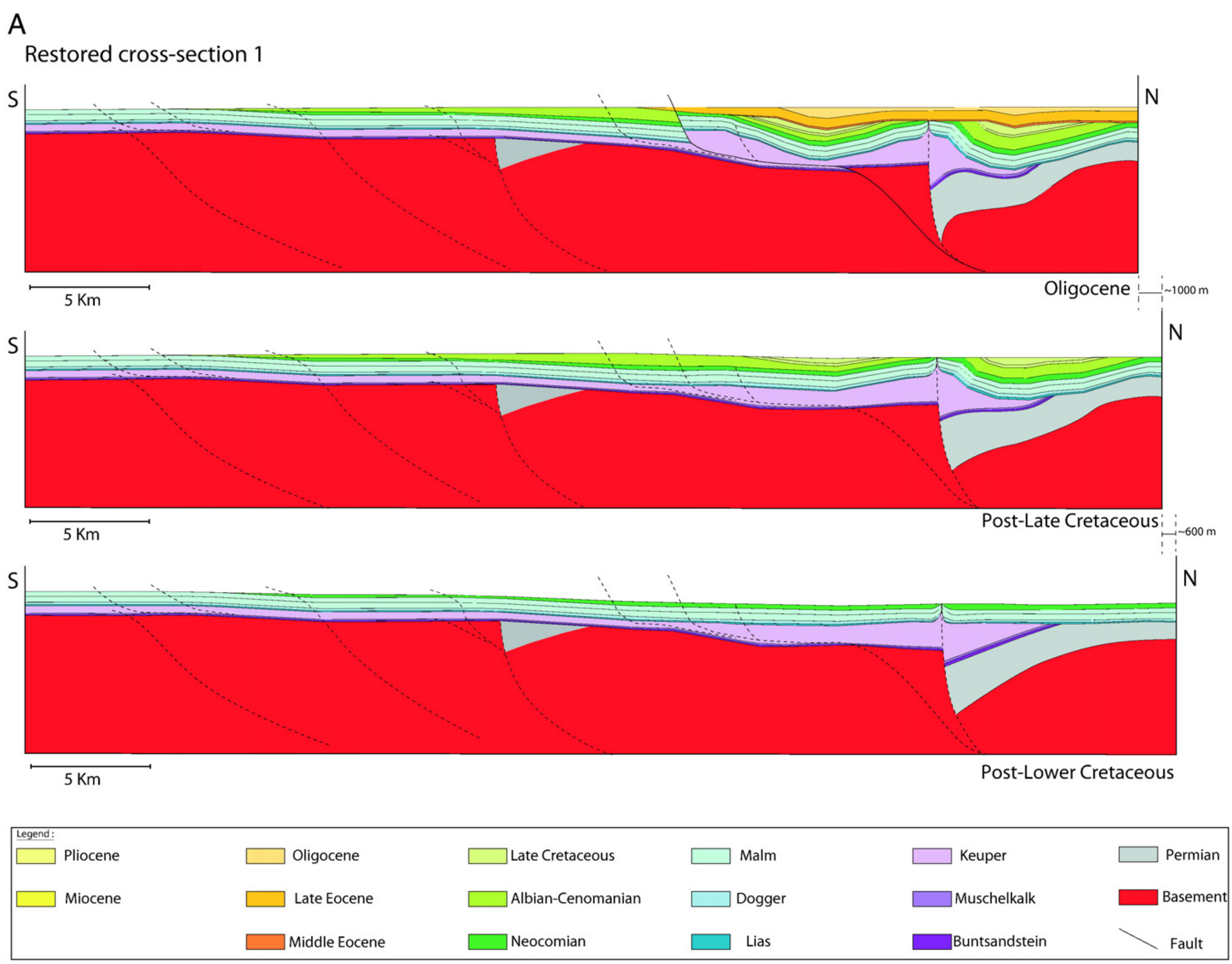

Fig. 5. Restored cross-sections at three different stages for the three balanced cross-sections presented in Fig. 4. A, Restoration for the western cross section at the Lower Cretaceous, Late Cretaceous and Oligocene (1, Fig. 4); B, similar restoration stages for the central (2, Fig. 4) cross section, the same legend as A; C, restored cross sections at the Lower Cretaceous, Late Cretaceous and Lower Miocene for the eastern cross-section, the same legend as A. The present-day topography is not indicated.

