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1 **Structural evolution of the superimposed Provençal and** 2 **Subalpine fold-thrust belts (SE France)**

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10 **Abstract**

11 Highlighting how crustal shortening is accommodated in space and time in fold-thrust belts is
12 a major issue for understanding the long-term tectonic evolution of orogenic systems. In this
13 study, we combine surface and subsurface data to build a 150 km-long sequentially restored
14 balanced cross-section across two superimposed foreland fold-thrust belts in southeastern
15 France: the upper Cretaceous-Eocene Provençal belt and Oligocene-Neogene Subalpine belt.
16 These belts are composed of inverted Paleozoic-Mesozoic basins with Mesozoic halokinetic
17 structures, basement thrusts, and cover thrusts-nappes detached in the Triassic evaporites. The
18 multiphase growth of the Provençal and Subalpine belts has been constrained along the
19 studied cross-section by syn-orogenic deposits and kinematic indicators of thrusting. The pre-
20 orogenic palinspastic reconstruction of the cross-section to Late Santonian shows a large
21 uplift zone in the center of the section (Durance and Valensole highs), which led to the
22 separation of the Beausset basin (South Provence basin) to the south from the Barles-Digne
23 basin (Vocontian basin) to the north. The Provençal shortening propagated ~NNE-ward from
24 the Beausset basin up to folds in the Barles area during the Latest Santonian to Eocene times.
25 Shortening value reaches 38 km and has been mainly accommodated by the inversion of the
26 South Provence basin-Durance high. The Subalpine shortening propagated ~SW-ward from
27 the Barles-Digne basin up to the Mediterranean coast during the Oligocene-Miocene to
28 Quaternary times. It reaches 35 km and has been mainly consumed by the inversion of the
29 thick Digne basin. The Provençal thrust wedge is characterized by distributed basement
30 thrusts reworking numerous structures inherited from the Variscan belt and Permian-

31 Mesozoic rifts. This structural style might have favored the development of confined foreland
32 basins, as the Arc and Rians basins. In contrast, the vertical stacking of the thick Digne Nappe
33 and Barles basement triangle zone in the Subalpine thrust wedge might have controlled the
34 large flexure of the Valensole foreland basin. The lack of Triassic evaporites in the Valensole
35 high probably explains that shortening was not transferred into the cover of this domain.
36 Consequently, the Provençal then the Subalpine shortenings might have been transferred more
37 deeply to induce the reactivation of basement faults into the external zones. The Mesozoic
38 halokinetic structures also strongly influenced the location of contractional deformation. This
39 study highlights that the crustal structural inheritances influenced the structural styles and
40 development of extensional basins and subsequent Provençal and Subalpine belts.

41 **Keywords**

42 Superimposed orogens; Thrust tectonics; Cross-section balancing; Sequential restoration;
43 Structural inheritance; Provence-Alps

44 **1. Introduction**

45 The structural evolution of fold-thrust belts is systematically controlled by crustal
46 heterogeneities (e.g., rheology, crust-mantle coupling and crustal thickness) which are mostly
47 inherited from previous tectonic stages of extensional, compressional, or strike-slip
48 deformation (e.g., Lutaud, 1957; Angelier and Aubouin, 1976; Zubieta-Rossetti et al., 1993;
49 Coward, 1996; Roure and Colletta, 1996; Beauchamp et al., 1999; Branquet et al., 2002; Kley
50 and Monaldi, 2002; Souquet et al., 2003; Mora et al., 2006; Alvarez-Marron et al., 2006;
51 Vergés et al., 2007; Espurt et al., 2008; Malavieille and Konstantinovskaya, 2010; Bellahsen
52 et al. 2012; Hinsch, 2013; Mouthereau et al., 2013; Tavani et al., 2015; Lacombe and
53 Bellahsen, 2016; Bestani et al., 2016; Butler, 2017; Calderon et al., 2017; Espurt et al., 2019a;
54 Jourdon et al., 2019, 2020; Spitz et al., 2020; Martín-González et al., 2021; Mouthereau et al.,
55 2021). These elements generate anisotropic behavior of the crust during contractional
56 deformation in the Provence, where the succession of geological events generated a complex
57 structural framework since Paleozoic (Matte, 2001; Lacombe and Jolivet, 2005; Guillot and

58 Menot, 2009; Schreiber et al., 2011; Advokaat et al., 2014; Tavani et al., 2018; van
59 Hinsbergen et al., 2020; Romagny et al., 2020; Angrand and Mouthereau, 2021). In this paper
60 we have chosen to study and discuss the anisotropic behavior of Provence crust and its role on
61 kinematic evolution of superimposed Provençal and the Subalpine fold-thrust belts (Fig. 1;
62 Lemoine, 1972; Siddans, 1979).

63 The Provençal and southwestern Subalpine foreland thrust wedges have contrasting thrust
64 vergences and time emplacement as recorded by structural data and well preserved syn-
65 tectonic foreland strata. Both thrust wedges are characterized by a far-field transmission of the
66 compressional stress into the most external zones that is still poorly characterized and not yet
67 quantified because of complex overlap of extensional and compressional deformations (e.g.,
68 Baudemont, 1985; Tempier, 1987; Delfaud et al., 1989; Guieu and Roussel, 1990; Philip,
69 2012; Bestani et al., 2016; Tavani et al., 2018). Thrust-fold structures related to the Provençal
70 compression develop at more than 180 km to the north from the Provence crystalline massifs
71 (Fig. 1). They were described in the Valensole foreland basin (Dubois and Curneille, 1978), in
72 the Castellane arc (de Lapparent, 1935; Goguel, 1937; Kerckhove and Antoine, 1963; Roux,
73 1968; Roux, 1970; Campredon and Giannerini, 1982), in the Barles area (Lemoine, 1972;
74 Haccard et al., 1989b), and in the Baronnies-Dévoluy area (Gidon et al., 1970; Lemoine,
75 1972; Gidon and Pairis, 1976; Fabre et al., 1985; Montenat et al., 2005). Likewise, the
76 occurrence of deformed Neogene to Quaternary rocks in southern Provence and the seismicity
77 data, suggest far-field reactivation of inherited structures more than 100 km to the south to the
78 Subalpine thrust front and as far as the Ligurian continental shelf (Guieu, 1968; Angelier and
79 Aubouin, 1976; Weydert and Nury, 1978; Dupire, 1985; Hippolyte et al., 1993; Champion et
80 al., 2000; Baroux et al., 2001; Chardon et al., 2005; Cushing et al., 2008; Béthoux et al., 2008;
81 Larroque et al., 2021; Fig. 1).

82 Low-temperature thermochronology methods which are used to constrain the chronology
83 of thrust propagation as in many fold-thrust belts (e.g., Ege et al., 2007; Jolivet et al., 2007;
84 Espurt et al., 2011; Parra et al., 2012; Labaume et al., 2016; Rat et al., 2019; Hernandez et al.,
85 2020), are poorly applicable in the study area because the Provençal and Subalpine thrust
86 wedges have experienced insufficient Cenozoic burial and the exhumation history of the inner
87 crystalline basement of the Provençal belt has been erased by the thermal event related to the
88 Oligocene-Miocene Ligurian-Provençal rift (Lucazeau and Mailhe, 1986; Machhour et al.,
89 1994; Morillon, 1997; Jakni, 2000; Bestani, 2015). The study of syn-tectonic strata with
90 growth stratal pattern deposited during contractional deformations remains thus essential for
91 reconstructing the chronology of successive deformational events, thrust propagation, and
92 foreland basin evolution (Vergés and Munoz, 1990). This approach can be combined with
93 fault slip analyses to constrain deformation regimes, thrust tectonic transport directions and
94 particularly late fault reactivation in younger and in appearance undeformed foreland strata
95 (Angelier, 1984, 1991). Balanced cross-section method has been used to restore
96 compressional tectonic structures to their initial stages and the geometry of the basins before
97 their deformation (Dahlstrom, 1969; Thomas and Bayona, 2002). Integration of precise timing
98 of thrusting provided by syn-orogenic deposits with cross-section balancing can be a powerful
99 tool for reconstructing shortening rates and restore deformation stages (Burbank and Vergés,
100 1994; Zapata and Allmendinger, 1996; Meigs and Burbank, 1997; Echavarría et al., 2003;
101 Heermance et al., 2008; Li et al., 2010).

102 This paper is a review and analysis of the geology and structure of the Provençal and
103 Subalpine thrust wedges in Central Provence. To decipher the complex multistage thrusting
104 history in space and time of these superimposed thrust wedges, we built a 150 km-long
105 balanced and restored cross-section across the widespread dissected Provençal Beausset, Arc
106 and Rians foreland basins, and broad Alpine Valensole foreland basin (Fig. 1) using available

107 surface and subsurface data. Foreland syn-tectonic deposits and fault slip data allow us to
108 validate an accurate kinematics model of crustal shortening budget. We discuss the role of
109 pre-existing basement-cover heterogeneities, and we confirm their dominance on long-term
110 tectonic evolution of the orogenic belts.

111 **2. Structural framework**

112 From the Mediterranean coast to the south, to the Baronnies-Barles area to the north (Fig.
113 1), the ~E-trending folds and thrusts mainly result from the ~northward Provençal
114 compression (so-called Pyrenean-Provençal compression; Bertrand, 1888; Zürcher, 1891;
115 Kilian, 1892; Lutaud, 1924; Philip and Allemann, 1982; Tempier, 1987). Shortening occurred
116 from the Latest Santonian-Campanian to the Late Eocene (~50 m.y.) in response to the N-S
117 convergence of the Eurasian and African plates through the Sardinia-Corsica continental
118 block (Lacombe and Jolivet, 2005; Schreiber et al., 2011; Advokaat et al., 2014; van
119 Hinsbergen et al., 2020; Fig. 2b). During Oligocene-Miocene times, the inner Provençal thrust
120 wedge has been affected by the E-W opening of the West-European rift then the NW-SE
121 opening of the Ligurian-Provençal back-arc basin between Provence and Corsica-Sardinia
122 block (Hippolyte et al, 1993; Gattacceca et al., 2007; Bache et al., 2010; Oudet et al., 2010;
123 Nury et al., 2016; Romagny et al., 2020; van Hinsbergen et al., 2020; Fig. 2c). A SW-ward
124 migration of the Alpine deformation front into the southern Subalpine zone and Provençal
125 zone occurred from the Oligocene-Miocene to the present day (~30 m.y.) in response to the
126 eastward continental subduction of Eurasia beneath Adria (Fig. 2d). It is characterized by two
127 major arcs detached above Triassic evaporites, the Diois-Baronnies-Western Provence Arc to
128 the west and the Digne-Castellane Arc to the east (Goguel, 1937; Siddans, 1979; Villeger and
129 Andrieux, 1987). These arcs are separated by the poorly deformed Neogene Valensole
130 foreland basin (Figs. 1 and 2d). The thrust systems are accommodated by NNE-trending
131 strike-slip faults (Middle Durance fault zone, Bès fault; Hippolyte et al., 2012; Guyonnet-

132 Benaize et al., 2015) inherited from the Paleozoic and Mesozoic extensional fault pattern
133 (Roure and Colletta, 1996).

134 Many previous local studies have described the structural architecture of the thrust systems
135 in the Provençal zone (Kilian, 1892; Mennessier, 1959; Aubouin and Mennessier, 1962;
136 Aubouin and Chorowicz, 1967; Morabito, 1967; Guieu, 1968; Mennessier, 1970; Tempier,
137 1987; Bestani et al., 2015; Guyonnet-Benaize et al., 2015; Bestani et al., 2016; Espurt et al.,
138 2019b), in the Valensole foreland basin (Dubois and Curneille, 1978; Hippolyte and Dumont,
139 2000) and in the Subalpine zone of the Digne-Castellane arc (Goguel, 1937, 1939, 1944;
140 Unalan, 1970; Lemoine, 1972; Gigot et al., 1974; Debelmas, 1974; Giannerini, 1978; Siddans,
141 1979; Gidon and Pairis, 1992; Laurent et al., 2000; Graham et al., 2012; Célini et al., 2020).
142 North of the Paleozoic basement units outcropping in the Maures and Tanneron-Estérel
143 massifs (Figs. 1 and 3), the Provençal zone consists in strongly deformed Mesozoic sediments
144 and upper Cretaceous-Eocene foreland sedimentary units covered in places by Oligocene and
145 Miocene-Pliocene rocks. Southward, it shows ENE-trending thrust sheets (Bandol, Sainte-
146 Baume, Etoile-Pierresca, Aurélien Mount) associated with two large synclines (Beausset and
147 Arc; Bestani et al., 2015). Northward, the structures are characterized by E- to ESE-trending
148 thrusts (Sainte-Victoire Mountain, Pourrières, Pallières, Vinon, Les Murras and Gréoux) and
149 small synclines (e.g., Rians and Saint-Julien) interfering with NW- to NNE-trending zones of
150 middle-upper Triassic evaporitic-carbonate (e.g., Barjols or Carcès Triassic zones; Fig. 3).
151 These Triassic sequences show significant internal fold systems and can be associated with
152 halokinetic synclines (Angelier and Aubouin, 1976; Baudemont, 1985; Espurt et al., 2019b).
153 Northward, Miocene-Pliocene sediments fill the Valensole foreland basin. Its morphology
154 corresponds to a ~55 km long and up to ~30 km plateau dissected by Quaternary rivers (Gigot
155 et al., 1974; Dubois and Curneille, 1978; Mercier, 1979; Fig. 3). Westward, the basin is
156 bordered by the NNE-trending Middle Durance fault system (e.g., Roure et al, 1992). This

157 fault controlled extensional subsidence to the northwest of the Manosque basin during the
158 Mesozoic and the Oligocene, and transpressional uplift during its Miocene inversion (e.g.,
159 Roure and Colletta, 1996; Guyonnet-Benaize et al., 2015). Along the northern and
160 northeastern edges of the Valensole foreland basin, Mesozoic to Neogene strata are deformed
161 by SW-verging anticlines (e.g., Lambruissier, Mirabeau-Mallemoisson, Aiglun, La Maurière)
162 and overthrust by the SW-verging Digne Nappe and Castellane arc (Goguel, 1963; Gigot et
163 al., 1974; Dubois and Curnelle, 1978; Faucher et al., 1988, Hippolyte and Dumont, 2000; Fig.
164 3).

165 **3. Stratigraphy**

166 The basement rocks and sedimentary successions in the Provençal and Subalpine zones are
167 summarized in Figs. 4 and 5 and described hereafter.

168 **3.1. Variscan basement and upper Paleozoic sedimentary successions**

169 Paleozoic rocks are exposed in the Maures and Tanneron-Estérel massifs, Cap Sicié
170 peninsula, and punctually in the eastern Provençal zone (Carcès) and Subalpine zone of
171 Digne-Barles (St-Geniez, Verdaches and in tectonic slices at the base of the Digne Nappe)
172 (Baudemont, 1985; Tempier, 1987; Guiomar, 1990; Crévola and Pupin, 1994; Fig. 3). In the
173 Maures and Tanneron massifs, the Variscan belt is formed by E-verging thrust stacking of
174 Cambrian-Silurian foliated metasediments including phyllites, micaschists and gneisses. The
175 upper Carboniferous stage is associated with N-S orogen-parallel transpressional shearing
176 synchronous with the opening of intramountain basin filled by terrigenous sediments and
177 emplacement of granitoid and dome structures (Crévola and Pupin, 1994; Onézime et al.,
178 1999; Rolland et al., 2009; Corsini et al., 2010; Schneider et al., 2014; Simonetti et al., 2020).
179 During the continental break-up of Pangea, the Variscan framework was unconformably
180 overlain by a more than 800 m-thick Permian detrital and volcanoclastic sequence

181 accumulated in a puzzle of intracontinental basins associated with NNE- to SE-trending
182 steeply-dipping normal faults (Bathiard and Lambert, 1968; Baudemont, 1985; Delfaud et al.,
183 1989; Toutin-Morin et al., 1993; Durand, 1993; Cassinis et al., 2003). Well data indicate a
184 heterogenous distribution of the upper Carboniferous-Permian depocenters under the
185 Mesozoic-Cenozoic cover of Provençal zone and Valensole foreland basin (e.g., Morabito,
186 1967; Bathiard and Lambert, 1968; Dubois and Curneille, 1978; Baudemont, 1985).

187 **3.2. Triassic to upper Santonian Tethyan, Vocontian and South Provence rift successions**

188 Paleozoic rocks are unconformably overlain by ~30-120 m-thick lower-middle Triassic
189 white to red fluvial conglomerates and sandstones, and red-green argillites of the
190 Buntsandstein facies (Fig. 4). The opening of the Tethys Ocean started with the deposition of
191 a middle-upper Triassic succession of dolomite, marls, carnageules, halite, anhydrite, and
192 gypsum layers. The Muschelkalk facies is 100 to 600 m-thick. This unit is composed by the
193 lower Anhydritgruppe evaporites mainly developed in Provençal domain (Fig. 4), and upper
194 Muschelkalk limestones (Mennessier, 1959; Caron, 1979; Brocard and Philip, 1989; Toutin-
195 Morin et al., 1993; 1994; Espurt et al., 2019b). It is overlain by dolostones and evaporitic
196 marls of Keuper facies, 40 to 450 m-thick in Provence and more than 1 km-thick in the Digne
197 basin (Arnaud et al., 1976; Baudrimont and Dubois, 1977; de Graciansky et al., 1989; Fig. 4).
198 Rhaetian rocks consist of 30 to 80 m-thick alternation of limestones and yellow-green marls.
199 Middle-upper Triassic successions also contain ankaratrite intrusions and volcanoclastic
200 levels (Caron, 1970).

201 Sedimentologic and structural studies in Provence (Philip et al., 1987; Espurt et al., 2019b)
202 and southern Subalpine zone (Coadou et al., 1971; Rousset et al., 1983; Dardeau, 1988;
203 Mascle et al., 1988; Lemoine and de Graciansky, 1988; Haccard et al., 1989b; Graham et al.,
204 2012; Célini et al., 2020) suggest that the Jurassic-Santonian succession is associated with
205 significant thickness and facies changes due to syn-rift basement faulting, thermal subsidence

206 and local halokinetic motions of the ductile Triassic evaporite layers. From Jurassic to
207 Barremian, the Provençal domain consisted in dominantly marine carbonates that built huge
208 platforms (Léonide et al., 2007; Masse et al., 2009) including shallow marine limestones,
209 dolomites, calcareous breccias and lower Cretaceous rudist-bearing limestones with Urgonian
210 facies (Fig. 4). The thickness of this carbonate succession varies from 300 m to 1 km. These
211 platform sequences are superseded by pelagic facies in the Digne basin during Jurassic times.
212 Major subsidence of the Digne basin is recorded by the deposition of up to 1.4 km-thick of
213 lower Jurassic pelagic shales (Coadou et al., 1971; Fig. 4). During Middle Jurassic, the
214 subsidence reaches the Barles area with the deposition of more than 1.8 km-thick upper
215 Bathonian-lower Oxfordian black shales and turbidites (Black Shales Formation). The
216 overlying upper Jurassic-lower Cretaceous pelagic sediments attest for persisting subsidence
217 associated with gravitational instabilities. It consists of ~220 m-thick upper Oxfordian clayed
218 limestones and Kimmeridgian-Tithonian limestones covered by ~350 m-thick Berriasian-
219 Barremian clayed limestones, breccias and turbidites (Rousset et al., 1983; de Graciansky and
220 Lemoine, 1988; Célini et al., 2020).