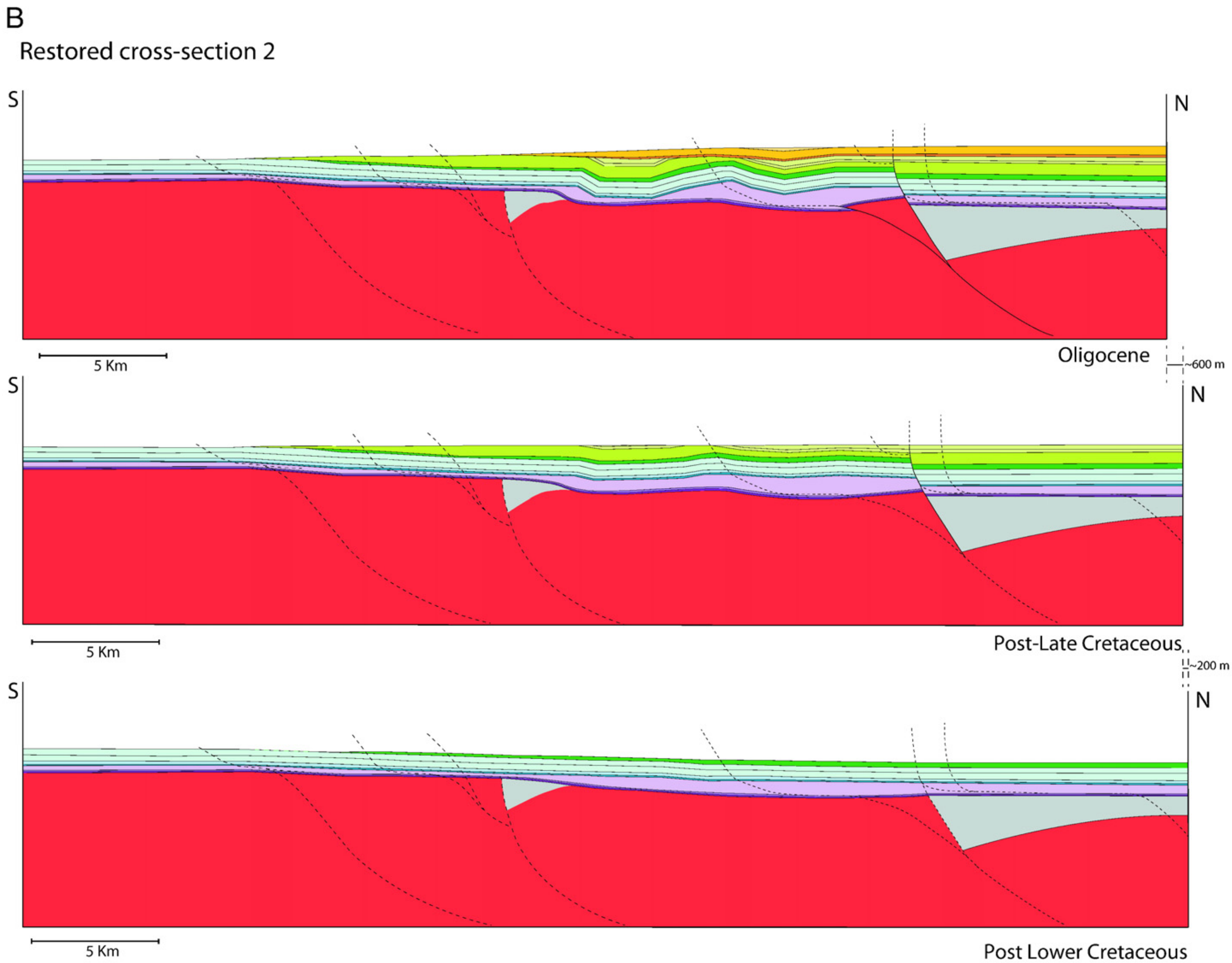

Fig. 5 (*continued*).

accounts for a third of the exhumation of the internal part of the Mercantour ECM, which was buried at about 15 km at 33–26 Ma (Sanchez et al., 2011b).