221 From Latest Barremian to the beginning of Cenomanian and probably until Late Santonian,
222 the opening of the Bay of Biscay is associated with the Valais Ocean through the Vocontian
223 basin and led to NNE-SSW extension between Eurasia and Sardinia-Corsica block (Advokaat
224 et al., 2014; Tavani et al., 2018; van Hinsbergen et al., 2020; Angrand and Mouthereau, 2021;
225 Fig. 2a). During this period, the uplift of the Durance high separates the Digne-Barles basin to
226 the north from the Beausset basin to the south. The Beausset basin can be interpreted as the
227 eastern prolongation of the Pyrenean Rift System (Philip et al., 1987; Hennuy, 2003; Bestani
228 et al., 2016; Tavani et al., 2018; Floquet et al., 2021; Angrand and Mouthereau, 2021; Fig.
229 2a). ~350 m-thick of Aptian-Albian pelagic black shales and gravitational deposits
230 accumulated in the southern edge of the Beausset basin (Philip et al., 1987), overlain by ~1.1

231 km-thick Cenomanian-Santonian marine rudist-rich carbonate including siliciclastic
232 sediments (e.g., La Ciotat conglomerates) deposited into local troughs sourced from an
233 emerged basement paleo-high between the Provence and the Sardinia-Corsica block (Philip,
234 1970; Hennuy, 2003; Floquet et al., 2005, 2006). Northward, middle Cenomanian-upper
235 Santonian sediments progressively onlap the Durance high from the Beausset basin to the
236 southern edge of the Arc basin (Fig. 4). The Durance high corresponds to a large E-trending
237 dome with low relief associated with major subaerial erosion and bauxite formation from
238 uppermost Albian to lowermost Cenomanian at least, and till Santonian (Aubouin and
239 Chorowicz, 1967; Masse and Philip, 1976; Laville, 1981; Guyonnet-Benaize et al., 2010;
240 Figs. 2 and 4). North of the Durance high, Aptian-Albian deposits reach ~170 m-thick in the
241 Digne-Barles basin (Rousset et al., 1983; Friès, 1987; Haccard et al., 1989b; Breheret, 1995),
242 and consist of marine black shales. The overlying Cenomanian-Santonian strata dominantly
243 consist of more than 1.5 km-thick marine marls and limestones in the Digne basin (Rousset et
244 al., 1983).

245 **3.3. Uppermost Santonian to Eocene foreland successions in Provençal zone**

246 Plate reconstructions suggest the onset of N-S convergence between Africa and Eurasia
247 that led to the contraction of Corsica-Sardinia block and Provence (Advokaat et al., 2014; van
248 Hinsbergen et al., 2020; Angrand et Mouthereau, 2021; Fig. 2b), as illustrated by the
249 transition from marine to continental environment in Latest Santonian in the Beausset basin.
250 The growth of Provençal belt initiates at ~84 Ma and continues until the Late Eocene toward
251 the north (Philip, 1970; Philip et al., 1987; Tempier, 1987; Lemoine et al., 2000; Lacombe
252 and Jolivet, 2005; Montenat et al., 2005; Espurt et al., 2012; Philip et al., 2017). The
253 Provençal foreland succession is characterized by local continental facies dated by
254 magnetostratigraphy (Westphal and Durand, 1990; Cojan et al., 2003) and biostratigraphy
255 (Garcia and Vianey-Liaud, 2001; Philip et al., 2017). Upper Cretaceous-lower Eocene

256 foreland sequences are exposed in the Beausset, Arc, Rians, Saint-Julien and Esparron-de-
257 Verdon basins (Fig. 4). They are described in Fig. 5a. The Valdonnian and Fuvelian facies are
258 of Latest Santonian-Early Campanian age and correspond to lacustrine and palustrine black
259 limestones units including lignite beds (Fabre-Taxy and Philip, 1964; Durand, 1980; Fig. 5a).
260 The Begudian facies is of Middle Campanian age and consists in lacustrine and fluvial
261 deposits. The Rognacian facies is of Late Campanian-Maastrichtian age. It consists in
262 dinosaur eggshells and bones-bearing red marls, sandstones and Rousset lacustrine limestone
263 deposits (Cojan, 1989; 1993; Garcia and Vianney-Liaud, 2001). In the Arc basin, the
264 Cretaceous-Cenozoic boundary is located inside the Galante Formation, a 5-10 m-thick fluvial
265 conglomerates made of well-rounded pebbles of basement quartzite sourced from the Maures
266 massif (Durand, 1962; Cojan, 1993). The Danian corresponds to the Vitrollian facies with red
267 silty marls including lacustrine microcodium limestones. The Selandian-Thanetian Bird
268 Eggshells Marls Formation comprises the lacustrine limestones of Meyreuil and Saint-Marc
269 in the Arc basin. In the Rians basin it also includes the Touars conglomerates with clasts
270 derived from the Maures massif. The Ypresian period includes deposition of the Langesse
271 limestone and red-grey marls with gypsum in the Arc basin, and the fluvial Bluish Sands
272 Formation including lacustrine Bithynia bearing limestones in the Rians basin and High Var
273 area (de Lapparent, 1938a, 1938b; Corroy, 1957; Durand and Tempier, 1962; Angelier, 1971;
274 Philip et al., 2017; Fig. 1). Finally, the Lutetian consists in Montaignet and Cuques lacustrine
275 limestones cropping out in the Arc basin. The upper Cretaceous-Eocene continental sequences
276 of the Provençal foreland basin include locally massive syn-orogenic alluvial-fan deposits
277 with growth strata recording thrust activity (Durand and Tempier, 1962; Guieu, 1968;
278 Angelier, 1971; Leleu et al., 2009; Espurt et al., 2012). They correspond to upper Cretaceous
279 Begudian-Rognacian breccias and Paleocene microcodium breccias with clasts derived
280 generally from the nearby Jurassic-Cretaceous growing reliefs (Fig. 5a).

281 **3.4. Oligocene syn-rift and Miocene post-rift successions in Provençal zone**

282 The Oligocene is a transitional period between the end of the Provençal compression and
283 the initial growth of the southern Subalpine zone (Nury, 1988). In the south and east of the
284 study area, the Provençal compressional structures were affected by the E-W opening of the
285 West-European rift then the NW-SE opening of the Ligurian-Provençal back-arc rift between
286 Provence and Corsica-Sardinia block (Hippolyte et al, 1993; Gattacceca et al., 2007;
287 Advokaat et al., 2014; van Hinsbergen et al., 2014; Bestani et al., 2016; Nury et al., 2016;
288 Romagny et al., 2020, van Hinsbergen et al., 2020; Fig. 2c). The result of these successive
289 extensional events is the formation of large intracontinental basins like the Marseille,
290 Aubagne, Aix-en-Provence, and Manosque basins (Fig. 3; Hippolyte et al, 1993). Smaller
291 Oligocene basins developed in southeastern Provence. They are the NE- to E-trending
292 Bandol, Signes-Méounes, Peypin, Saint-Zacharie and Nans basins/troughs (Popoff, 1973;
293 Nury, 1988) (Figs. 3, 4 and 5a). In the High Var area (Fig. 1), Oligocene sediments fill N- to
294 NNW-trending narrow basins (Touraine, 1967; Angelier, 1971; Angelier and Aubouin, 1976;
295 Giannerini, 1978; Philip et al., 2017) like the La Combe, Bourdas and Plan d'Auron troughs
296 (Figs. 3, 4 and 5a). In the overall, syn-rift sediments correspond to lacustrine and clastic
297 deposits including marls and locally gypsum intercalations as in the Manosque basin (Fig. 3),
298 but the subsidence rate was ten times larger during the Ligurian-Provençal rifting than during
299 the West-European rifting (Hippolyte et al., 1993).

300 The drifting of the Ligurian-Provençal basin associated with the counterclockwise rotation
301 of the Corsica-Sardinia block during the Miocene times led to the flooding of the inner
302 Provençal basement thrust wedge (Guieu and Roussel, 1990; Hippolyte et al, 1993;
303 Gattacceca et al., 2007; Oudet et al., 2010; Bestani et al., 2016; Nury et al., 2016). In the
304 onshore domain, Miocene (Burdigalian(?) to Tortonian) continental strata with marine
305 intercalations filled N- to NW-trending incised valleys and basins (e.g., Les Maurras, Barjols,

306 Nans, Signes/Méounes and Bandol). These deposits consist of basal red breccias, sandstones
307 and marls, and upper fluvial conglomerates (Mennessier, 1959; Cornet, 1965; Angelier and
308 Aubouin, 1976) generally originated from the erosion of nearby Jurassic reliefs (Les Maurras
309 basin) or from the Maures massif (Nans basin) (Figs. 3 and 4; Cornet, 1961). In the west of
310 the study area (Luberon-Nerthe; Fig. 1), shallow marine environment is associated with the
311 formation of concomitant wave-cut platforms overlain by marine molasse including
312 limestones and grainstones (Champion et al., 2000; Besson, 2005; Oudet et al., 2010).
313 Basaltic clasts related to Messinian intrusions and lava flows of Evenos are found in the upper
314 conglomerate sequences (Pliocene?) of the Bandol basin (Coulon, 1967; Baubron, 1984).
315 Pliocene to Quaternary sediments are mainly found within paleo-valleys that were incised
316 during the Messinian and then filled by Gilbert type deltas after the reflooding of the
317 Mediterranean Sea at 5.46 Ma (Clauzon, 1978; 1982, Bache et al., 2012).

318 **3.5. Cenozoic foreland successions in Subalpine zone: Barles zone and Valensole foreland** 319 **basin**

320 In the Barles zone, folded Mesozoic sequences are unconformably overlain by the poorly
321 dated (upper Eocene-lower Oligocene?) continental Basal Breccia Formation (Lemoine, 1972;
322 Haccard et al., 1989a,b; Fig. 5b). It is composed of breccia channels interbedded with red
323 shales. The breccia channels are composed of clasts and olistoliths, exclusively sourced from
324 local Jurassic and Cretaceous limestone strata, included in a red shaly matrix with
325 microcodium (Gigot et al., 1974; Haccard et al., 1989b). The Basal Breccia Formation is
326 interpreted as sedimentary drape sequences (with differential compaction) sealing the paleo-
327 morphology of E-trending Eocene Provençal folds (Lemoine, 1972; Maillart et al., 1987;
328 Haccard et al., 1989b). However, the breccia beds also exhibit large sedimentary fans which
329 could be interpreted as growth strata deformed by progressive tectonic uplift (Maillart et al.,
330 1987). The Basal Breccia Formation might be laterally equivalent to upper Eocene

331 nummulitic limestones deposited immediately to the east in the Digne-Castellane basin
332 (Haccard et al., 1989b; Fig. 4) or post-date it and of Early Oligocene age (Lichorish and Ford,
333 1998).

334 The growth of the southwestern Subalpine zone started during the Oligocene and continued
335 throughout the Neogene. Oligocene rocks belonging to the Barles zone and northern edge of
336 the Valensole basin consist of basal fluvial sandstone and red clay named as the Red
337 Continental Molasse Formation and the upper brackish deposits named as the Grey Molasse
338 Formation (Haccard et al., 1989b; Fig. 5b). This succession reaching more than 450 m-thick is
339 interpreted as deposited as the distal foreland sediments of the inner Alps (Maillart et al.,
340 1987; Gidon, 1997; Ford et al., 1999). Like the Basal Breccia Formation, the Red Continental
341 Molasse Formation includes large sedimentary fans in the Barles zone and northern edge of
342 the Valensole basin which could be interpreted as growth strata deformed by progressive limb
343 rotation (Gidon, 1997). The Valensole foreland basin is then filled by more than 3 km-thick
344 marine to continental Miocene-Pliocene strata with strong facies and thickness variations
345 (Gigot et al., 1974; Gidon and Pairis, 1988; Haccard et al., 1989a,b; Clauzon et al., 1989;
346 Cojan et al., 2013). It is connected westward to the Forcalquier and Cucuron-Pertuis basins
347 (Fig. 3). Field studies and well data in the center part of the basin give a global insight of the
348 Miocene-Pliocene sedimentary succession and architecture (Dubois and Curnelle 1978;
349 Clauzon et al., 1989; Couëffe et al., 2004; Besson, 2005; Figs. 4 and 5b). The transgressive
350 Aquitanian-Burdigalian Lower Marine Molasse Formation consists in tidal facies deposited in
351 major subsiding environment that pass northward into siliciclastic deposits sourced from the
352 inner Alps domain (Couëffe et al., 2004). The Langhian to Tortonian-lower Messinian
353 succession is composed of marls, coarse-grained sandstones and limestones alternations of the
354 Upper Marine Molasse Formation that pass laterally to prograding fluviodeltaic conglomerate
355 sequences of the Valensole 1 Formation (Beaudouin et al., 1966, 1970; Dubois and Curnelle

356 1978; Mercier, 1979; Aguilar and Clauzon, 1981; Haccard et al., 1989b; Clauzon et al., 1989;
357 Besson, 2005; Cojan et al., 2013; Fig. 5b). The Valensole 1 Formation comprises Durance
358 conglomerates rich in siliciclastic clasts westerly sourced by the paleo-Durance river and
359 Subalpine conglomerates rich in local Mesozoic limestone clasts easterly sourced (Haccard et
360 al., 1989b). The Valensole 2 Formation mainly consists of marine clay and continental
361 deposits (conglomerates, sandstones and claystones) of Pliocene to Quaternary age (Clauzon,
362 1979; Mercier, 1979; Clauzon et al., 1989). Locally, near Oraison (Fig. 3), a Messinian
363 canyon was mapped by Dubar (1983, 1984). It allows the distinction between the Valensole 1
364 and Valensole 2 Formations, separated by a major discontinuity, the Messinian Erosional
365 Surface (Mercier, 1978, 1979; Clauzon, 1979; Dubar, 1983, 1984; Clauzon, 1996). The
366 Tanaron Formation is the youngest stratigraphic level in the footwall of the Digne Nappe
367 (under the La Robine half-klippe) in the northern edge of the Valensole foreland basin (Gigot
368 et al., 1974; Figs. 3 and 5b). The age of the Tanaron Formation is still a matter for debate. It is
369 composed of mixed massive olistoliths (Triassic gypsum, Callovian-Oxfordian black shales,
370 upper Jurassic-lower Cretaceous limestones, upper Eocene-Oligocene continental molasses
371 and Miocene marine molasse) mainly sourced from the collapse of the Barles fold zone
372 (Gigot et al., 1974). Rare olistoliths of upper Cretaceous carbonates and nummulitic
373 limestones are interpreted to come from the erosion of the Digne Nappe (Haccard et al.,
374 1989b). The olistoliths are included in grey-red marls and sandstones. Gigot et al. (1974),
375 Crumeyrole et al. (1991) and Hippolyte et al. (2011) pointed out that the Tanaron Formation
376 infills a very large scale erosional feature. At a time where the Valensole conglomerates were
377 considered as Pliocene, Gigot et al. (1974) assigned a Late Pliocene age to the overlying
378 Tanaron Formation. The geological map (Haccard et al., 1989a,b) ascribes a Latest Pliocene
379 age for this formation. However, Gidon and Pairis (1988, 1992) emphasized that the Tanaron
380 Formation is not dated biostratigraphically. The mapping of Messinian-Zanclean paleo-

381 valleys around Digne revealed that a Zanclean erosional surface has truncated the Tanaron
382 Formation at La Bonnette and constrains the age of this formation as older than the Late
383 Pliocene (Hippolyte et al., 2011). Furthermore, these authors suggested that if the erosional
384 surface at the base of the Tanaron Formation was the Messinian Erosional Surface related to
385 the desiccation of the Mediterranean basin at 5.6 Ma (Clauzon et al., 1996; Bache et al., 2012;
386 Roveri et al., 2014) the age of the Tanaron Formation would be Late Miocene to Early
387 Pliocene (Hippolyte et al. 2011).

388 **3.6. Décollement levels**

389 Inherited Variscan metamorphic foliation and upper Carboniferous-Permian extensional
390 faults controlled décollement in the basement as observed in the Cap Sicié and Maures
391 massifs and also suggested by subsurface data under the Provence cover (Tempier, 1987;
392 Roure and Colletta, 1996; Cushing et al., 2008; Espurt et al., 2019b). The Paleozoic
393 framework was intermittently reactivated during the Mesozoic rifting and during the
394 Provençal and Alpine shortening episodes. The basal Muschelkalk and Keuper units are
395 interpreted as main regional decollement levels of the sedimentary cover above the Paleozoic-
396 Triassic (Buntsandstein) basement (Bathiard and Lambert, 1968; Caron, 1979; Gidon, 1982;
397 Gidon and Pairis, 1992; Fig. 4). This middle-upper Triassic evaporitic-carbonate package
398 shows major thickness variations related to local depocenters during rifting or tectonic
399 duplication and disharmonic deformations during the compressive stages (Goguel, 1939;
400 Bathiard and Lambert, 1968; Gigot et al., 1974; Toutin-Morin et al., 1993, 1994; Baudemont,
401 1985; Espurt et al., 2019b). Middle Jurassic black shales, upper Jurassic-Berriasian limestones
402 and Oligocene-Miocene sequences (Fig. 4) promote disharmonic deformations with internal
403 folding and shearing particularly in the Barles and Vélodrome structures (Gidon and Pairis,
404 1992).

405 **4. Cross-sectional structural architecture**

406 To illustrate the structural architecture of the Provençal and Subalpine thrust wedges, we
407 constructed a 150 km-long crustal-scale cross-section from the Mediterranean coastal zone to
408 the south to the Digne-Barles region to the north (Figs. 3 and 6). The cross-section trends
409 N175°E in the coastal basement units and the Beausset basin, then N002°E across the Sainte-
410 Baume Nappe and Aurélien Mount thrust, and finally N024°E across the northern Provençal
411 thrusts, Valensole basin up to the Digne Nappe. This section was constructed according to
412 thrust tectonics concepts (e.g., Dahlstrom, 1969; Hossack, 1979; Elliot, 1983; Suppe, 1985).
413 The section is localized in a zone only weakly affected by the Oligocene rifting (Bestani et al.,
414 2016). We used existing 1:50,000 geological maps of French Geological Survey (BRGM) and
415 new surface structural data from several fieldworks. We also used existing cross-sectional
416 data of Kilian (1892), Goguel (1939, 1944), Morabito (1967), Gigot et al. (1974), Debelmas,
417 1974; Dubois and Curnelle (1978), Combes (1984), Haccard et al., 1989b; Gidon and Pairis
418 (1992), Hippolyte and Dumont (2000), Guyonnet-Benaize et al. (2015), Bestani et al. (2015,
419 2016) and Célini et al. (2020). Thrust geometries at depth and basement geometry have been
420 constrained using exploration wells and geophysical data. We used nine major exploration
421 wells (from south to north: Gréoux (G3, G2, G1, G4, S1, S4), Les Mées (LM1), and Mirabeau
422 (M1, M2)) some of them (G1, S1, LM1, M2) reaching the Paleozoic substratum (Fig. 3). We
423 interpreted two seismic reflection profiles, VL85-P and 75DV-2, that cover the central-
424 southern part of the Valensole basin over a distance of ~28 km. These profiles have been
425 reprocessed by Agence de l'Eau Rhône Méditerranée Corse and available for this study (Fig.
426 3). A time-depth conversion of seismic profiles was performed using the MOVE software and
427 seismic velocities of sedimentary intervals provided by BRGM. Finally, fault location at
428 depth is interpreted by some present day deep-located earthquakes (Cushing et al., 2008).

429 In the southern part of the Provençal zone, the cross-section trace located between the Cap
430 Sicié and the Aurélien Mount thrust follows the section of Bestani et al. (2015) (Figs. 3 and

431 6). This section has been updated from previous version in terms of displacement magnitude
432 of the thin-skinned thrust sheets and basement deformations based on new structural
433 interpretations and new geological mapping (Laville et al., 2018). Because the section trace is
434 located lateral to the Cap Sicié peninsula (~10 km to the east), its thrust geometry has been
435 projected onto the section. To the north of the Aurélien Mount thrust, the cross-section has
436 been fully constructed for this study and is presented for the first time (Fig. 3). The proposed
437 balanced cross-section is one possible construction, but it is the most realistic solution
438 consistent with the available surface and subsurface structural data presented in this work.

439 The construction, balancing and restoration of the cross-section were performed with the
440 MOVE structural modeling software following thrust tectonic concepts and flexural-slip
441 algorithm assuming constant bed length and thickness (Dahlstrom, 1969; Boyer and Elliott,
442 1982; Suppe, 1983; Suppe and Medwedeff, 1990). For the middle Jurassic black shales, and
443 Oligocene-Miocene sequences, in zone where these ductile units promote strong internal
444 deformations (Barles folds and Vélodrome), we assumed an area balance approach (e.g.,
445 Mitra and Namson, 1989; Mitra, 2002; Butler, 2013; Butler et al., 2020). For the ductile
446 middle-upper Triassic evaporitic-carbonate units which may move in three dimensions (and
447 their volumes can change as a result of dissolution, fluid migration and surface erosion during
448 successive stages deformation and diapirism), we also achieved as much as possible an area
449 balance approach. For deep crustal level, we also assumed an area balancing approach. The
450 orientation of the cross-section is orthogonal to the trend of the structures of both orogenic
451 systems (see section 6). No major strike-slip fault system is crossed except in the Huveaune
452 zone (Fig. 3; Guillemot et al., 1973; Philip, 2018).

453 **4.1. The Provençal zone**

454 **4.1.1. Surface structure**

455 The surface structure of the Provençal zone is described from south to north, i.e., from the
456 inner coastal basement units (preserved from the Ligurian-Provençal rifting) and thrust sheet
457 domain, to outer thrust systems of High Var (Fig. 6a,b).

458 The Cap Sicié peninsula is formed by two imbricate thrusts. The upper N-verging Cap
459 Sicié thrust is composed of Variscan phyllites and quartzites covered by N-dipping lower-
460 middle Triassic strata to the north (Fig. 7a-c). The Cap Sicié thrust is folded by the lower
461 Saint-Mandrier thrust leading to the formation of the Saint-Mandrier half-window (Zürcher,
462 1893; Haug, 1925; Tempier, 1987), made of lower-middle Triassic and Permian strata at the
463 surface (Figs. 6b and 7a,c,d) and Variscan phyllites and quartzites at depth.

464 North of the Cap Sicié basement imbricate, the sedimentary cover of the Provençal zone is
465 detached northward above ductile middle-upper Triassic layers and associated Triassic highs
466 (Gouvernet, 1963; Bathiard and Lambert, 1968; Tempier, 1987; Bestani et al., 2016; Espurt et
467 al., 2019b). The Bandol cover thrust including the Bandol syncline is transported at the front
468 of the Cap Sicié thrust. The Bandol thrust is emplaced above the Saint-Cyr Triassic high
469 where lower Santonian carbonates unconformably overlain middle Triassic limestones
470 (Philip, 1967; Philip et al., 1985, 1987; Espurt et al., 2019b; Fig. 8). Klippen of Triassic rocks
471 of the Bandol thrust rest on upper Santonian and lower Campanian (Valdonnian-Fuvelian
472 facies) strata of the Beausset syncline (Bertrand, 1887; Gouvernet, 1963) (Fig. 3). The Sainte-
473 Baume Nappe transported northward the Beausset syncline (Fig. 9; Guieu, 1968; Bestani et
474 al., 2015; Philip et al., 2018; Philip, 2019). Its hanging wall ramp consists in upper Triassic to
475 lower Cretaceous strata dipping 10°-15° southward, deformed by out-of-sequence thrusts
476 (Fig. 9c; Bestani et al., 2015). Its footwall is constituted by the tight Plan d'Aups syncline
477 cored by lower Campanian strata (Fuvelian facies), Pic des Corbeaux anticline, and broad La
478 Lare anticline mantled by upper Jurassic limestones. Small klippen of Jurassic and Barremian
479 rocks in the Huveaune zone (Saint-Zacharie and Nans basins) indicate the northward advance

480 of the Sainte-Baume Nappe (Figs. 3 and 6b; Guieu, 1968). The substratum of the Huveaune
481 zone is constituted by middle-upper Triassic rocks and relatively thin Jurassic sequences
482 which are unconformably covered by upper Cretaceous (Valdonnian-Fuvelian and Begudian-
483 Rognacian facies) and Oligocene strata (Fig. 3; Philip, 2018, 2019). The Aurélien Mount
484 thrust transported northward Jurassic limestones above the Arc syncline (Figs. 6b and 10;
485 Aubouin and Chorowicz, 1967; Guieu, 1968; Popoff, 1973). The hanging wall ramp of the
486 Aurélien Mount thrust shows Jurassic strata dipping 12° southward. The Sainte-Baume Nappe
487 and Aurélien Mount thrust, and southern limb of the Arc syncline are cut by NE-trending
488 normal faults (Bestani et al., 2016). This normal faulting is contemporaneous with opening of
489 the Aubagne, Peypin, Saint-Zacharie and Nans extensional basin framework during the
490 Oligocene (Figs. 3 and 6b; Nury, 1988; Hippolyte et al., 1993).

491 The surface and well data show that the Arc basin is an asymmetrical syncline
492 characterized by a N-thinning sedimentary wedge in first lower Santonian rudist carbonates,
493 then upper Santonian-lower Campanian (Valdonnian-Fuvelian) to middle Campanian-
494 Maastrichtian (Begudian-Rognacian) continental syn-orogenic facies (from ~ 1.1 km-thick to
495 the south, to less than 250 m-thick to the north; Fig. 6b). The core of the syncline is
496 constituted by ~ 200 m-thick flat Danian to Ypresian strata that form the Cengle plateau (Figs.
497 3 and 11a). Along the section, the Arc syncline is transported northward above the Pourrières
498 thrust constituted by 12° S-dipping Jurassic-lower Cretaceous strata (Fig. 11a). The Pallières
499 thrust system transports northward the Ollières syncline cored by upper Santonian-Campanian
500 (Valdonnian-Fuvelian-Begudian) continental facies strata. (Fig. 6b). The Pallières thrust
501 system is composed by two units. The upper unit corresponds to a $\sim 5^\circ$ S-dipping homocline
502 made of upper Jurassic-lower Cretaceous limestones. The lower unit involves upper Jurassic
503 limestones unconformably covered by Danian *Microcodium* breccias (Fig. 11c). The Pallières
504 thrust system overthrusts northward the E-trending Rians syncline filled by more than 700 m-

505 thick Danian to Ypresian syn-orogenic package including massive alluvial fan deposits (Fig.
506 11c; Tempier, 1963; Mennessier, 1970; Angelier, 1971; Angelier and Aubouin, 1976; Philip
507 et al., 2017). The Pourrières and Pallières thrusts form the eastern continuation of the bi-
508 verging Sainte-Victoire Mountain thrust system, while to the east they branch into the Barjols
509 Triassic zone (Figs. 3 and 11a,b). The N-verging Vinon thrust includes the Mont-Major
510 anticline locally thrusting southward the northern flank on the Rians syncline and Triassic
511 evaporite-cored anticline of Valavés (Figs. 3 and 6b; Mennessier, 1970). Northward, the
512 Vinon thrust dips 12° to the south and emplaces upper Jurassic limestones on Paleocene strata
513 of the Saint-Julien syncline (Fig. 12a). The geometry of the N-verging Les Maurras thrust is
514 revealed by the well G2 and by intensively sheared Oxfordian marly limestones as observed
515 in the field (Fig. 12b). The Les Maurras thrust is unconformably covered by poorly deformed
516 red sandstones and thick breccias layers, and Pliocene conglomerates of the Valensole
517 Formation. Eastward, the footwall of the Les Maurras thrust is constituted by the La Mourotte
518 syncline filled by Paleocene and Ypresian strata (Fig. 3; Mennessier, 1970; Angelier and
519 Aubouin, 1976). Finally, the N-verging Gréoux thrust emplaced lower Jurassic rocks over
520 lower Cretaceous limestones and Paleocene strata of the Esparron-de-Verdon syncline (Fig.
521 12c). This thrust dips $\sim 10^\circ$ southward at surface and increases at depth at $\sim 33^\circ$ southward as
522 revealed by the exploration wells G1 and G2 (Morabito, 1967). The thrust systems of northern
523 Provençal zone are later cut by NW-trending normal faults (e.g., Ginasservis fault) which are
524 associated with the development of La Combe, Bourdas, Plan d'Auron basins filled by
525 Oligocene sediments (Fig. 3; Angelier and Aubouin, 1976; Philip et al., 2017).

526 The southern Provençal zone shows locally inherited Mesozoic halokinetic structures
527 associated with normal faults previously described by Bestani et al. (2015) and Espurt et al.
528 (2019b) as the Saint-Cyr and Huveaune Triassic highs. The northern Provençal zone connects
529 eastward with the large Triassic depression of Barjols including halokinetic structures (Fig.

530 3). Along the cross-section, upper Jurassic-lower Cretaceous strata of the Pourrières thrust
531 depict growth wedge geometry and faulting, suggesting that this thrust could be superimposed
532 on an early SW-dipping normal fault zone with halokinetic movement of Triassic evaporites
533 at depth (Figs. 3 and 6b). Similarly, well data show strong thickness variations in upper
534 Jurassic carbonates into the Les Maurras thrust (Fig. 12d), suggesting early halokinesis.

535 **4.1.2. Deep structure**

536 Cross-section construction suggests that the Provençal zone is characterized by a
537 heterogenous geometry of the basement-cover interface resulting from basement inheritance
538 and basement-involved shortening. The basement of Provençal zone shows two styles of
539 deformation along the cross-section (Fig. 6b). Southward, low-angle basement thrusts (Cap
540 Sicié and Saint-Mandrier thrusts, and deeper thrust under the Sainte-Baume Nappe) propagate
541 through the Variscan metamorphic foliation or inherited shear zones (Espurt et al., 2019b).
542 The extensional geometry of the upper Carboniferous-Permian basins at depth is well
543 preserved as imaged by E- and N-trending seismic profiles, for instance, in the Luc basin at
544 the northwestern flank of the Maures massif (Fig. 13; Baudemont, 1985). The basins mainly
545 correspond to half-grabens filled by more than 1 km-thick upper Carboniferous-Permian
546 strata and associated with WNW- to NNW-dipping normal faults. These faults inherited from
547 the rifting of Pangea can have been locally reactivated during the Mesozoic rifting events
548 (Toutin-Morin et al., 1993, 1994; Espurt et al., 2019b). During the Provençal shortening, the
549 extensional features were cut and transported passively above N-verging basement thrusts
550 (Bathiard and Lambert, 1968; Baudemont, 1985; Tempier, 1987; Espurt et al., 2019b). The
551 shortening of the basement thrusts fed the slip of cover thrusts of the southern Provençal zone
552 (Bandol, Sainte-Baume and Aurélien Mount). In the northern Provençal domain, field and
553 geophysical data, and cross-section constructions (Biberon, 1988; Roure and Colletta, 1996;
554 Espurt et al., 2012, 2019) allow to infer the geometry and depth of the basement cover

555 interface. These data also suggest a heterogenous distribution of the upper Carboniferous-
556 Permian to Triassic depocenters. The base of the Arc syncline is inferred at ~2.2 km bsl and
557 the base of the Rians syncline at ~1.7 km below sea level (bsl). At the boundary between the
558 Provençal zone and Valensole basin (footwall of the Gréoux thrust), the basement depth is
559 located at ~1.4 km bsl. Surface data and cross-section construction suggest that the Paleozoic
560 basement is located at a shallow depth (~800 m bsl) under the Valavès Triassic anticline.
561 Contrary to the south, the basement structure of northern part of Provençal zone is interpreted
562 as half-grabens delimited by major NE-dipping faults (Fig. 6b). These basement thrusts are
563 interpreted as short-cut thrusts propagating through inherited upper Carboniferous-Permian to
564 Triassic half-grabens (Espurt et al., 2012; 2019b). In this zone, shallow-depth earthquakes are
565 observed (e.g., Valavés anticline; Figs. 3 and 6a). They could be tectonically connected with
566 deeper events occurring along basement faults as suggested by cross-section construction
567 (Cushing et al., 2008).

568 **4.2. The Subalpine zone: Digne Nappe, Barles fold system and Valensole foreland basin**

569 **4.2.1. Surface structure**

570 The Digne Nappe, Barles fold system and northern edge of the Valensole basin are well
571 exposed in a N-S section along the Bès valley (Figs. 3 and 14). The stratigraphy and the main
572 tectonic structures along this valley have been described for a long time by geologists (e.g.,
573 Goguel, 1939, 1944; Debelmas et al., 1970, 1974; Lemoine, 1972; Gigot et al., 1974; Gidon,
574 1982, 1989; Debelmas et al., 1983; Gidon and Pairis, 1986, 1988, 1992; Haccard et al.,
575 1989b). This area around the Barles village became famous for fieldworks with students
576 coming from national and international universities. More recent studies discussed the timing
577 of deformation (Fournier et al., 2008; Hippolyte et al., 2011) and tried to decipher Jurassic
578 salt tectonics through an overprint of Cenozoic contractional tectonics (Graham et al., 2012;

579 Célini et al., 2020). The surface structure of the Subalpine zone is described from north to
580 south, i.e., from the inner Digne-Barles units to Valensole foreland basin.

581 The Digne Nappe involves a thick (up to 4.5 km) Jurassic-Cretaceous pelagic succession
582 initially deposited in the Digne basin, and covered by Eocene-Oligocene strata northeastward
583 (Fig. 4; Goguel, 1944; Rousset et al., 1983). The paleo-temperature (~100 °C) recorded by
584 Oligocene-Miocene rocks in the footwall of the Digne Nappe indicates that the thickness of
585 the Digne Nappe was about 2.9 km above the Barles tectonic half-window (Schwartz et al.,
586 2017a,b; Figs 6c and 14). The Digne Nappe involves locally basement tectonic slices (Fig.
587 15) and internal folding (Guiomar, 1990). It is regionally detached SW-ward above a thick
588 Triassic evaporitic-carbonate layer and overthrusts the Barles fold system and the
589 northeastern edge of the Valensole foreland basin (Figs. 6c and 14). North of Digne, its
590 frontal part is constituted by the syncline-shaped La Robine half-klippe which overlies the
591 upper Miocene-lower Pliocene sequences of the Thoard syncline. Northward, the hanging-
592 wall of the Digne Nappe displays a large antiform, leading to the formation of the erosional
593 Barles half-window (Figs. 6c and 14). Along the cross-section, the Digne Nappe roots north
594 of the Verdaches anticline in thin layer of middle-upper Triassic strata. The E-trending
595 asymmetric Verdaches anticline is underlined by middle Triassic limestones and basal
596 Triassic sandstones which unconformably overlain upper Carboniferous coal and sandstone
597 beds (Figs. 6c and 15; Haccard et al., 1989b; Guiomar, 1990). The northern backlimb dips 10°
598 northward while the southern forelimb dips 25° southward. The structure of the Barles
599 antiform corresponds to very tight anticlines (La Petite Cloche, La Grande Cloche and La
600 Maurière anticlines along the cross-section) mantled by upper Jurassic-lower Cretaceous
601 limestones and cored by Callovian-Oxfordian black shales (Fig. 14). The anticlines are
602 bounded by synclines cored by uppermost Eocene-Oligocene strata that unconformably
603 overlain Mesozoic sequences (Fig. 16; Lemoine, 1972; Gidon, 1982; Haccard et al., 1989a).

604 Cross-section construction suggests that the Barles fold system is detached southward above
605 middle-upper Triassic strata (Fig. 6c). The La Petite Cloche thrust transports southward the
606 Les Sauvans syncline filled by thick Callovian-Oxfordian black shales (Fig. 6c). Its vertical to
607 overturned northern limb is the Barre de Chine armed by ~200 m-thick lower Jurassic
608 limestones overlain by a huge volume of Triassic shales, cagneules and gypsum of the Barles
609 diapir (Arnaud et al., 1976; Haccard et al., 1989a; Gidon and Pairis, 1992; Graham et al.,
610 2012). Although the Barles fold zone and Digne Nappe recorded strong contractional
611 deformation, field data show structural evidence of pre-existing halokinetic structures and
612 normal faulting (Gidon, 1982). For instance, the minibasin structures in La Robine half-klippe
613 and of the Les Sauvans syncline-Barre de Chine flap are interpreted as resulting from
614 halokinetic motion of Triassic evaporites during the Jurassic-Early Cretaceous times (Graham
615 et al., 2012; Célini et al., 2020).

616 South of the Barles anticlines, the three-dimensional Vélodrome syncline develops,
617 involving mainly Miocene rocks (Marine Molasse and Valensole 1 Formations)
618 unconformably covered by the olistolithic Tanaron Formation (Fig. 14). This complex
619 structure exhibits important contractional growth strata along syncline limbs with fold-
620 accommodation faults including out-of-syncline thrusts and disharmonic folding (Goguel,
621 1939; Gigot et al., 1974; Pairis and Gidon, 1987; Gidon, 1989; Gidon and Pairis, 1992).
622 Recently, it has been speculated that the three-dimensional structure of the Vélodrome may
623 correspond to a Miocene minibasin that sank into a hypothetical Oligocene salt layer (Célini
624 et al., 2020; Célini, 2020). The Vélodrome structure corresponds to the northern limb of the
625 Thoard syncline. The thickness of the Miocene-Pliocene syn-tectonic deposits increases
626 southward, from 1 km in the Vélodrome to more than 3.3 km in the core of the Thoard
627 syncline (Figs. 6c and 17). This syncline is transported southward onto the Aiglun and
628 Mirabeau-Mallemoisson S-verging fault-propagation folds armed by Miocene conglomerate

629 beds and probably detached in Triassic strata. The northern Aiglun anticline displays a
630 backlimb dipping 35-40° northward and a forelimb dipping 70° southward (Fig. 18a). The
631 southern Mirabeau-Mallemoisson anticline has a 40° N-dipping backlimb and a vertical
632 forelimb (Fig. 18b). Well data indicate that this later is associated with a thrust dipping 38°
633 northward. Well data also reveal that a sub-thrust develops in the footwall block. South of the
634 Bléone river (Fig. 3), Miocene strata are deformed by the Quaternary Lambruissier fold-
635 propagation fault (Hippolyte and Dumont, 2000) probably detached in the upper Triassic
636 strata. It is characterized by a 10-15° N-dipping backlimb and a 80° S-dipping forelimb. The
637 displacement on Lambruissier thrust decreases rapidly westward, explaining why it is not
638 visible in map view as well as in subsurface data near Les Mées borehole (Figs. 3 and 19).
639 Southward, the Valensole foreland basin is filled by upper Miocene conglomerates of the
640 Valensole 1 Formation topped by the Valensole 2 Pliocene Formation whose surface forms
641 the slightly S-dipping Puimichel plateau (Fig. 18c). In the southern edge of the Valensole
642 basin, upper Miocene-Pliocene rocks lie unconformably on lower Cretaceous strata.

643 **4.2.2. Deep structure**

644 In the southern part of the Valensole basin, reprocessed seismic reflection profiles and well
645 data have been used to interpret the geometry of hidden thrust systems under the Miocene-
646 Pliocene sedimentary infill (Fig. 19). The seismic profile 75DV-2 confirms that the Miocene
647 strata are unconformably deposited on top of a deep basement high, the Les Mées high,
648 underlined by folded Mesozoic strata with a thin (less than 30 m-thick) Triassic succession
649 (Dubois et Curnelle, 1978). The uplift of the Les Mées high is controlled by a N-dipping
650 basement ramp proved by subsurface data (Dubois and Curnelle, 1978). The basement ramp
651 branches upward into the Triassic strata then splits into two S-verging branches deforming
652 Miocene-Pliocene deposits under the Asse river as observed on the seismic profile VL85-P.
653 These hidden compressional structures are interpreted as Provençal folds and thrusts (Dubois

654 and Curnelle, 1978), slightly reactivated during the Alpine compression (Hippolyte and
655 Dumont, 2000; Godard et al., 2020).

656 Like the Provençal zone, cross-section construction and subsurface data illustrates that the
657 Subalpine zone (including the Valensole foreland basin) is characterized by a heterogenous
658 geometry of the basement-cover interface resulting mainly from basement-involved
659 shortening. Northward, the Carboniferous and lower Triassic teguments of the Verdaches
660 anticline were uplifted, as a result of reverse motions on a reactivated upper Carboniferous
661 normal fault (Fig. 15). However, the structural culmination of the Verdaches anticline is
662 associated with a deeper basement thrust wedge propagating under the Valensole foreland
663 basin in the Triassic evaporites as initially proposed by Goguel (1944) and Gigot et al. (1974).
664 Southward, cross-section construction suggests that the depth of basement is located at ~4.2
665 km bsl under the axis of the Thoard syncline (Fig. 6c) which is consistent with previous cross-
666 section constructions in this zone (Combes, 1984; Graham et al., 2012). Its depth decreases
667 progressively toward the south, at ~2.2 km bsl under the Mirabeau-Mallemoisson anticline
668 and at ~1 km bsl in the Les Mées basement thrust. Southward, the depth of basement-cover
669 interface goes down at ~2.5 km bsl, then decreases gradually toward the south (Figs. 6c and
670 19).