In this period of time, the continental break-up and opening of the Ligurian Sea due to the counterclockwise rotation of the Corsica–Sardinia block took place and led to the formation of south-dipping normal faults on the thinned Provence margin (e.g., Jolivet and Faccenna, 2000; Réhault et al., 2012; Sage et al., 2011). Finally, during the Late Miocene, the emplacement of the Argentera–Mercantour ECM close to the surface (Sanchez et al., 2011a) was accommodated by large-scale basement thrusting (Sanchez, 2010) and led to the folding of the sedimentary cover and to the inversion of inherited normal faults in the foreland.

4.2. Structural style of the Castellane fold-and-thrust belt

4.2.1. Amounts of shortening and implications for the reactivation of inherited pre-Alpine structures

Based on previous concept of a major basement–cover decoupling below the Castellane fold-and-thrust belt (e.g., Labaume et al., 1989; Ritz, 1992) Laurent et al. (2000) estimated an amount of shortening in the cover of 18 ± 1.5 km, and assumed that the accommodation of this shortening in the basement was taken up on a large-offset thrust rooting down beneath the Argentera–Mercantour Massif and the Dôme du Barrot. According to this interpretation, the décollement level is very efficient and all the reverse faults observed in the sedimentary cover are not connected in the basement immediately beneath. Reactivation of old normal faults, which have structured the region in Permian, Liassic and Early Cretaceous times (Dardeau, 1988; Dardeau and De Graciansky, 1987) was not regarded as significant events affecting the Mesozoic sedimentary cover. However, the choice of a thin-skinned tectonic style in the Castellane fold-and-thrust belt presents some problems: (i) the cumulated estimated shortening is rather high (up to 30% rather than 10 to 27% in the rest of the Alps, see Fig. 6 and Bellahsen et al., 2014) and a large part of it (~40%) is accommodated by the most frontal thrust near the city of Grasse (Fig. 3), where it accommodates a 2-fold increase of the sedimentary cover thickness; (ii) prior to the onset of the compressional deformation, the Triassic décollement level was offset by a large number of Mesozoic normal faults and can hardly be interpreted as a continuous weak level during later compression, its activation as a flat and continuous surface being thus highly improbable on a simple geometrical basis; and (iii) under the Cheiron Mountain (Fig. 3), the high topography is explained by a 34-fold tectonic overthickening of the Early Triassic units (which are