671 **5. Restoration of the Mesozoic extensional basin framework**

672 The balanced cross-section has been restored just before the Provençal compression
673 considering the top of upper Santonian marine strata as a regional flat template from the
674 South Provence basin domain (preserved from the Beausset syncline to the southern edge of
675 Arc syncline) to the Vocontian domain to the north (Digne Nappe). This template connects
676 with the subaerial erosional flat surface of the Durance high, which was used to unfold the
677 Mesozoic-Jurassic sedimentary cover in the northern Provençal belt, the Valensole high and
678 the Barles zone (Figs. 4 and 20a).

679 The pre-shortening restoration illustrates the initial extensional framework with a large
680 uplift zone (Durance high and Valensole high) in the center of the section, which led to the
681 separation of the Beausset basin (as part of the South Provence basin) to the south and the
682 Digne basin (as part of the Vocontian basin) to the north (Fig. 20a). The southern Beausset
683 rift-related structures and the southern edge of the Durance high have preferentially
684 accommodated the Provençal shortening along the Bandol thrust, Sainte-Baume Nappe and
685 Aurélien Mount thrust whereas the northern Digne rift-related structures have preferentially
686 accommodated the Alpine shortening of the Digne Nappe. Restoration of the Beausset basin
687 geometry is well determined by rift sequences preserved from erosion in the hanging-wall and
688 footwall of the Sainte-Baume Nappe (Fig. 20a). In contrast, cross-section restoration suggests
689 that a large part of the frontal Digne Nappe has been eroded (Fig. 20a). This eroded part
690 might correspond to the adjacent halokinetic Turriers basin located to the west and
691 transported ahead of the Digne Nappe along the studied cross-section (Gidon, 1982; Célini et
692 al., 2020, 2021). The restoration suggests that the Digne basin was initially located at ~8 km
693 to the north from the Barre de Chine and Barles diapir (Fig. 20a; Gigot et al., 1974).

694 In agreement with the structural data of Baudemont et al. (1985) and Espurt et al. (2019b),
695 cross-section restoration suggests that the basement incorporates a set of ~E- to ENE-trending
696 ~N-dipping normal faults associated with Permian half-graben systems (Fig. 20a). These
697 basement faults are reactivated intermittently during the Tethyan, Vocontian and South
698 Provence extensional episodes (Espurt et al., 2019b). They played a first order control on the
699 subsidence of the Beausset and Digne basins and thick syn-rift Jurassic-middle Cretaceous
700 sedimentary infilling (~2.4 km-thick in the Beausset basin and ~4.5 km-thick in the Digne
701 basin). This strong sediment load in the Beausset and Digne depocenters could have induced
702 the evacuation of the ductile Triassic evaporitic-carbonate layers from under the subsiding
703 zones toward local halokinetic features. This early halokinesis drives early folding of the

704 sedimentary pile including large subsiding synclines to narrow minibasins structures
705 characterized with growth wedges, normal faulting, and gravitational instabilities together
706 with local salt domes and diapirs initiated from basement faults (e.g., Saint-Cyr, Huveaune,
707 High Var and Barles halokinetic areas; Graham et al., 2012; Bestani et al., 2015; Espurt et al.,
708 2019b; Célini et al., 2020; Fig. 20a).

709 The southern edge of the Beausset basin show major E-W thickness and facies changes in
710 the Jurassic-middle-upper Cretaceous succession that depict depocenters and structural highs
711 with minor normal faulting (Philip, 1967, 1970; Philip et al., 1987; Espurt et al., 2019b). For
712 instance, the western Saint-Cyr Triassic high located in the footwall of the Bandol thrust, was
713 probably controlled by a N-dipping basement fault together with Triassic salt movement at
714 depth. Similar geometry has been used to restore the structures located in the footwall blocks
715 of the future Sainte-Baume Nappe and Aurélien Mount thrust (e.g., Huveaune Triassic high;
716 Guillemot et al., 1973; Bestani et al., 2015; Philip, 2019). In the Durance high, the halokinetic
717 activity during Late Jurassic-Early Cretaceous is more discrete with small sediment trapping
718 and normal faulting (e.g., future Pourrières and Les Maurras thrusts). Its result is the
719 formation of Triassic domes and long-wavelength syncline depocenters where the Jurassic-
720 Cretaceous platform shows successions with thickness changes (Fig. 20a; Espurt et al.,
721 2019b).

722 The pre-shortening restoration also suggests that fold structures observed in the footwall
723 (Barles diapir and Barre de Chine flap, Les Sauvans minibasin and Barles folds) and hanging-
724 wall (La Robine) of the Digne Nappe have been initiated by early halokinetic deformations
725 through the Jurassic and maybe until the Early Cretaceous (Fig. 20a; Graham et al., 2012;
726 Célini et al., 2020, 2021).

727 **6. Tracking superimposed compressions using syn-orogenic sedimentation** 728 **and fault slip data**

729 Deciphering the timing of deformations and shortening directions in the superimposed
730 Provençal and Subalpine thrust wedges are fundamental for produce a robust sequential
731 restoration model. The low-temperature thermochronology data available in the study area are
732 in the Oligocene-Miocene footwall rocks of the Digne Nappe, upper Cretaceous Arc basin
733 and Paleozoic basement of Maures and Tanneron-Estérel massifs (Jakni, 2000; Bestani, 2015;
734 Schwartz et al., 2017a). In the Maures massif, Oligocene-Miocene cooling ages are mainly
735 related to a reheating of the Paleozoic basement during the back-arc rifting of the Ligurian-
736 Provençal basin associated with Miocene volcanism (Jakni, 2000; Bestani, 2015). The cooling
737 ages in sedimentary cover are interpreted as detrital or partially reset ages by small reburial
738 below the Mesozoic-Cenozoic sedimentary cover or tectonic covering. In the Subalpine zone
739 thermal modeling of low-temperature thermochronological data suggested a cooling of the
740 Barles half-window at ~6 Ma (Schwartz et al., 2017a).

741 To constrain the timing of the thrusting, we used syn-orogenic basin infill and growth
742 strata (Suppe et al., 1992). The syn-orogenic deposits are either related to the Provençal or
743 Subalpine compression. They are described by previous studies (e.g., Gigot et al., 1974;
744 Dubois and Curnelle, 1978; Maillart et al., 1987; Haccard et al., 1989a; Espurt et al., 2012) or
745 highlighted in this work. The analysis of the temporal and spatial distribution of these
746 deposits was used to determine the growth and/or reactivation of contractional structures
747 along our section and the progressive advance of the thrust wedges toward the forelands (Fig.
748 21). In Provençal zone, there is no evidence of growth strata in Neogene deposits. However,
749 numerous examples of deformed Oligocene and Miocene-Pliocene strata testify to the
750 presence of Alpine contractional reactivation of pre-existing Provençal thrusts and Ligurian-
751 Provençal extensional structures (e.g., Hippolyte et al., 1993).

752 Fault slip data were used to determine regional stress regimes and the thrust transport
753 directions (Angelier, 1984, 1991). Variations in stress regimes through geological times can
754 be stratigraphically dated by reconstructing paleo-stresses at different stratigraphic levels
755 (Hippolyte et al., 1993). Furthermore, the fault slip analysis in growth strata deposits can
756 provide a direct dating of paleo-stresses (Hippolyte et al., 1992). Paleo-stress studies have
757 already been done in the Provençal zone in the Oligocene basins, the Arc syncline and the
758 Sainte-Victoire Mountain by Combes (1984), Gaviglio (1985), Lacombe et al. (1992),
759 Hippolyte et al. (1993), Guignard et al. (2005), and Espurt et al. (2012), and in the Subalpine
760 zone of Digne-Barles by Combes (1984), Fournier et al. (2008) and Hippolyte et al. (2012). In
761 this study, fault slip data were collected in zone with single deformation phase and far from
762 oblique structures or ramps producing local rotations and stress deviations (Lacombe, 2012)
763 as observed locally in the Sainte Victoire Mountain (Espurt et al., 2012). We measured
764 striated fault surfaces along the cross-section at twenty three new sites and computed paleo-
765 stresses using the Angelier method (INVD method, Angelier, 1990). We added to our data
766 five previously published fault sites along our cross-section that were analyzed with the same
767 method in the southern flank of Sainte-Victoire Mountain (Espurt et al., 2012), in the
768 Valensole basin (Hippolyte and Dumont, 2000), and in the Digne area (Hippolyte et al.,
769 2012). The trends of the computed shortening directions (maximum horizontal stress axes σ_1)
770 are presented in Fig. 21 and all computed data and diagrams in Table 1 and Fig. 22.

771 **6.1. Record of the Provençal shortening**

772 The map of Fig. 21a shows the Provençal compressional structures and associated upper
773 Cretaceous to Eocene syn-orogenic basins. In the southernmost part of the Provençal zone,
774 the transition from marine to continental deposits in the Latest Santonian of the Beausset
775 syncline could be related to basement stacking in the inner part of the orogen (Philip, 1970;
776 Philip et al., 1987; Tempier, 1987). The growth of the Sainte-Baume Nappe is recorded by

777 somewhat later Maastrichtian breccias (Rognacian facies) preserved in the folded footwall
778 block (Fig. 9b; Aubouin and Chorowicz, 1967; Corroy and Philip, 1964; Guieu, 1968; Philip
779 et al., 2018). Northward, Rognacian facies strata lie unconformably on middle-upper Triassic
780 rocks and record the uplift of the Huveaune antiform during the Late Cretaceous (Bestani et
781 al., 2015; Philip, 2018, 2019). Syn-tectonic deposits are not preserved along the Aurélien
782 Mount thrust. Because it overthrusts Maastrichtian (Rognacian facies) strata of the Arc
783 syncline (Fig. 10a), we can infer that this thrusting might have occurred during the Latest
784 Cretaceous like the neighboring Sainte-Baume and Etoile Nappes and then later during the
785 Paleocene (Fig. 3; Aubouin and Chorowicz, 1967; Guieu, 1968; Popoff, 1973). The initial
786 growth of the Pourrières thrust is attested by a major pinch out of the continental upper
787 Cretaceous sequences with intercalated breccia beds in middle Campanian strata (Begudian
788 facies) (Durand and Mennessier, 1964; Popoff, 1973). A later activity can be also inferred for
789 the Pourrières thrust as this thrust connects laterally to the S-verging Sainte-Victoire
790 Mountain thrust with Paleocene growth strata (Espurt et al., 2012; Fig. 11 a,b). The Paleocene
791 growth of the Pallières, Vinon, Les Maurras and Gréoux thrust systems is clearly recorded by
792 alluvial fan systems of microcodium breccias with local growth strata infilling the Rians,
793 Saint-Julien, La Mourotte and Esparron-de-Verdon synclines (Figs. 4, 11c and 12; Angelier,
794 1971; Philip et al., 2017). These thrusts and basins have recorded ongoing deformation maybe
795 until the Early Eocene (Angelier and Aubouin, 1976; Philip et al., 2017).

796 In the Valensole basin, subsurface data indicate that the Les Mées and Mirabeau-
797 Mallemoisson structures overprint Provençal thrusts which have been eroded and sealed by
798 basal Miocene strata (Fig. 19; Dubois and Curnelle, 1978; Gigot et al., 1981a). We can
799 speculate a similar pre-structuration of the neighboring Lambruissier and Aiglun structures.
800 Farther to the north evidence of Provençal deformation is also found in the Barles area (Fig.
801 3). In the Barles tectonic half-window, all E-trending compressional folds have been eroded

802 and unconformably covered by the Basal Breccia Formation with growth stratal architectures
803 (Lemoine, 1972; Maillart et al., 1987; Haccard et al., 1989a,b). For instance, onlaps and fan
804 shaped-geometry of the breccia beds can be observed in the northern limb of the Feissal
805 syncline west of Barles (Fig. 16b). Although the age of these deposits remains poorly
806 constrained (Latest Eocene-Early Oligocene) this occurrence of growth stratal wedging with
807 reverse faulting provides evidence for compressional syn-tectonic sedimentation in the
808 northernmost parts of the Provençal belt in Late Eocene times (Priabonian?). Interestingly,
809 well dated evidence for late Provençal shortening is found farther west in the E-trending
810 Eygalayes syncline of the Baronnies area (Fig. 3). Here, Montenat et al. (2005) described
811 Lutetian to Bartonian lacustrine and conglomerate deposits with growth stratal geometries
812 which recorded the initial folding of Eygalayes syncline during Middle-Late Eocene before its
813 reactivation during the Alpine compression.

814 To determine the trend of the compressional stress in the Provençal thrust wedge, we
815 measured fault-slip data at eleven sites into syn-tectonic conglomerates or growth strata
816 described above, and in Jurassic-Cretaceous units of clearly define Provençal thrusts (Fig.
817 21a). Our paleo-stress computation shows evidence for a regional ~NNE-trending
818 compression as previously shown by Gaviglio (1985), Lacombe et al. (1992) and Espurt et al.
819 (2012) in the Sainte-Victoire Mountain-Arc syncline area. Syn-depositional tilting and
820 faulting of the Danian growth strata in the Sainte-Victoire Mountain as well as in the Rians
821 basin (Figs. 11b, 21a and 22) show that this ~NNE-trending compression occurred at least
822 during the Danian (Espurt et al., 2012). In the Barles area, paleo-stress studies by Fournier et
823 al. (2008) and Hippolyte et al. (2012) show a complex polyphase history with structures with
824 different axial trends and ages interfering. Syn-depositional faulting in the Basal Breccia
825 Formation growth strata in the slightly deformed ~E-trending Feissal syncline indicates a ~N-
826 trending shortening (Fig. 16b). This trend is consistent with of the compression characterized

827 by Fournier et al. (2008) in the folded Mesozoic limestones of the Barles canyon. Thus, we
828 can infer that the ~NNE-trending Provençal compression propagated northward to the
829 outermost part of the belt in Late Eocene times (Fig. 21a) as previously suggested by
830 Montenat et al. (2005) for the Baronnies area.

831 **6.2. Record of the Alpine shortening**

832 The map of Fig. 21b shows the Alpine compressional structures and associated syn-
833 orogenic basins. In the Barles area, folding was recorded immediately after the Provençal
834 compression by growth stratal geometries in the Oligocene continental Red Molasse
835 Formation, then in the Grey Molasse Formation south of La Maurière anticline (Maillart et
836 al., 1987). This Oligocene deformation is consistent with growth stratal geometries in the
837 Esparron syncline as described by Gidon (1997) and Célini et al. (2021) more to the north
838 (Fig. 3). The growth of the La Maurière anticline (and probably also other inner folds) was
839 then recorded by Aquitanian-Burdigalian Lower Marine Molasse Formation, then Langhian,
840 Serravallian and Tortonian-lower Messinian conglomerates with growth stratal architecture in
841 the Vélodrome structure (Fig. 14; Gigot et al., 1974; Pairis and Gidon, 1987; Gidon, 1989;
842 Gidon and Pairis, 1992). West of La Robine half-klippe along the Duyes valley, growth strata
843 in upper Miocene subalpine conglomerates are identified in second order fold of the Thoard
844 syncline (Figs. 3 and 17; Haccard et al., 1989b). The upper Miocene-lower Pliocene
845 olistolithic Tanaron Formation which unconformably lies above folded Oligocene-upper
846 Miocene rocks, recorded the final growth of the Barles fold system before the emplacement of
847 the Digne Nappe (Gigot et al., 1974; Pairis and Gidon, 1987; Hippolyte et al., 2011; Fig. 14).

848 Based on thermal modeling of low-temperature thermochronological data from lower
849 Miocene rocks of the Vélodrome structure, Schwartz et al. (2017a) proposed that the thrusting
850 of the Digne Nappe started before the cooling of its footwall at ~6 Ma. They proposed that the
851 modeled upper Miocene cooling and exhumation was controlled by regional basement

852 tectonic thickening under the Valensole foreland basin. However, almost all of samples of the
853 Barles half-window have Late Pliocene or younger minimum apatite (U-Th)/He cooling ages
854 (Schwartz et al., 2017b) which is consistent with the proposed Late Miocene-Early Pliocene
855 age of the Tanaron Formation in the footwall of the Digne Nappe (Hippolyte et al., 2011).
856 Southward, the Aiglun, Mirabeau-Mallemoisson, Lambruissier, Les Mées and Asse structures
857 deform upper Miocene conglomerates of the Valensole 1 Formation (Fig. 18; Haccard et al.,
858 1989a). No growth strata that could support a syn-sedimentary shortening of these folds, was
859 found on the field and subsurface. This suggest that the growth of these folds occurred after
860 Late Miocene times (Hippolyte and Dumont, 2000; Godard et al., 2020).

861 Although Provençal and Oligocene structures are often unconformably covered and sealed
862 by Miocene-Pliocene sequences in the Provençal zone, numerous evidence of contractional
863 deformations are observed. South of the Valensole foreland basin, the Gréoux, Les Maurras
864 and Vinon thrusts are unconformably covered by Miocene-Pliocene conglomerates of the
865 Valensole 1-2 Formations. Along the fault traces, these conglomerates show striated and
866 sheared pebbles (Figs. 21b, 22 and 23a), suggesting small reverse-reactivation of the inherited
867 thrusts. The eastern edge of the Vinon thrust, the Rians syncline and Pallières thrust are also
868 unconformably overlain by Oligocene conglomerate beds of the La Combe half-graben (Fig.
869 3; Angelier and Aubouin, 1976). The conglomerate beds are also tilted and cut by reverse
870 faults (Fig. 23b). This evidence of Alpine shortening is consistent with perched Miocene
871 strata on the Mont-Major and Pallières thrusts; similar strata are located ~100 m lower on the
872 Valavés anticline to the north (Fig. 3; Angelier and Aubouin, 1976). South of the Aurélien
873 Mount thrust, Miocene conglomerates, sandstones and shales are folded and faulted, and
874 unconformably overlie Campanian shales (Begudian facies; Fig. 23c) or middle Triassic
875 Muschelkalk limestones (Espurt et al., 2019b). To the south, in the Signes and Méounes
876 grabens, that belong in the Sainte-Baume Nappe, Oligocene-Miocene strata are locally

877 exposed (Nury and Rousset, 1986). At the northern edge of Signes graben, vertical to
878 overturned Oligocene-Miocene conglomerates unconformably overlie Campanian-
879 Maastrichtian continental strata (Begudian-Rognacian facies; Répelin, 1922) (Fig. 24a). These
880 field data reveal Alpine contractional deformation in this zone. Evidence of Alpine shortening
881 are also observed in the Oligocene basins of Saint-Zacharie, Peypin, Destrousse, Aubagne and
882 Marseille (Fig. 3). The Oligocene strata infilling these basins are generally affected by normal
883 faults (Hippolyte et al., 1991), but they are also later folded (Guieu, 1968; Weydert and Nury,
884 1978; Dupire, 1985) and faulted by strike-slip and reverse faults (Hippolyte et al., 1993). For
885 example, the surface cross-section of Fig. 24b shows that the Peypin basin is an Oligocene
886 half-graben bounded to the southwest by the Mauvais Vallon normal fault. This Oligocene
887 basin develops on the eastern edge of the Etoile-Pierresca nappe system (Fig. 3), which
888 propagated northward above the Arc syncline during the Provençal compression (Aubouin
889 and Chorowicz, 1967; Guieu, 1968). The present day syncline shape of the Peypin basin
890 associated with disharmonic folds and strike-slip faults in the Oligocene strata results from
891 Alpine shortening. Similar deformed Oligocene (Miocene?) strata are also described
892 westward in the Etoile Nappe near the Les Maurins trough (Figs. 3 and 23d; Nury and
893 Raynaud, 1986; Hippolyte et al., 1993; Villeneuve et al., 2018). Alpine shortening is also
894 found at the southern edges of the Marseille and Aubagne basins (Figs. 3 and 23e). In these
895 areas, lower Cretaceous limestones of the Carpiagne anticline are interpreted as overthrust on
896 northward tilted to overturned and faulted Oligocene limestone and conglomerate beds
897 (Guieu, 1968; Weydert and Nury, 1978; Dupire, 1985; Villeneuve et al., 2018). A last
898 example of Alpine shortening is recognized in the southernmost Bandol basin along the
899 Mediterranean coast where Oligocene-Miocene conglomerates (Gouvernet, 1963) dip 20°
900 toward the south and are cut by numerous small-scale reverse faults (Fig. 23f).