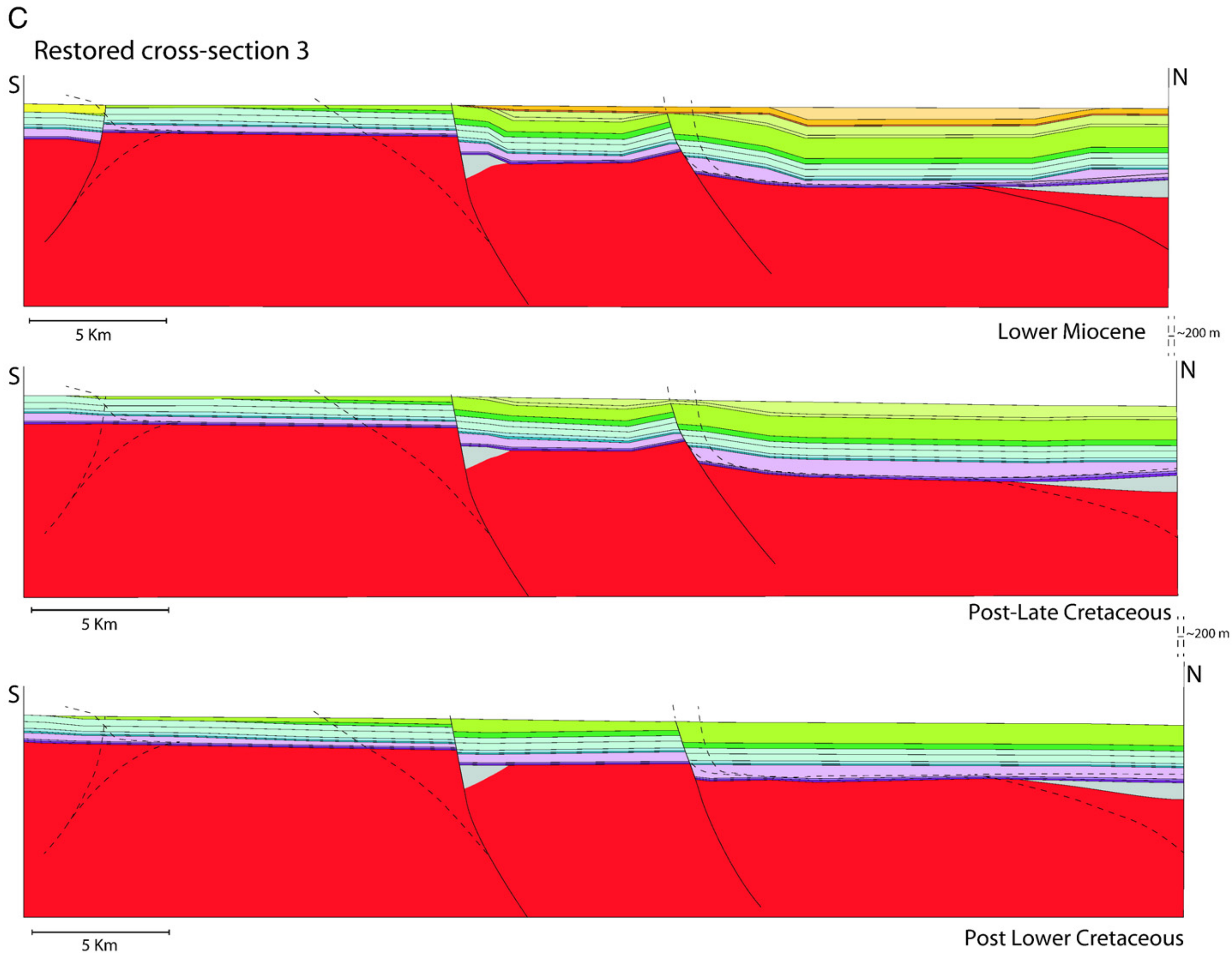

Fig. 5 (*continued*).

made of limestones and sandstones mechanically coupled to the basement) from 50 m (initial thickness) up to 1700 m (Fig. 3). More generally, in Laurent et al. (2000), the basement top is represented by a flat-lying surface, which seems unlikely because of all the extensional tectonic events that marked this region from the Permian rifting (Lapierre et al., 1999) to the Miocene Ligurian Sea opening (e.g. Jolivet and Faccenna, 2000; Sage et al., 2011). Therefore, several problems arise from the thin-skin tectonic deformation model in terms of mechanical and structural consistencies.

In the present paper, the cross-sections are drawn in a thick-skin mode taking into account an irregular basement–cover interface, as due to normal fault offsets during rifting phases. Such thick-skin deformation in subalpine domains was already proposed by several authors in other regions (Bellahsen et al., 2012; Coward, 1996; Espurt et al., 2012; Tozer et al., 2002). The calculated shortening ratio ascribed to the Alpine (Miocene) deformation on the westernmost cross-sections is about 20%, while amounts of shortening increase from the western and southwestern Alps toward the Central Alps (e.g. Bellahsen et al., 2014; Ford and Lickorish, 2004; Kempf and Pfiffner, 2004; Sinclair, 1997). That amount is very close to other shortening ratios estimated in External Western Alps (Fig. 6) in the northern Mont Blanc massif (Burkhard and Sommaruga, 1998), the Bornes massif (Affolter et al., 2008), and the Vercors–Oisans massifs (Bellahsen et al., 2012; Deville et al., 1994; Phillipe et al., 1998), where the shortening ratios never exceed 27%. Similarly, in Provence Ventoux–Lure area, the amount of Oligocene–present (Alpine) shortening is 6–11% (Ford and Stahel, 1995; Fig. 6), while it can be neglected in the Sainte-Victoire further south (Espurt et al., 2012).