901 To reconstruct the orientation of the Alpine compression along our cross-section in the
902 Subalpine zone and Provençal zone, we measured striated faults in seventeen sites in
903 Oligocene, Miocene and Pliocene strata (Fig. 21b). Paleo-stress computation reveals a
904 widespread ~NE-trending compression from the northern edge of the Valensole foreland
905 basin to the Aubagne and Marseille basins to the south, and up to the Bandol basin along the
906 Mediterranean coast. This ~NE-trending compression is contemporaneous with the deposition
907 of upper Miocene growth strata in the Thoard syncline (Fig. 17) and observed up to Pliocene
908 strata of the Digne area (Hippolyte et al., 2012). Generally, shortening direction may differ
909 from stress orientation depending for example on the amount of displacement on strike-slip
910 faults. It was demonstrated that a SSW transport direction of the Digne Nappe was related to
911 an about NE-SW trend of compression at thrust front (Hippolyte et al., 2012). The post-
912 Oligocene NE-SW trend identified in Provence is thus similar to the contemporaneous trend
913 of compression in the Alps. That locally the Miocene-Pliocene trend of compression was
914 similar to the upper Cretaceous-Eocene one may result from stress reorientation along
915 Provençal structures. The distinction between the two stress fields was possible only because
916 we worked in sites where the deformation can be dated. According to other works around the
917 study area, this ~NE-trending Alpine compression which propagated southwestward into the
918 Subalpine zone then, in the whole Provençal zone (Fig. 21b), probably prevailed since the
919 Middle Miocene (Combes, 1984; Gaviglio, 1985; Villeger and Andrieux, 1987; Ritz, 1992;
920 Champion et al., 2000; Clauzon et al., 2011).

921 **6.3. Modern compressional stress**

922 Beyond the predominant Miocene-Pliocene NE-trending compression presented above,
923 modern NE- and NNW-trending compressional stresses have been identified by previous
924 studies. Active shortening in the study area is revealed by earthquakes ($M_w < 3.5$) distributed
925 in the sedimentary cover and Paleozoic basement (Nicolas et al., 1990; Ritz, 1992; Baroux et

926 al., 2001; Cushing et al., 2008). The modern state of stress revealed by earthquake focal
927 mechanisms in the Digne-northern Valensole foreland basin area is consistent with NE-
928 trending shortening (Baroux et al., 2001; Cushing et al., 2008). It would be responsible for the
929 Quaternary deformation in the Digne Nappe east of the Bès fault, as well as at the
930 Lambruissier and Les Mées anticlines of the Puimichel plateau (Collina-Girard and
931 Griboulard, 1990; Hippolyte and Dumont, 2000; Hippolyte et al., 2012; Godard et al., 2020;
932 Fig. 18c). The Provençal zone is also characterized by a modern NNW-trending strike-slip
933 regime along the Middle Durance left lateral strike-slip fault (Ritz, 1992; Baroux et al., 2001;
934 Cushing et al., 2008; Mazzotti et al., 2020). This NNW-trending stress regime is consistent
935 with fault slip data collected in Pliocene rocks along the Middle Durance fault zone (Guignard
936 et al., 2005) or very locally in Oligocene rocks of the Marseille basin (Hippolyte et al., 1993).
937 The origin of the modern shortening in Subalpine and Provençal areas is still debated. Some
938 authors proposed that they could be controlled by the buoyancy forces in the core of the Alps,
939 driving compression in the foreland (Le Pichon et al., 2010; Walpersdorf et al., 2018) with
940 local deviations along pre-existing faults (e.g., Guignard et al., 2005), or related to the modern
941 Africa/Europe convergence (Ritz, 1992; Baroux et al., 2001; Mazzotti et al., 2020).

942 **7. Structural evolutionary model of the superimposed Provençal and** 943 **Subalpine foreland thrust wedges**

944 Based on the above mentioned structural data together with the review of deformation in
945 space and time, we built a sequentially balanced cross-section illustrating the long-term thrust
946 kinematic and shortening budget of superimposed Provençal and Subalpine thrust wedges
947 from Latest Santonian to present day (Fig. 20). The structural evolutionary model presented
948 here allows to discuss the role of crustal-scale rift-related anisotropies on vertical partitioning
949 of the shortening, foreland flexure and syn-orogenic sedimentation. The amounts of

950 calculated shortening/extension are minimum because the footwall and hanging wall cut-offs
951 of some structures are not clearly determined.

952 **7.1. Cross-section trace and thrust transport directions in the frame of the Provençal and** 953 **Subalpine compressions**

954 Drawing cross-sections approximately parallel to the slip direction, i.e., parallel to the
955 thrust transport direction is a prerequisite for cross-section balancing and restorations, and
956 quantifying shortening (Dahlstrom, 1969; Elliot and Johnson, 1980). However, it appears
957 relatively difficult to study superimposed orogenic deformations with a single and continuous
958 section. In this study, we proposed to show and evaluate the role of pre-existing structures on
959 subsequent deformations on a continuous cross-section with three segments of different trends
960 (Fig. 3) that respect the structural grain of the main structures and the shortening directions.
961 Thus, the cross-sectional trace chosen for this study integrates with tolerable obliquity the
962 different shortening directions of the two mountain ranges: the mean direction of compression
963 trends NNE in the Provençal thrust wedge while it trends NE in the Subalpine thrust wedge
964 (Fig. 21). Small variations in the direction of compressional are related to strike-slip
965 structures (e.g., Bès fault, site 14 in Fig. 21b; Hippolyte et al., 2011). The magnetic fabric
966 geometry also suggests SW-directed tectonic transport in all the Subalpine chains (Aubourg et
967 al., 1999). These trends are consistent with the trace of the balanced cross-section in the zone
968 where maximum horizontal shortening is recorded (Fig. 21 and see below). Therefore, our
969 balanced cross-section that trend ~N to NNE can be sequentially restored to illustrate the
970 structural evolution of Provençal and Subalpine superimposed foreland thrust wedges. To the
971 south, the cross-section is somewhat oblique to the Alpine shortening direction, which implies
972 that the amount of Alpine shortening is probably underestimated in this zone.

973 **7.2. Kinematic restoration of the Provençal thrust wedge**

974 The distribution of the upper Cretaceous foreland deposits permits to restore the first
975 episode of thrust propagation in the inner part of the Provençal foreland between Latest
976 Santonian and Maastrichtian (Fig. 20b). The transition from marine to continental deposits
977 described in the uppermost Santonian strata of the Beausset syncline could be related to inner
978 basement thrust stacking of the Provençal wedge related to the inversion South Provence
979 basin and basement thrusting. The emplacement of the Cap Sicié basement thrust fed at least
980 the slip of the Bandol cover thrust (other inner cover thrusts may have been eroded) above the
981 Saint-Cyr Triassic high. The activity of these thrusts probably occurred during the
982 Campanian-Maastrichtian because lower Campanian rocks (Fuvelian facies) are the youngest
983 strata in the footwall of the Bandol thrust. The initial growth of the Sainte-Baume Nappe and
984 deformation of Plan d'Aups and Pic des Corbeaux folds are recorded by upper Campanian-
985 Maastrichtian (Begudian-Rognacian facies) syn-tectonic conglomerates. It is fed by the deep
986 Saint-Mandrier thrust. The uplift of the Huveaune Triassic high is recorded by strata of
987 Fuvelian or Rognacian facies unconformably resting on the Triassic basement, although these
988 strata could have been transported onto the incipient Huveaune Triassic high during the
989 overthrusting of the Etoile Nappe (Fabre et al., 1975; Philip, 2019). The restoration shows that
990 the displacement of the deep basement Sainte-Baume thrust sheet was first accommodated by
991 the Arc syncline and then by the Pourrières frontal ramp in the Middle Campanian-
992 Maastrichtian. This frontal early growth might also result from the reactivation of basement
993 faults to the north, similarly to this proposed for the early reactivation of the Sainte-Victoire
994 Mountain deep basement thrusts (Espurt et al., 2012). Northward, the eroded area of the
995 Valensole high could be interpreted as a forebulge during the Latest Cretaceous (Fig. 20b).

996 The Paleocene-Eocene period (Fig. 20c) corresponds to the complete emplacement of the
997 southern basement thrusts and complete tectonic covering of the Sainte-Baume Nappe and
998 Aurélien Mount thrust. Shortening propagates northward in the cover units (Pourrières,

999 Pallières, Vinon and Gréoux thrusts) as recorded by Paleocene syn-tectonic conglomerates
1000 and subsiding thrust-top basins (i.e., Arc, Rians, Saint-Julien, La Mourotte, Esparron-de-
1001 Verdon synclines; Fig. 3). Although a huge volume of Triassic evaporites is suspected under
1002 the Pallières, Vinon and Gréoux units, the restoration suggests that Provençal thrusting also
1003 propagated northward at depth level into the basement of the Valensole high. The shortening
1004 recorded in cover of the outer part of the Provençal wedge was partly fed from the south (e.g.,
1005 Pourrières and Pallières thrusts) but was also controlled by the inversion of basement
1006 inherited fault system along S-verging thrusts with inferred shortcut trajectories. Although we
1007 have no growth strata in subsurface data because of the erosion at the base of the Miocene
1008 strata (Fig. 19), the initial growth of the Asse-Les Mées basement thrust system may have
1009 occurred with a similar structural style (S-verging basement wedge) during this stage. The
1010 growth of the Lambruissier, Mirabeau-Mallemoisson and Aiglun thrusts might also have
1011 started as small domes during this stage (Dubois and Curneille, 1978). According to field
1012 data, the restoration also shows the propagation of the deformation front up to the northern
1013 edge of Valensole high as shown by the reactivation of the halokinetic folds of the Barles area
1014 during the Late Eocene (Fig. 20c; Lemoine, 1972). Interestingly, the late Provençal shortening
1015 has been recorded west of Barles area by Lutetian-Bartonian growth strata in the small
1016 Eygalayes syncline (Montenat et al., 2005; Fig. 3). In addition, structural and stratigraphic
1017 data in onshore Languedoc region confirm that Provençal shortening was active up to
1018 Priabonian time (Séranne et al., 2021). An alternative interpretation is that the growth of the
1019 Barles fold system comes from to the north from the early Alpine belt (Ford et al., 1999).
1020 However, the N-trending compression reconstructed from growth strata (Fig. 16b), and the E-
1021 W trend of the folds in this zone are rather consistent with the Provençal compression, the
1022 Alpine compression being oriented more to the NE (Fig. 21). This outer Provençal shortening

1023 in the northern portion of the Valensole high up to and beyond the Barles zone implies that
1024 shortening was transmitted at depth through the basement under the Valensole high.

1025 The Provençal thrust wedge accommodated a minimum of 38 km (i.e., 17%) of total
1026 shortening (Fig. 20c). The main part of this shortening, i.e., 32 km, is accommodated by the
1027 inversion of the South Provence basin to the south in the Provençal zone, the remaining 6 km
1028 is accommodated in the Valensole high (Les Mées thrust) and Barles area. The sequential
1029 restoration (Fig. 20a,b,c) allow to calculate a shortening amount of 8.9 km between Latest
1030 Santonian and Maastrichtian (i.e., $\sim 0.5 \text{ km Ma}^{-1}$), and of 29.1 km between Danian and Late
1031 Eocene (i.e., $\sim 0.9 \text{ km Ma}^{-1}$).

1032 **7.3. Oligocene breakup of the Provençal zone**

1033 South of the Valensole high, most of the structural development in the Oligocene is related
1034 to the breakup of the Provençal thrust wedge by N-S, NNW-SSE and NE-SW extensional
1035 fault systems (Fig. 20d; Angelier and Aubouin, 1976; Giannerini, 1978; Nury, 1988;
1036 Hippolyte et al., 1993). The amount of extension remains small ($\sim 1.6 \text{ km}$) in the continental
1037 Provençal margin and focused south of the cross-section in the Provençal zone (Bestani et al.,
1038 2016). Most of the normal faults are reactivated thrust ramps, strike-slip faults (Roure et al.,
1039 1992) or diapiric features (Angelier and Aubouin, 1976; Philip, 2018, 2019), which
1040 accommodate the E-W opening of the West-European rift, then the NNW-SSE stretching of
1041 the southern Provence crust due to the Ligurian-Provençal back-arc rifting between Provence
1042 and Corsica-Sardinia block (Hippolyte et al., 1990, 1991, 1993; Bestani et al., 2016).

1043 **7.4. Kinematic restoration of the Subalpine thrust wedge**

1044 Following the Provençal shortening, the early orogenic building of the internal Alpine
1045 wedge during Late Eocene-Oligocene times has been probably associated with major flexural
1046 subsidence north of the Barles area (Fig. 20d; Ford et al., 2006; Graham et al., 2012).

1047 According to field data, the cross-section suggests small reactivation (less than 500 m of
1048 shortening) of the pre-existing halokinetic folds of the Barles area (Maillart et al., 1987;
1049 Gidon, 1997; Célini et al., 2021). This Oligocene deformation might be interpreted as distal
1050 shortening in the Oligocene Alpine foreland.

1051 The Late Miocene restored stage (Fig. 20e) shows the southward shift of subsidence into
1052 the Valensole high that became a large foreland basin. The syn-orogenic Miocene
1053 sedimentary package of the Valensole foreland basin exhibits a wedge-shaped geometry, with
1054 a more than 3 km-thick sequence in the depocenter zone and a few hundred meters in the
1055 south of basin. Southward, the Provençal zone was uplifted without fault reactivation. There,
1056 thin marine to continental sequences fill paleo-valleys overprinting the paleo-morphology of
1057 the Provençal thrusts, or cover marine abrasion surfaces (Mennessier, 1959; Cornet, 1965;
1058 Angelier and Aubouin, 1976; Besson, 2005; Oudet et al., 2010). The Provençal zone may
1059 represent a flexural forebulge, accompanying the subsidence of the Valensole foredeep
1060 depozone. This uplift is likely to have also been controlled by the Ligurian-Provençal passive
1061 margin setting to the south. The initial subsidence of the Valensole basin was first recorded by
1062 the transgressive Aquitanian-Burdigalian Lower Marine Molasse Formation on Mesozoic
1063 rocks previously folded and faulted during the Provençal compression. Subsidence continues
1064 through the Langhian to Tortonian (early Messinian?) with the deposition of the Upper
1065 Marine Molasse Formation that passes laterally and vertically to prograding fluvio-deltaic
1066 conglomerates of the Valensole 1 Formation. Synchronous syn-folding sedimentation is
1067 localized in the northern edge of the Valensole basin where growth strata in clastic sediments
1068 of the Vélodrome structure record the activity of the La Maurière anticline and other inner
1069 folds through the Miocene (Pairis and Gidon, 1987). Such deformation and uplift of the
1070 Barles fold system require the existence of basement thrust wedge under the Barles area.
1071 Although the ductile behavior of the Triassic detachment undeniably participated to the

1072 growth of the folds in cover, we propose that shortening was mainly controlled by the
1073 inversion of the deep Verdaches half-graben (named here the Verdaches horse) and the
1074 formation of an inferred frontal basement thrust (named the Barles horse) that propagated
1075 southward in the Triassic evaporites under the Thoard syncline. To the south of the Valensole
1076 basin, no growth strata are found within the upper Miocene conglomerate sequences,
1077 suggesting no contractional reactivation of the pre-existing Provençal structures.

1078 The comparison between the Late Miocene and present day stages (Fig. 20e,f) allows to
1079 evaluate the upper Miocene-Pliocene thrusting of the Digne Nappe above the Valensole and
1080 the southward transfer of the shortening into the Valensole foreland basin and Provençal zone.
1081 The olistolithic Tanaron Formation in the stratigraphically highest part of the Thoard syncline
1082 indicates strong erosion of the La Maurière anticline (Pairis and Gidon, 1987). The Digne
1083 Nappe moved above the Valensole basin mainly after the deposition of the upper Miocene-
1084 lower Pliocene Tanaron Formation. The present antiformal shape of the Digne Nappe is
1085 related to renewed basement thrusting under the Barles area, leading to the development of
1086 the present erosional Barles half-window. In this way, the Digne Nappe was folded later
1087 probably during the Pliocene by the underlying basement thrust wedge (Schwartz et al.,
1088 2017a,b). During the Pliocene-Quaternary, shortening of the basement thrust wedge was
1089 transferred southward through a Triassic décollement in the Valensole basin, leading to the
1090 complete development of the Aiglun, Mirabeau-Mallemoisson and Lambruissier anticlines.
1091 This period is also marked by the contractional reactivation of Provençal thrusts in the
1092 Valensole basin and Provençal zone, and inversion of Oligocene-Miocene basins as far as
1093 Aubagne, Marseille and Bandol basins along the Mediterranean coast. This outer Alpine
1094 shortening through the Provençal zone implies that part of the shortening was transmitted at
1095 depth through the continental crust.

1096 Globally, the kinematic restoration shows that the Subalpine thrust wedge is characterized
1097 by a first order shortening transfer from hinterland to foreland zones (Fig. 20d,e,f). However,
1098 the structural evolution of the Subalpine Thoard syncline and Digne Nappe show interaction
1099 between thrusting and syn-tectonic sedimentation and erosion disturbing here temporally this
1100 forward thrusting sequence as observed in many natural examples and analog and numerical
1101 models (e.g., Baby et al., 1995; Bonnet et al., 2008; Uba et al., 2009; Fillon et al., 2013; Erdős
1102 et al., 2015; Butler, 2020; Wu et al., 2021). The structural inversion in the southern margin of
1103 the Digne basin can be interpreted as a triangle zone model (McMechan, 1985; Price, 1986;
1104 Jamison, 1993). The large accumulation (up to 3.3 km-thick) of Miocene syn-tectonic
1105 sediments at the front of Barles basement triangle zone may have influenced the frictional
1106 properties (increase of the vertical stress) of the potential Triassic décollement under the
1107 Thoard syncline (Fig. 14b). We suggest that from Oligocene to Late Miocene this behavior
1108 may have prevented forward thrust propagation, localizing in a first step the deformation front
1109 at the northern edge of the Valensole basin in the Thoard-Vélodrome zone (Wu et al., 2021).
1110 When stresses were too strong and the Barles triangle zone could no longer accommodate the
1111 shortening by folding in the Thoard-Vélodrome zone, the broad Digne Nappe activated and
1112 cuts the pre-structured folds of the Barles area (Gidon and Pairis, 1992; Fig. 20e,f),
1113 characterizing an out-of-sequence kinematics as suggested by Vann et al. (1986). After the
1114 emplacement of the Digne Nappe, the development of the basement thrust wedge continued
1115 (folding of the Digne Nappe) and its shortening was transferred southward in the Aiglun,
1116 Mirabeau-Mallemoisson and Lambruissier cover structures.

1117 The Subalpine thrust wedge accommodated a minimum of ~35 km (i.e., 19%) of total
1118 shortening between Oligocene and present day (Fig. 20f). The main part of this shortening,
1119 i.e., 22 km, is accommodated by the Digne Nappe, 10 km in the Barles fold system and
1120 northern Valensole foreland basin, the rest, i.e., 3 km, is accommodated southward by

1121 reactivated Provençal basement thrusts (south of the Lambruissier anticline). The sequential
1122 restoration (Fig. 20d,e,f) allows to calculate a shortening amount of 13.8 km between
1123 Aquitanian and Late Miocene (i.e., $\sim 0.6 \text{ km Ma}^{-1}$ for the Aquitanian-Late Miocene period),
1124 and of 21.2 km between the Late Miocene and present day (i.e., $\sim 2.1 \text{ km Ma}^{-1}$). The
1125 calculated upper Miocene to present day shortening rate is three times larger than the post-
1126 Zanclean slip (0.7 km Ma^{-1}) along the Bès fault deduced from the 2.3 km displacement of a
1127 dissected post 3.4 Ma infilled drainage network (Hippolyte et al., 2012).