In the Jura, Bornes, Bauges, and Vercors massifs, the structural style is thin-skinned and the cover shortening is transferred at depth to inner-more basement thrust faults (Bellahsen et al., 2012; Deville et al., 1994; Phillipe et al., 1998; Rossi and Rolland, 2014), except in the Jura where recent basement thrusts were activated (see Lacombe and Mouthereau, 2002, and references therein). On the contrary, the Provence belts around the Sainte Victoire area (Espurt et al., 2012; Lacombe and Jolivet, 2005) and the Castellane fold-and-thrust belt (this study) present reactivated basement thrusts all across their length and especially close to their front. In summary, our shortening estimates are lower than those previously published in the same area (32%) by Laurent et al. (2000) (see Fig. 3), but seem consistent with other shortening estimates in the External Alps (Fig. 6). The discrepancy between the results of this study and the one in Laurent et al. (2000) comes mainly from a different interpretation of the structure of the frontal thrust at depth.

In our study, shortening ratios tend to increase from east to west from about 9% to about 21%. This variation is partly due to an increasing contribution of the Pyrenean shortening phase from east to west, but this has to be rather small. Indeed, this contribution is minor in the

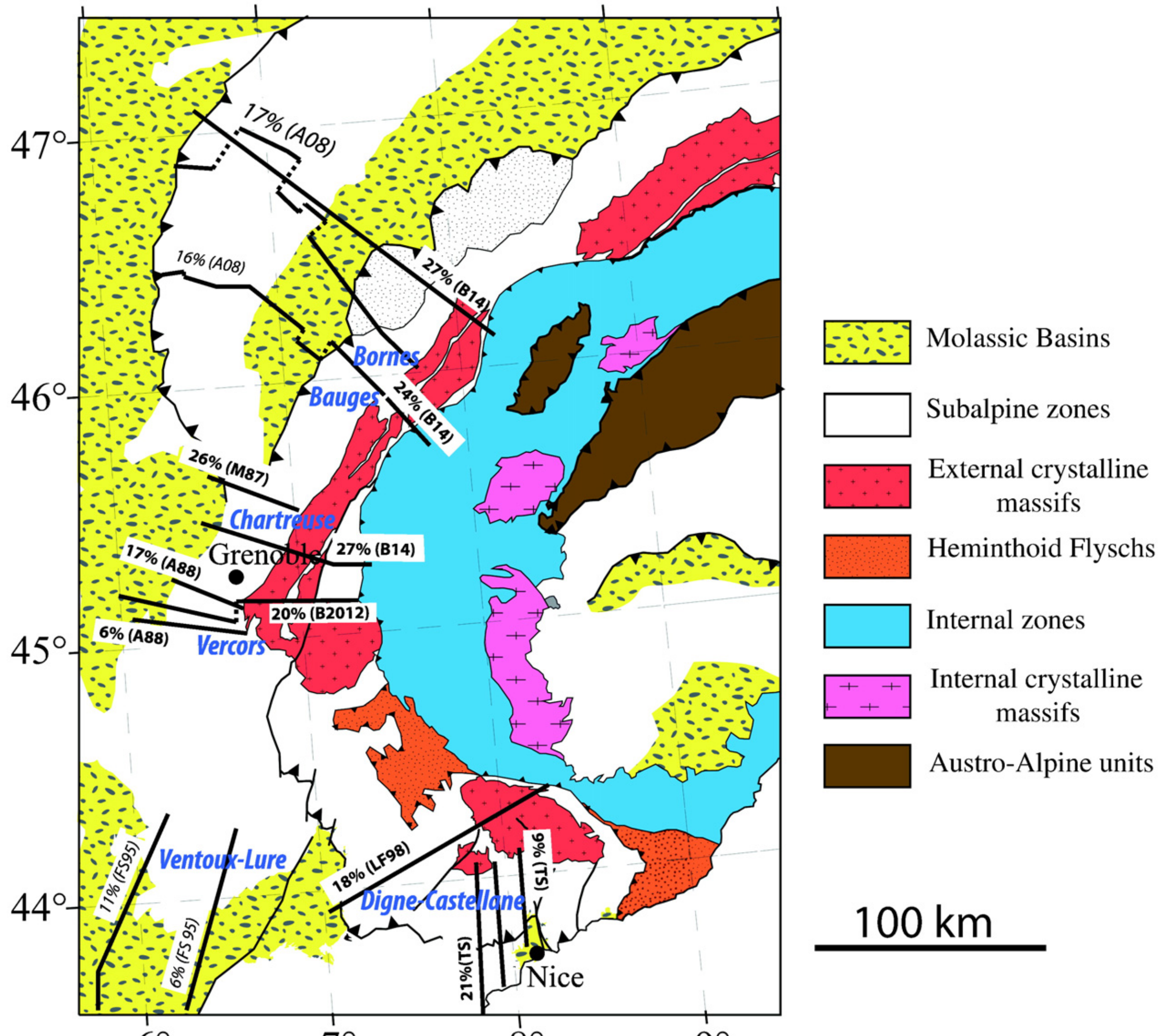

Fig. 6. Synthesis of shortening ratios and modes of deformation in the subalpine chains (after Affolter et al., 2008; Arpin et al., 1988; Bellahsen et al., 2012, 2014; Ford and Stahel, 1995; Lickorish and Ford, 1998; Mugnier et al., 1987; and this study). Italic/bold letters indicate a dominant thin-skin or thick-skin deformation mode, respectively.

Castellane fold-and-thrust belt (less than 2 km), where most of the shortening is post-Oligocene in age, while in the Sainte-Victoire area the 25–34% shortening rate is ascribed only to the Pyrenean phase (Espurt et al., 2012). Moreover, this lateral variation also coincides with a change in pre-Alpine extensional setting as expressed by the reconstructed Permian rift thickness, which significantly decreases from west to east. This observation is in agreement with the opening of Permian rifts as pull-apart basins along the Pangea margin (e.g., Arthaud and Matte, 1977; Ziegler and Stampfli, 2001). Furthermore, it is likely that the Pyrenean extension stage has enhanced this Permian configuration (Lagabrielle et al., 2010). During the Mesozoic, Western Alps and French South-East basin may be regarded as a transfer domain between two extended domains (the Pyrenean and the Valais domains). The following phase of Pyrenean N–S compression resulted in an increased inversion toward the west, and this inversion was significantly less in the Western Alps especially east of the Castellane fold-and-thrust belt.

4.2.2. Geodynamic significance of the thick-skin deformation mode

While many foreland thrust-and-fold belts have long been interpreted as thin-skin structures, some recent works suggest that inherited basement structures are often reactivated during recent compressional episodes. This has been demonstrated in the Pyrenean–Provençal orogenic system, in southwestern Alps, in Zagros and in Taiwan for instance (e.g., Lacombe and Mouthereau, 2002; Roure and Colletta, 1996, and references therein; Filleaudeau and Mouthereau, 2011; Jackson, 1980). Thick-skin tectonic deformation of foreland systems can occur because of several factors, among which: an inefficient, discontinuous or inexistent décollement level, a mechanical weakening of fault zones in the basement due to fluid–rock interactions (e.g., Wibberley, 2005 and references therein), and/or a thermal weakening of the crust inherited from the last thermal event (Mouthereau et al., 2013) or due to syn-orogenic tectonic burial (Bellahsen et al., 2012). More generally, a recent compilation of collisional orogens worldwide shows a clear difference between "recent" (i.e., Phanerozoic) lithospheres characterized by thick-skin tectonics and a low amount of strain (<35%) and old cratonic keels where the deformation obeys a thin-skin mode and where shortening ratios can reach more than 70% (Mouthereau et al., 2013), although tectonic modes might also be influenced by other factors such as the sedimentary cover thickness.