1128 **7.5. Foreland basin development**

1129 The propagation of the Provençal and Subalpine thrust wedges led to tectonic loading and
1130 foreland basin flexure (DeCelles and Giles, 2001). Both thrust wedges exhibit similar flexure
1131 angle toward hinterland zones of $\sim 7^\circ$ (Fig. 20c,f). However, the Provençal and Subalpine
1132 thrust wedges are characterized by foreland basins with contrasted developments and
1133 geometries. The Provençal thrust wedge is characterized by distributed basement thrusts along
1134 the cross-section reworking numerous structures inherited from the Variscan belt and
1135 Permian-Mesozoic rifts. As previously suggested by Roure and Colletta (1996), the inversion
1136 of this complex structural framework might have favored the development of several upper
1137 Cretaceous to Eocene foreland depocenters (Beausset, Arc, Rians, Saint-Julien, La Mourotte,
1138 Esparron-de-Verdon basins; Fig. 3), incorporated into the thrust wedge (Fig. 20c). In contrast,
1139 the Subalpine thrust wedge is characterized by the stacking of the thick Digne Nappe and
1140 Barles basement triangle zone. This duplex induced strong tectonic loading which might have
1141 controlled the large flexure of the Valensole basin, partly incorporated into the thrust wedge
1142 (Fig. 20f).

1143 **8. Discussion**

1144 **8.1. Basement inheritance and Triassic décollement level**

1145 Inherited or neo-formed décollement levels have strong influence on the vertical
1146 partitioning of the shortening in fold-thrust belts (e.g., Torres Carbonell et al., 2017).
1147 Inherited discontinuities in the Paleozoic basement (inherited faults, shear zones, or
1148 metamorphic foliations) act as decoupling levels that connect at depth with lower crustal level
1149 (e.g., Espurt et al., 2008; Lacombe and Bellahsen, 2016; Bellahsen et al., 2016). Salt and
1150 ductile shale levels are also recognized as major decollement levels in the sedimentary cover.
1151 The tectonic style of both Provençal and Subalpine foreland thrust wedges consists in mixed
1152 thick- and thin tectonic styles: deep basement thrust wedges kinematically linked with
1153 shallow thrust wedges detached in middle-upper Triassic evaporites. The basal Muschelkalk
1154 and Keuper units constitute major decollement levels in the sedimentary cover (Caron, 1979;
1155 Gidon, 1982; Baudemont, 1985; Tempier, 1987; Gidon and Pairis, 1992; Gidon, 1997). The
1156 structural data and cross-section restorations reveal that the shortening propagated through
1157 heterogenous basement frameworks involving the reactivation of pre-existing structures
1158 related to Permian and Mesozoic rift events (Bathiard and Lambert, 1968). The thinner crust
1159 of the former Mesozoic rift basins (i.e., the Beausset and Digne basins) associated with
1160 ductile lower crust might have favored the inversion of basement normal faults and the
1161 formation of inner thick-skinned thrust wedges during convergence. This style of shortening
1162 is most likely to occur in a young lithosphere like Europe (Mouthereau et al., 2013;
1163 Mouthereau et al., 2021). Our kinematic model shows that the Valensole high, separating the
1164 Beausset and Digne basins, remained a basement block poorly deformed by successive
1165 Provençal and Alpine orogens (Fig. 20a). If we analyze the distribution of the Triassic
1166 evaporites in the study area, we observed that these latter have controlled the transfer of the
1167 thin-skin tectonic displacement into the cover and the geometry of Provençal and Subalpine
1168 thin-skinned thrust wedges. The Provençal and Subalpine external compressional structures
1169 developed along the pinchouts of the Triassic evaporites located onto the flanks of the

1170 Valensole high. The near absence of potential Triassic layer overlying the basement of
1171 Valensole high (Les Mées area) and the relatively thin Mesozoic cover (less than 1 km-thick;
1172 Figs. 19 and 20) seem to have prevented the propagation of the décollements at shallow level
1173 into the cover. The restoration suggests that both Provençal and Alpine compressional stresses
1174 have consequently been transmitted under the basement of the Valensole high (Fig. 20c,f)
1175 perhaps onto a deep crustal (lower crust?) detachment level (e.g., Lacombe and Mouthereau,
1176 2002; Mouthereau et al., 2013, 2014; Lacombe and Bellahsen, 2016; Laborde et al., 2019;
1177 Mouthereau et al., 2021). During the Oligocene-Miocene Ligurian-Provençal back-arc
1178 extension, the southern Provence crust (Maures massif) was reheated, thinned, and weakened
1179 (Guieu and Roussel, 1990; Jakni, 2000; Bestani et al., 2016). This may explain the
1180 involvement of the basement in the shortening related to the Alpine compression as far as the
1181 Marseille-Bandol zone to the south. Such outer thick-skin deformation related to thermal
1182 event was already suggested by Jourdon et al. (2014) in the external Castellane area (Fig. 1).
1183 In the same way, the Provençal shortening has been transmitted under the Valensole high up
1184 to the weakened zone of the Vocontian rift during the Late Eocene times.

1185 On a larger scale (Fig. 1), the cross-section presented in this study and previously
1186 published cross-sections constructed across the southwestern Subalpine thrust front (Lickorish
1187 and Ford, 1998; Laurent et al., 2000; Jourdon et al., 2014; Guyonnet-Benaize et al., 2015;
1188 Bestani et al., 2016) show that the irregular basement architecture inherited from Paleozoic
1189 and Mesozoic tectonic stages and the distribution of the Jurassic-Cretaceous basins underlain
1190 by Triassic evaporites controlled the along-strike geometry of the Subalpine thrust front
1191 (Roure and Colletta, 1996; Macedo and Marshak, 1999). The Diois-Baronnies-Western
1192 Provence Arc (Villegier and Andrieux, 1987) and Digne-Castellane Arc (Goguel, 1937)
1193 coincide with Jurassic-Cretaceous depocenters (more than 5600 m-thick in the western
1194 Provence, and 1300-2200 m-thick in the Castellane Arc; Ménard, 1979; Laurent et al., 2000;

1195 Guyonnet-Benaize et al., 2015; Bestani et al., 2016) whereas the Valensole high is overlain by
1196 a thin succession (150-1000 m-thick). We suggest that the shape of the Diois-Baronnies-
1197 Western Provence Arc (Villegier and Andrieux, 1987) and the Digne-Castellane Arc (Goguel,
1198 1937) has been preferentially controlled by these paired Jurassic-Cretaceous basins underlain
1199 by huge volumes of Triassic evaporites. It was well established that the southern units of these
1200 arcs accommodated Alpine shortening in Miocene. In the Western Provence Arc, the
1201 Luberon, Costes and Trévaresse thrusts recorded shortening in Middle-Late Miocene as
1202 revealed by growth strata deposits and uplifted staircase wave-cut platforms (Anglada and
1203 Colomb, 1967; Champion et al., 2000; Besson, 2005; Clauzon et al., 2011). In the southern
1204 units of the Castellane Arc, growth strata in the upper Miocene-Pliocene conglomerates of the
1205 southeastern edge of the Valensole basin, and Eoulx-Brenon, Roque-Esclapon and Aubarède
1206 synclines more to the east have also recorded the Alpine shortening (Gigot et al., 1976;
1207 Giannerini et al., 1977; Giannerini, 1978; Dardeau et al., 2010; Fig. 1). In map view, the
1208 Valensole high can be interpreted as a reentrant between the Diois-Baronnies-Western
1209 Provence Arc and the Digne-Castellane Arc. The thinning of the Triassic evaporites onto the
1210 northeastern flank of the Valensole high prevented the large SW-ward propagation of the
1211 thin-skinned shortening, which was accommodated into the Aiglun, Mirabeau-Mallemoisson
1212 and Lambruissier anticline, probably recently and partly during the Quaternary for the
1213 Lambruissier anticline (Hippolyte and Dumont, 2000). The along-strike structural variations
1214 in the kinematic of the Subalpine thrust systems were accommodated by the development of
1215 NNE-trending strike-slip systems and oblique structures on lateral ramps on both sides of the
1216 Valensole foreland basin (e.g., Middle Durance fault zone and Bès fault; Laurent et al., 2000;
1217 Hippolyte et al., 2012; Guyonnet-Benaize et al., 2015; Fig. 1). Along the study area, the
1218 transfer of shortening into the basement of Provence induced the small reactivation of

1219 inherited basement faults under the central-southern Valensole basin and Provençal zone,
1220 causing their uplift during the Pliocene-Quaternary.

1221 Although the shape of the Eocene Provençal frontal thrusts has been strongly modified by
1222 the Alpine deformation, we also show that the along-strike configuration of the Mesozoic
1223 basins has controlled the ~NNE-ward propagation of the thin-skinned Provençal thrust front
1224 (Fig. 1). In the Western Provence Arc, balanced cross-sections of Bestani et al. (2016)
1225 illustrating the geometry of the Provençal fold-thrust belt at the end of Eocene show that the
1226 combination of thick sedimentary cover and Triassic evaporitic detachment layers favored the
1227 propagation of the thin-skinned thrust front up to the Baronnies area north of the Ventoux-
1228 Lure thrust system (Montenat et al., 2005). Similarly, restored balanced cross-sections of
1229 Laurent et al. (2000) and Jourdon et al. (2014) suggest that the Provençal cover detached
1230 easily northward up to the Vocontian basin thank to the Triassic decollements. During this
1231 period, the NNE-trending Middle Durance fault zone and probably the Barjols Triassic zone
1232 (Figs. 1 and 3) accommodated the along-strike kinematic of the Provençal thrust systems
1233 (Bestani et al., 2016).

1234 **8.2. Accommodation of the shortening by halokinetic structures**

1235 During shortening, the pre-existing halokinetic structures were generally reactivated,
1236 localizing preferentially the shortening in the cover. The initial geometry of these pre-existing
1237 halokinetic structures can be strongly modified during the contractional deformations as
1238 documented, for instance, in the Pyrenees (Saura et al., 2016; Labaume and Teixell, 2020;
1239 Burrel and Teixell, 2021), in Provence (Angelier and Aubouin, 1976; Espurt et al., 2019b), in
1240 the Alps (Dardeau et al., 1990; Granado et al., 2018; Célini et al., 2020, 2021), in the Zagros
1241 (Callot et al., 2007, 2012), and in the eastern Carpathians (Tămaş et al., 2021). Shortening
1242 induces the lateral and vertical evacuation of the mobile evaporitic material. Inherited salt
1243 dome structures constitute weak zones, which can evolve in diapirs with upturning flanks

1244 leading to thrust welding. Minibasins related to sedimentary loading evolve in tight synclines.
1245 As revealed by the palinspastic restoration of Fig. 20a and previous works in the study area,
1246 the Provençal and Subalpine thrust wedges incorporate numerous pre-existing halokinetic
1247 structures that initially developed during Mesozoic rifting stages (e.g., Graham et al., 2012;
1248 Bestani et al., 2015, 2016; Espurt et al., 2019b; Célini et al., 2020, 2021; Wicker and Ford,
1249 2021) and locally also during the Oligocene (Angelier and Aubouin, 1976). The Mesozoic
1250 halokinetic structures associated with variable thickness of Triassic evaporites and
1251 sedimentary cover undeniably generated anisotropic behavior in deformation during
1252 shortening. In the following, we discuss the role and evolution of some of these pre-existing
1253 halokinetic features in the Provençal and Subalpine thrust wedges.

1254 Halokinesis and its relationship with basement faulting has been previously analyzed and
1255 quantified by Espurt et al. (2019b) in eastern Provençal zone where a succession of early
1256 halokinetic folds have been later deformed by the Provençal and Alpine compressions. This
1257 work has provided new ideas leading to the reinterpretation of some structures located further
1258 west along the cross-section studied in this paper. For instance, the emplacement of the
1259 Bandol thrust was controlled by the Saint-Cyr Triassic high developed earlier at least up to
1260 the Early Santonian (Philip, 1967; Philip et al., 1987). The initial geometry of the dome has
1261 been squeezed and probably decapitated by the Bandol thrust. Shortening may have also
1262 driven salt migration and the northward expulsion of the Saint-Cyr Triassic evaporites onto
1263 the floor of the Beausset syncline (Fig. 8). The dip of the northern limb of the dome, made of
1264 lower Santonian limestone beds, has been enhanced during Provençal shortening and sheared
1265 by low-angle reverse faulting. Although we have no direct evidence of early halokinetic
1266 motion in the footwall block of the Sainte-Baume Nappe, we inferred that its emplacement
1267 might have been initiated on an early halokinetic structures above a N-dipping basement
1268 normal fault. However, we interpret the present day tight geometry of the Plan d'Aups

1269 syncline as mostly resulting from the emplacement of the far-travelled Sainte-Baume Nappe
1270 and contractional shear into its footwall block (e.g., Bertagne tectonic slice; Fig. 9b). A
1271 tectonic model implying an early growth (Jurassic?) of diapiric structures can be proposed for
1272 the Huveaune Triassic high (Bestani et al., 2015; Philip, 2018; 2019). This high is interpreted
1273 as an early dome controlled by a huge volume of Triassic evaporites overlying a basement
1274 fault (Guillemot et al., 1973). The restoration suggests that the dome has been strongly
1275 squeezed by Provençal shortening. As suggested by analog models in Callot et al. (2007), the
1276 initial shape of the dome controlled the propagation of the hanging-wall flat of the Aurélien
1277 Mount thrust through southern limb of the former dome. Surface exposure of the halokinetic
1278 structures after the Provençal compression might explain why some of them (e.g., the
1279 Huveaune or Barjols Triassic zones; Fig. 3), constituted by ductile material, have been easily
1280 reactivated during the Oligocene extension and Alpine compression (Angelier and Aubouin,
1281 1976; Philip, 2018, 2019; Espurt et al., 2019b).

1282 North of the Arc syncline, field data and cross-section restoration suggest that thrust
1283 systems are superimposed on Jurassic-lower Cretaceous salt domes and large synclines with
1284 local normal faults (Fig. 20a,b,c). Folding of the Rians syncline may have initiated during the
1285 Jurassic. However, its present tight geometry could rather result from the combined effects of
1286 halokinetic processes and thrusting (Cojan, 1993). We suggest that syn-tectonic sedimentary
1287 loading of the Paleocene-Eocene Rians syncline has favored its sinking into the thick ductile
1288 Triassic evaporitic-carbonate layer (Fig. 20c) as previously shown in the Sivas basin in
1289 Turkey by Kergaravat et al. (2016).

1290 In the Subalpine zone, cross-section restoration and field data also confirm that halokinetic
1291 features located in the hanging-wall and footwall of the Digne Nappe have localized and
1292 accommodated Alpine shortening. As suggested by the restored stage (Fig. 20a), the Digne
1293 Nappe involves frontal halokinetic structures (lateral continuation of the Turriers unit (now

1294 eroded) and La Robine; Célini et al., 2020). These structures are not only passively
1295 transported within the nappe but recorded major internal contractional deformation. The
1296 Barles diapiric structure is interpreted as a major halokinetic structure that localized the
1297 leading edge of the Digne Nappe (Fig. 20a; Graham et al., 2012). Beneath this thrust, and in
1298 contrast to the cross-sections of Graham et al. (2012) and Célini et al. (2020), the Barles fold
1299 system is not interpreted as an allochthonous unit transported above inferred deep cover sub-
1300 thrust. Although we have no subsurface data at Barles to highlight the thrust system geometry
1301 at depth, our construction favors a simple detachment fold interpretation as previously
1302 proposed by Goguel (1944) (Fig. 14b). The La Grande Cloche and La Maurière anticlines are
1303 detached from the underlying basement along Triassic evaporites. We agree that in places,
1304 folding has been initiated during Mesozoic rifting as a result of motion of the Triassic
1305 evaporites together with normal faulting as observed immediately north of the La Grande
1306 Cloche anticline in the Les Sauvans minibasin and Barles diapir. These pre-existing folds
1307 naturally localized contractional deformation during shortening. The present squeezed
1308 geometry of the detachment folds at Barles and complete overturning of the Barre de Chine
1309 flap were likely favored by ductile behavior of Triassic evaporites, but also by compaction of
1310 the ductile Callovian-Oxfordian shales in fold cores (Fig. 14c). Thus, the present day
1311 squeezed geometry of the Barles folds has been likely enhanced by shortening and the
1312 emplacement of the overlying Digne Nappe. Recently, Célini et al. (2020) interpreted the
1313 Neogene Vélodrome syncline as a mini-basin structure and the contact between the La
1314 Maurière anticline and the Vélodrome as a salt weld. These interpretations require a salt or
1315 allochthonous salt sheet Oligocene in age but the Oligocene strata exposed near Esclangon are
1316 fluvial deposits without any evidence of salt level.

1317 To sum up, Mesozoic halokinetic structures belonging in the South Provence and
1318 Vocontian extensional basins were rejuvenated and squeezed by Provençal and Alpine

1319 compressions. These structures localize preferentially the shortening and a significant amount
1320 of shortening might have been also accommodated by internal deformation within evaporitic
1321 layers. It appears therefore important to perform microstructural analysis into the evaporitic
1322 layers of salt-core structures to better constrain the amount of internal deformation
1323 accommodated within these structures during subsequent shortening (Malavieille and Ritz,
1324 1989; Rowan et al., 2019; Tămaş et al., 2021).

1325 **9. Conclusion**

1326 In this study, available surface and subsurface data together with the construction of a 150
1327 km-long sequentially restored balanced cross-section were combined to provide a regional
1328 overview of geology, structural evolution and quantification of crustal shortening budget of
1329 the superimposed Provençal and Subalpine fold-thrust belts in Central Provence region.
1330 Along the studied cross-section, the multistage growth of the fold-thrust belt systems has been
1331 remarkably recorded by well-preserved syn-orogenic deposits, growth strata and kinematic
1332 indicators of thrusting in the foreland basins. The major points arising from this study case are
1333 as follow:

1334 - The Provençal and Subalpine thrust wedges are characterized by a mixed thick- and thin-
1335 skinned tectonic style including basement thrusts propagating through inherited basement
1336 discontinuities (Variscan structures and normal faults delimiting upper Carboniferous-
1337 Permian basins, Mesozoic Tethyan, Vocontian and South Provence basins, and Oligocene
1338 basins) and cover thrusts detached in the middle-upper Triassic evaporites. The palinspastic
1339 restoration of the thrust wedges to Late Santonian shows a large uplift zone in the center of
1340 the study area (Durance high and Valensole high), which led to the separation of the Beausset
1341 basin (as part of the South Provence basin) to the south and the Digne basin (as part of the
1342 Vocontian basin) to the north. These extensional basins reactivated ~N-dipping normal faults
1343 related to upper Carboniferous-Permian basins.

1344 - The Provençal shortening propagated NNE-ward from the inverted Beausset basin up to
1345 the Barles area during Latest Santonian-Campanian to Late Eocene (i.e., ~50 m.y.). The
1346 Subalpine shortening propagated SW-ward from the inverted Digne basin up to the
1347 Mediterranean coast during Oligocene-Aquitainian to present (i.e., ~30 m.y.). In Provence, the
1348 Provençal and Alpine episodes of shortening are clearly separated by the Oligocene rifting
1349 events including the West-European rift followed by the back-arc rift of the Ligurian-
1350 Provençal basin destroying the inner Provence crystalline thrust wedge. In the studied area,
1351 NNW-SSE extension was mainly accommodated by crustal thinning and upper crustal ENE-
1352 trending normal faults together with major phases of erosion and sedimentation. In contrast,
1353 the recognition of upper Eocene, Oligocene and Miocene growth strata in the Subalpine zone
1354 of Barles might suggest a continuum of contractional deformation.

1355 - The amount of horizontal crustal shortening, derived from sequential restoration, is
1356 relatively similar in both thrust wedges: 38 km (17%) in the Provençal thrust wedge and 35
1357 km (19%) in the Subalpine thrust wedge.

1358 - By distributing the deformation through numerous inherited Permian-Mesozoic basement
1359 structures, the Provençal thrust wedge might have favored the development of confined
1360 foreland basins, as the Arc and Rians basins. In contrast, the vertical stacking of the thick
1361 Digne Nappe and Barles basement triangle zone in the Subalpine thrust wedge might have
1362 controlled the large flexure of the Valensole foreland basin.