In the Alpine and Pyrenean domains, it is noteworthy that both thick- and thin-skinned structural styles have been deciphered. Thick-skinned styles in foreland belts are found in southern belts, i.e. South-

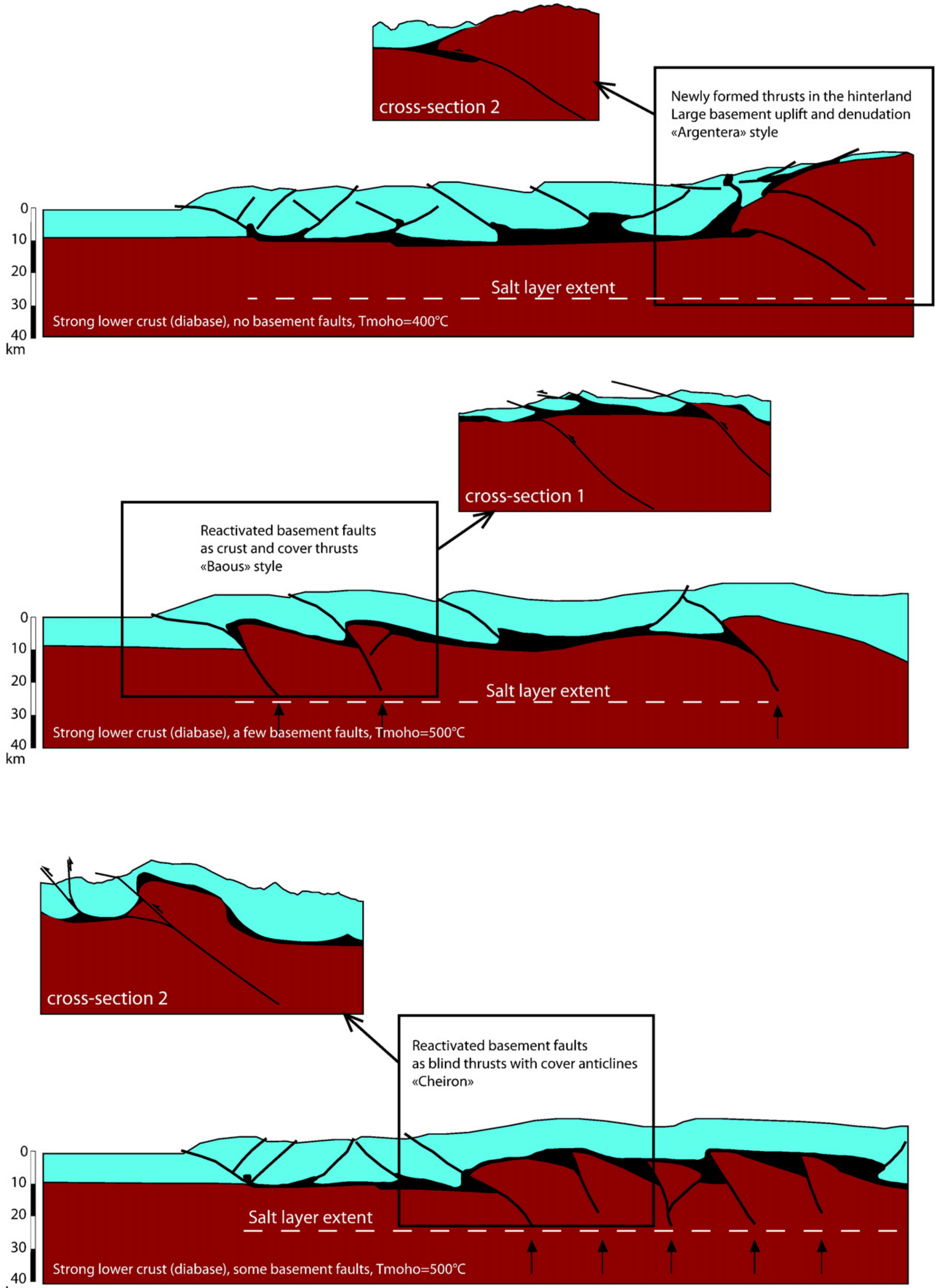

Fig. 7. Simplified interpretation of three numerical models by Nilfouroushan et al. (2013) with different Moho temperatures, initial salt layer extent (dashed white line) and number of pre-existing basement faults (vertical arrows). Insets show details of the cross-sections (Fig. 4) that compare well with these models.

West Alps and Provence domain. In both cases, these belts were activated just after a significant stage of crustal thinning: the Pyrenean Albo-Cenomanian rifting and the Ligurian Miocene rifting. Moreover, the Ligurian rifting is linked to the African slab roll-back and back-arc calc-alkaline volcanism, both offshore and onshore Provence (Jolivet and Faccenna, 2000; Réhault et al., 2012). In this back-arc context, the European crust was probably heated and weakened by fluid and magma circulation. This may explain the thick-skinned structural style that affected this weak crust.

A comparison with recently published numerical models of fold-and-thrust belt evolution with a salt layer and reactivated basement faults (Nilfouroushan et al., 2013) explains satisfactorily the different compressional structures visible on our cross-sections (Fig. 7), although none of the models was specifically set up to correspond to the geological setting of the Castellane region. The uplift of the Argentera–Mercantour Massif is compatible with the activation of high-angle thrusts in the hinterland, visible on a numerical model where the Moho temperature is low (400 °C, Fig. 7 top section). Reactivation of inherited normal faults in the foreland domain associated with a strong coupling between the cover and the basement occurs when the lower crust is weak enough to deform in a ductile manner, i.e. with a higher Moho temperature (500 °C, Fig. 7B middle and lower sections). This result compares well to the deformation observed in front of the Baous and plateaus of Caussols and Calern in the outermost part of our cross-sections. Finally, for a similar Moho temperature, reactivation of basement normal faults in the central part of the models is responsible for a large antiformal uplift of the cover underlain by several blind basement thrusts in a similar way as the Cheiron structure (although the models do not consider possible ancient basin inversion). Compared to these models, our cross-sections advocate for: (i) a rather strong lower crust with a Moho temperature colder in the hinterland than in the foreland, which is compatible with a recent heating due to the opening of the Ligurian basin to the south; and (ii) a pervasive inherited basement fabric of south-dipping normal faults that are reactivated as thrusts during the Alpine compressional episode (Fig. 4). This fabric is consistent with the existence of an E–W Permian basin between the Maures and Tanneron massifs and with numerous evidences of Early Cretaceous N–S extension.

5. Conclusion

Three balanced cross-sections have been drawn across the Castellane fold-and-thrust belt in south-west Alps, which are used to constrain the amount and style of deformation of multiple shortening events at the junction of Pyrenean and Alpine fold belts. The reconstructed geometries allowed us to propose a dominantly thick-skinned structural style for this fold-and-thrust belt, accounting for 9–21% of horizontal shortening, mainly during the Miocene to Pliocene times. These shortening ratios are consistent with other shortening estimates around the Alps fold-and-thrust belt arc. Alpine shortening was accommodated by tectonic inversion of Permian/Lower Cretaceous normal faults, as high-angle thrusts, inversion of Permian basins and short-cuts.

Restored cross-sections highlight two different compressional deformation stages:

(1) A post-Late-Cretaceous stage concomitant with the onset of the Pyrenean orogeny is characterized by wide folds in the sedimentary cover, which are overlain with a slight angular unconformity by Tertiary sediments.
(2) Alpine (i.e., Miocene) deformations are much more significant, and result in a tightening of previously formed Pyrenean structures, low-angle thrusting in the sedimentary cover and high-angle basement fault inversion, or basin inversion and short-cuts.

Estimated cumulated amounts of shortening range from 21 to 9%, respectively from west to east, which is only partly explained by a lateral decrease of Upper Cretaceous–Eocene Pyrenean shortening toward the east. This domain of the Alps foreland was thus preserved during the Pyrenean phase and only deformed in the Miocene, while the Provence domain to the west underwent significant Pyrenean compression and very little Alpine compression. In contrast, the Nice Arc to the east underwent polyphased and orthogonal deformation during the Alpine phase, in the Oligocene and Miocene times.

A comparison with recently published numerical models of fold-and-thrust belt evolution with a salt layer and reactivated basement faults explains satisfactorily the different compressional structures visible on our cross-sections. This comparison agrees for a rather strong lower crust with a Moho temperature colder in the hinterland as opposed to in the foreland, which is compatible with a recent heating due to the opening of the Ligurian basin to the south, and for a significant basement anisotropy inherited from Permian and Cretaceous extensional phases.

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