1363 - The sequential restoration suggests that the shortening has been mainly controlled by the
1364 inversion of the South Provence basin-Durance high (Provençal shortening) to the south, by
1365 the inversion of the Vocontian basin (Alpine shortening) to the north, and by other factors as
1366 the pinch out of the Triassic décollement level onto the Valensole high and early halokinetic
1367 structures. At the scale of our cross-section, we show that syn-tectonic sedimentation has
1368 small and localized impact on thrusts kinematics (i.e., Digne Nappe).

1369 - This study highlights that the development of pre-shortening extensional basins and long-
1370 term structural evolution of the Provençal and Subalpine fold-thrust belts have been
1371 controlled by inherited crustal structures. These crustal heterogeneities have greatly
1372 influenced the propagation of the contractional deformation and foreland basin development
1373 in space and time.

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1389 **Data Availability**

1390 Datasets related to this article can be consulted by contacting J. Balansa (balansa@cerege.fr)

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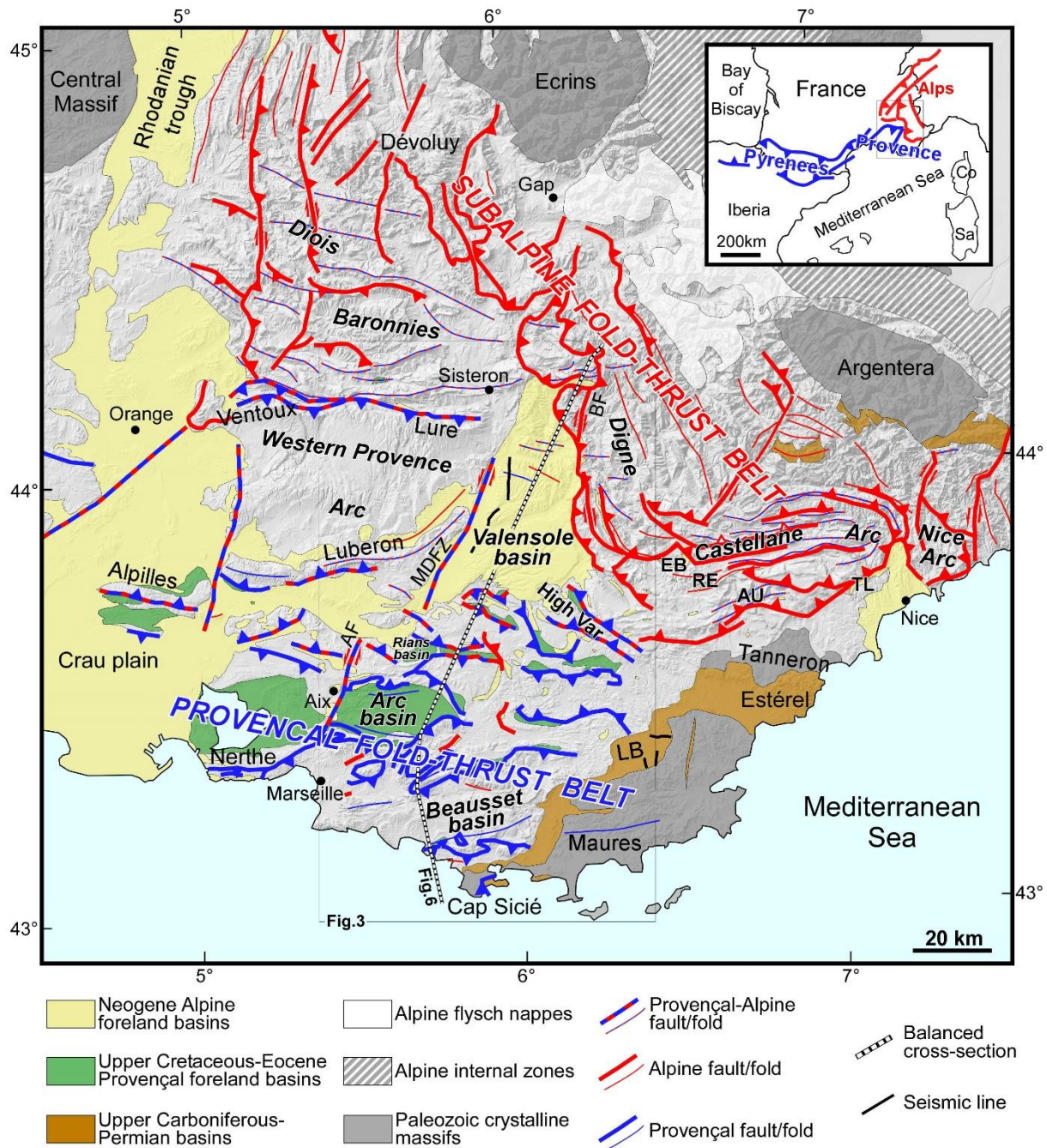
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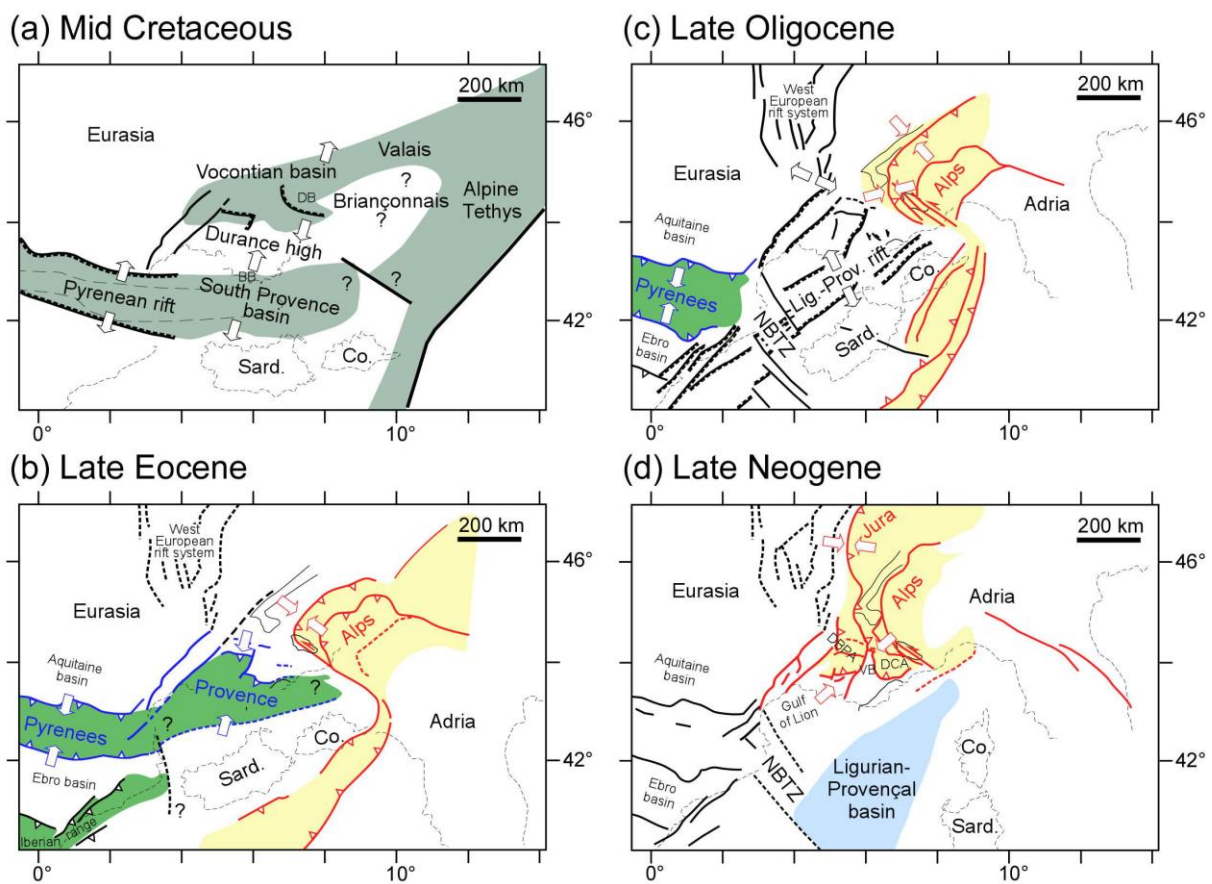
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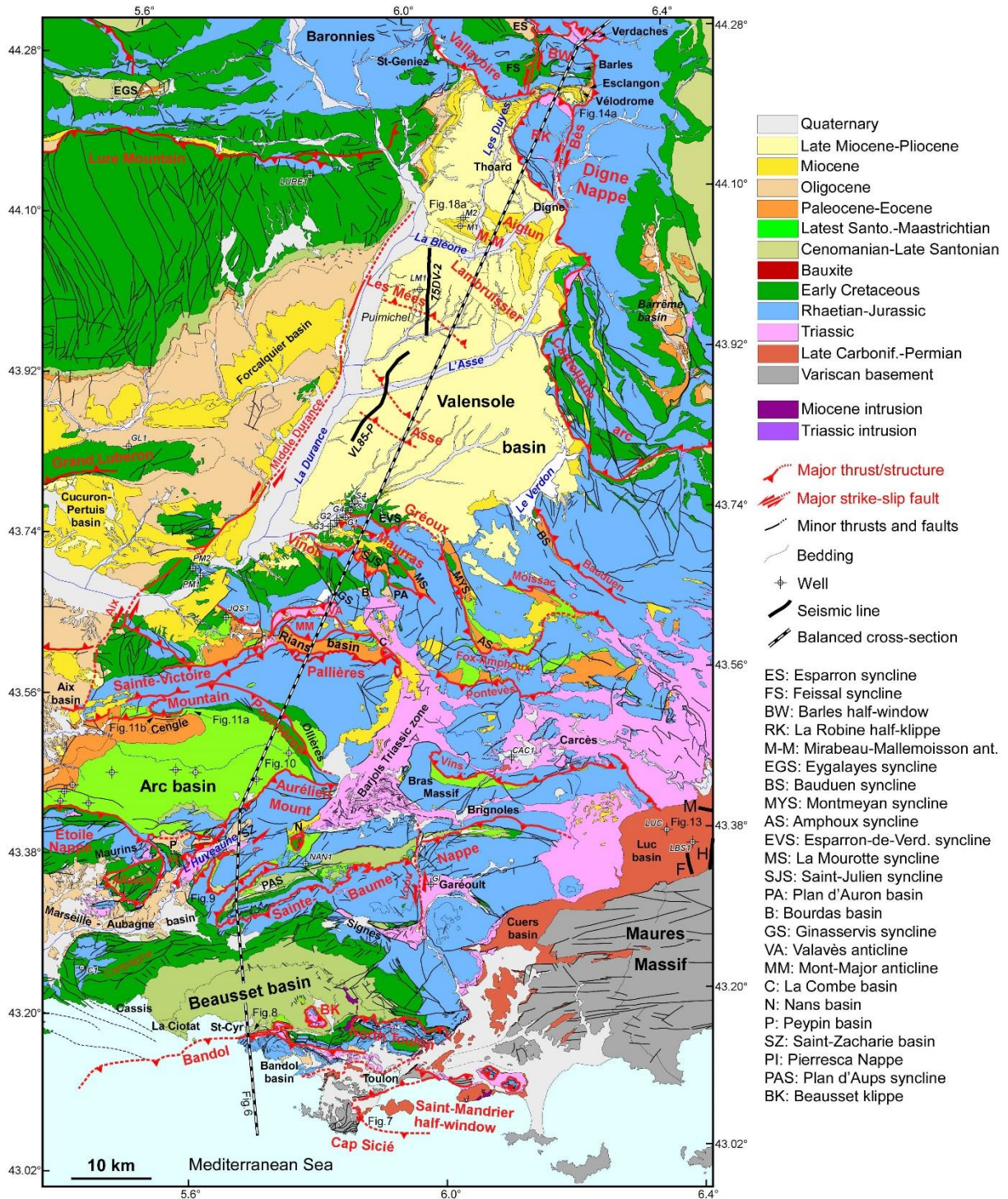
2330 **Fig. 1.** Synthetic structural map of the junction of the Provençal and Subalpine fold-thrust
 2331 belts in southeastern France. The map shows the traces of the main structures and main
 2332 polyphase structures compiled from Guieu (1968), Gidon et al. (1970), Lemoine (1972),
 2333 Gidon and Pairis (1976), Angelier and Aubouin (1976), Dubois and Curneille (1978), Dupire
 2334 (1985), Haccard et al. (1989a), Hippolyte et al. (1993), Lemoine et al. (2000), Champion et al.
 2335 (2000), Montenat et al. (2005), Chardon et al. (2005), Cushing et al. (2008), Espurt et al.

2336 (2012), Bestani et al. (2016), Villeneuve et al. (2018) and Espurt et al. (2019b), and from this
 2337 study. Thin black square indicates location of the geological map of Fig. 3. Dashed thick
 2338 black and white line shows the trace of the balanced and restored cross-section of Fig. 6
 2339 across the upper Cretaceous Beausset and upper Cretaceous-Eocene Arc and Rians foreland
 2340 basins and Neogene Valensole foreland basin. MDFZ: Middle Durance fault zone. AF: Aix
 2341 fault. BF: Bès fault. LB: Luc basin. Locations of Miocene-Pliocene syn-orogenic basins of
 2342 Eoulx-Brenon (EB), Roque-Esclapon (RE), Aubarède (AU) and Tourette-sur-Loup (TL) in
 2343 the Castellane Arc are also shown.



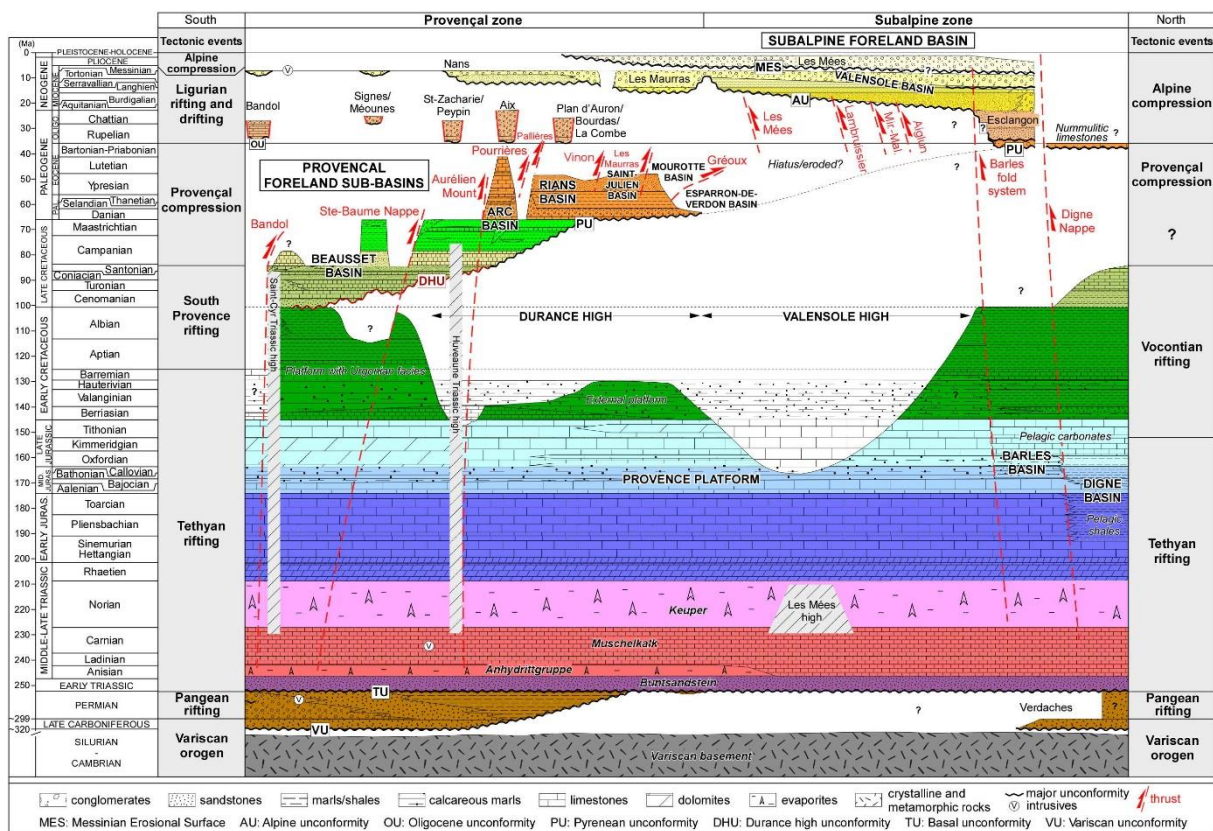
2344
 2345 **Fig. 2.** Geological map of the central Provençal and southwestern Subalpine zones. This is a
 2346 compilation of existing geological maps of BRGM (1/50000) of Mennessier (1963), Goguel
 2347 (1964), Flandrin et al. (1964), Dorkel et al. (1966), Mennessier and Modret (1966), Catzigras
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2349 Blanc et al. (1973), Blanc et al. (1974), Bordet et al. (1976), Blanc et al. (1977), Mennessier et
 2350 al. (1979), Gigot et al. (1981b), Rousset et al. (1983), Haccard et al. (1989a), Gigot et al.
 2351 (2013) and Laville et al. (2018), our field observations and subsurface data. Thick dashed
 2352 black and white line shows the trace of the balanced and restored cross-section of Fig. 6.



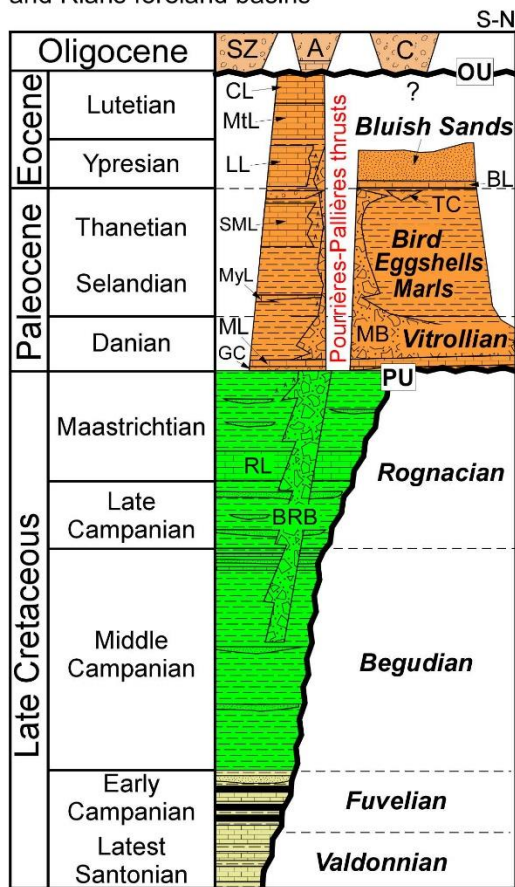
2353

2354 **Fig. 3.** Geological map of the central Provençal and Subalpine zones. This is a compilation of
 2355 existing geological maps of BRGM (1:50,000) of Goguel (1966), Rouire et al. (1969a,b),
 2356 Rouire et al. (1970), Rouire et al. (1974), Rouire et al. (1977), Rouire et al. (1979), Rouire et
 2357 al. (1981), De Graciansky et al. (1981), Gigot et al. (1981b), Haccard et al. (1989a), Rousset
 2358 et al. (1983), Gigot et al. (2013) and Laville et al. (2018), our field observations and
 2359 subsurface data. Thick dashed black and white line shows the trace of the balanced and
 2360 restored cross-section of Fig. 6.

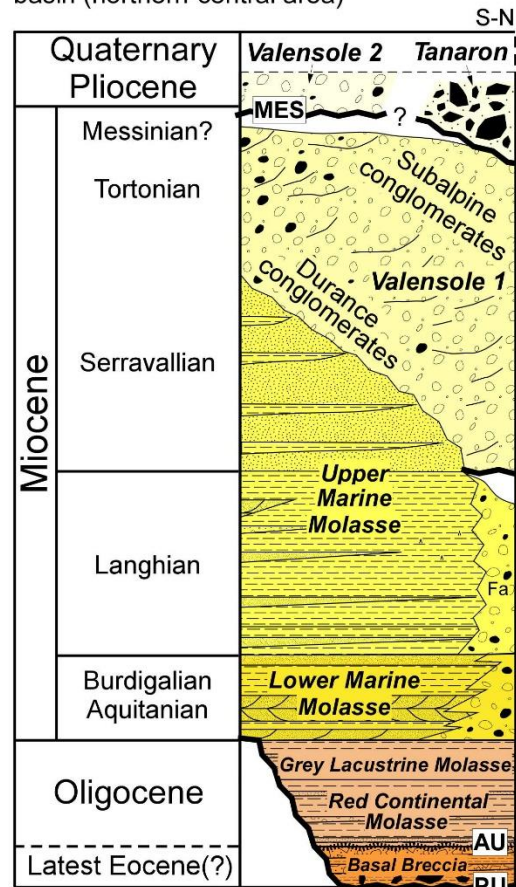


2362 **Fig. 4.** Synthetic stratigraphic chart along the studied cross-section across the Provençal zone
 2363 and southwestern Subalpine zone. Locations of the Oligocene-Miocene Bandol,
 2364 Signes/Méounes, Saint-Zacharie/Peypin, Nans, Aix, Plan d'Auron/Bourdass/La Combe basins
 2365 are shown in Fig. 3. See details in text for references.

(a) Upper Cretaceous-Eocene infill of the Arc and Rians foreland basins



(b) Cenozoic infill of the Valensole foreland basin (northern-central area)



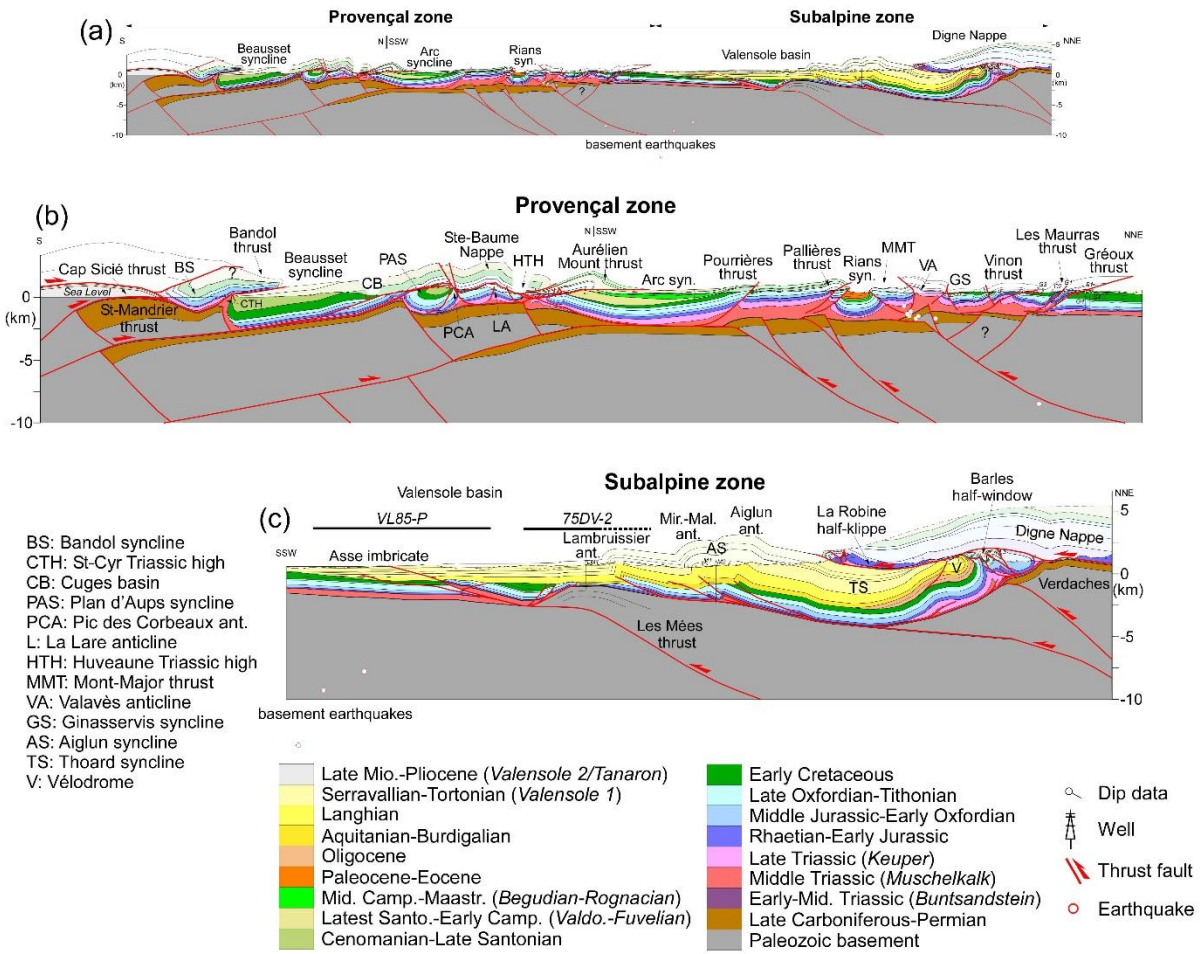
MES: Messinian Erosional Unconformity
 AU: Alpine unconformity
 OU: Oligocene unconformity
 PU: Pyrenean unconformity
 SZ: Saint-Zacharie basin
 A: Aix basin
 C: La Combe basin

CL: Cuques limestones
 MtL: Montaiguet limestones
 LL: Langesse limestones
 TC: Touars conglomerates
 BL: Bithynies limestones
 SML: Saint-Marc limestones
 MyL: Meyreuil limestones

MB: Microcodium breccias
 ML: Microcodium limestones
 GC: Galante conglomerates
 RL: Rousset limestones
 BRB: Begudian-Rognacian breccias
 Fa: Facibelle

2366

2367 **Fig. 5.** Details of the syn-orogenic sedimentary successions of the (a) upper Cretaceous-
 2368 Eocene infill of the Arc and Rians foreland basins and (b) Cenozoic infill of the Valensole
 2369 foreland basin. Locations of the Oligocene Saint-Zacharie, Aix and La Combe basins are
 2370 shown in Fig. 3. See details in text for references.



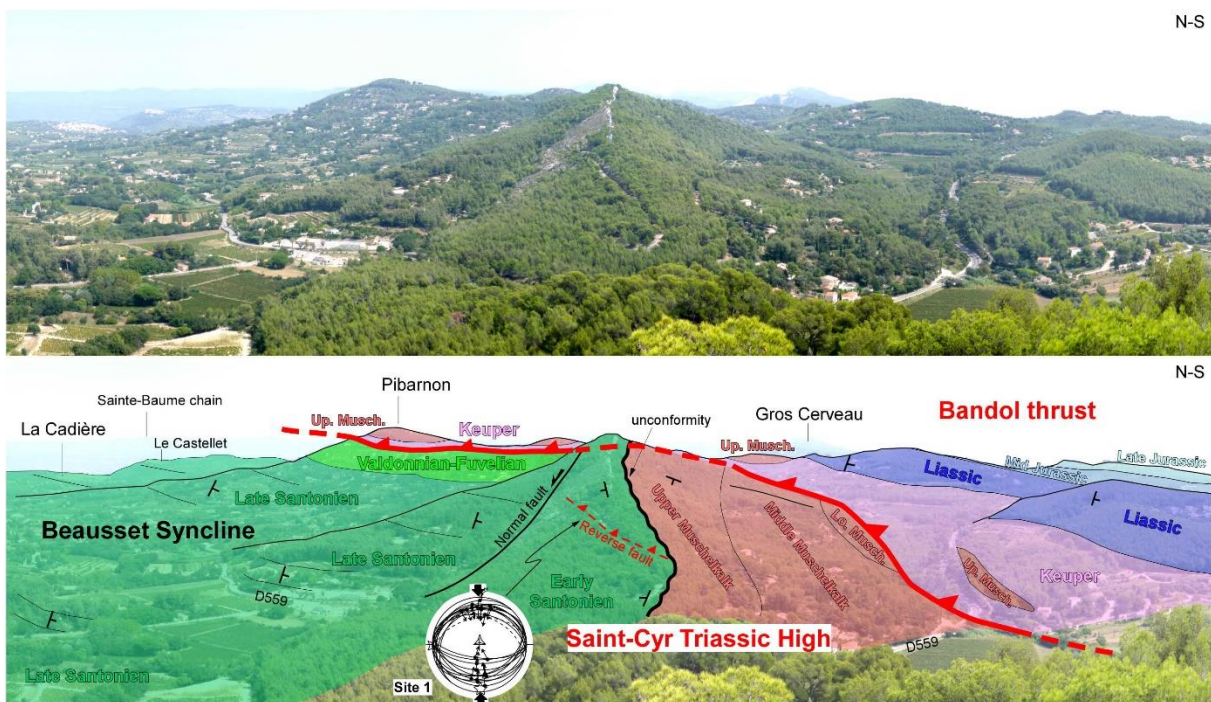
2371

2372 **Fig. 6.** (a) Cross-sectional geometry of the Provençal and southwestern Subalpine foreland
 2373 thrust wedges. Details of the (b) Provençal zone and (c) Subalpine zone. For location of
 2374 section trace, see Fig. 3. Location of seismic profiles VL85-P and 75DV-2 (Fig. 19) are
 2375 shown. Note that transparency has been applied to the eroded structures.



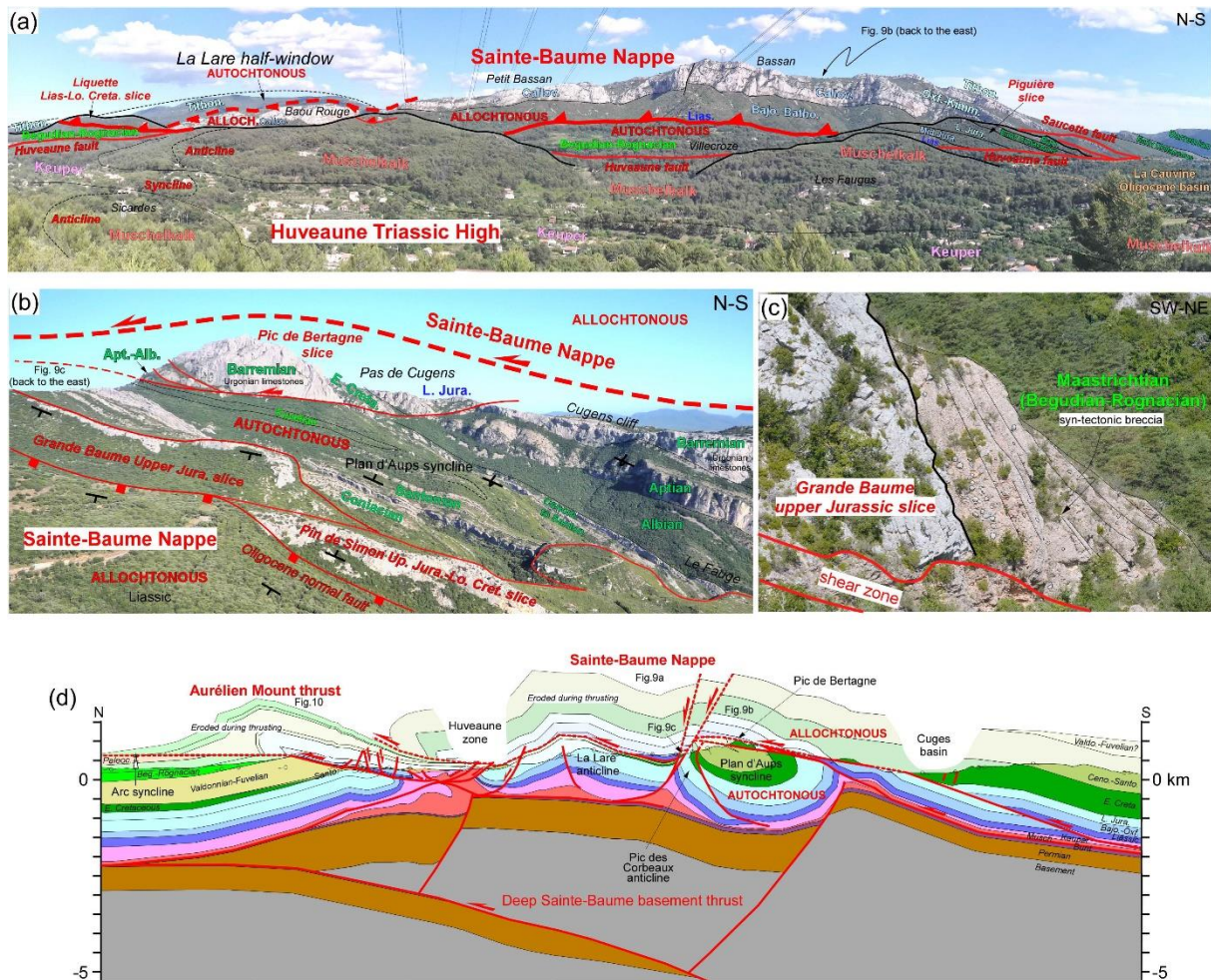
2376

2377 **Fig. 7.** (a) Panoramic view of the Saint-Mandrier half-window and folded N-verging Cap
 2378 Sicié basement thrust (view looking westward; 43.070431°, 5.907101°). For location, see also
 2379 Fig. 3. (b) Permian strata in the southern limb of the Saint-Mandrier anticline (43.067175°,
 2380 5.913673°). (c) North-dipping lower-middle Triassic (Buntsandstein) conglomerates and
 2381 sandstones at the front of the Cap Sicié thrust (43.112546°, 5.792338°). (d) Stratigraphic
 2382 contact between lower-middle Triassic sandstones (Buntsandstein) and middle Triassic
 2383 carbonates (Muschelkalk) at the northern front of the Cap Sicié thrust (43.113050°,
 2384 5.791026°).



2385
 2386 **Fig. 8.** Panoramic view of the Bandol thrust and southern limb of the Beausset syncline
 2387 (modified from Philip et al., 1985 and Espurt et al., 2019b) (view looking eastward;
 2388 43.165794°, 5.709031°). For location, see also Fig. 3. The N-verging Bandol thrust is
 2389 detached above upper Triassic evaporites (Keuper) embedding upper Muschelkalk carbonate
 2390 slices (e.g., Pibarnon) and overthrusts Valdonnien-Fuvelian facies strata of the Beausset
 2391 syncline. In the southern limb of the Beausset syncline, the Saint-Cyr Triassic high is formed
 2392 by middle Triassic (Muschelkalk) limestones, cargneules and volcano-detritic basaltic facies.

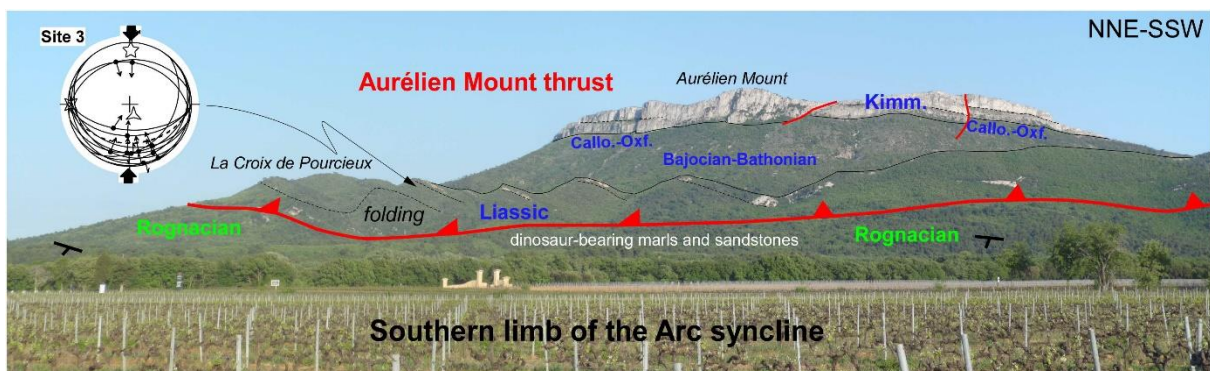
2393 The upper Muschelkalk fossiliferous limestones are unconformably overlain by lower
 2394 Santonian conglomerates and rudist bearing limestones. Striated faults measured in N-dipping
 2395 lower Santonian limestones are consistent with a N-trending compression. Fault data and
 2396 bedding (dashed line) have been projected in an equal area stereogram, lower hemisphere.



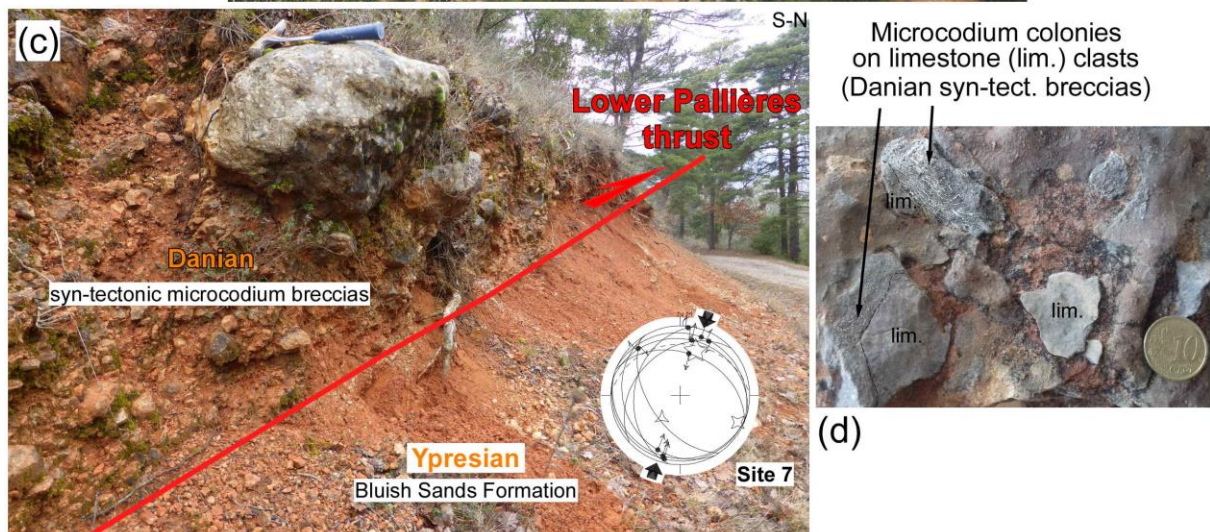
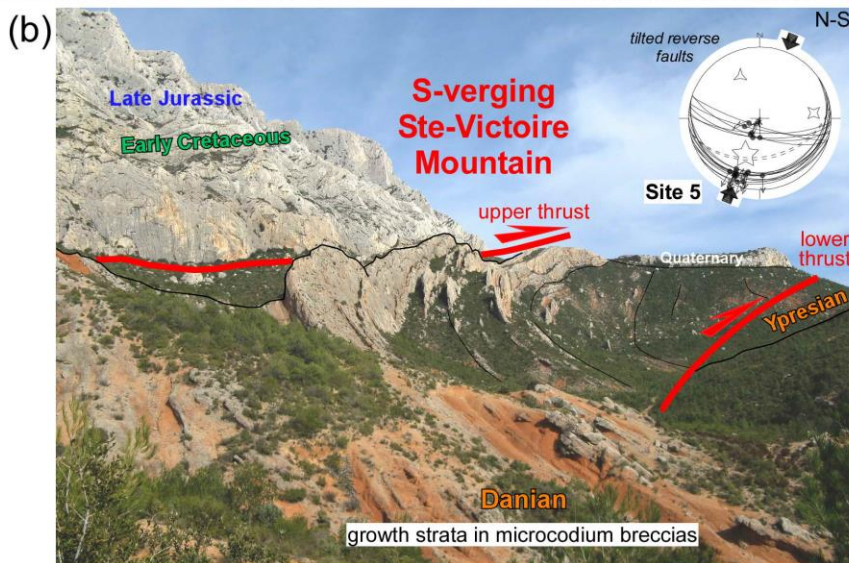
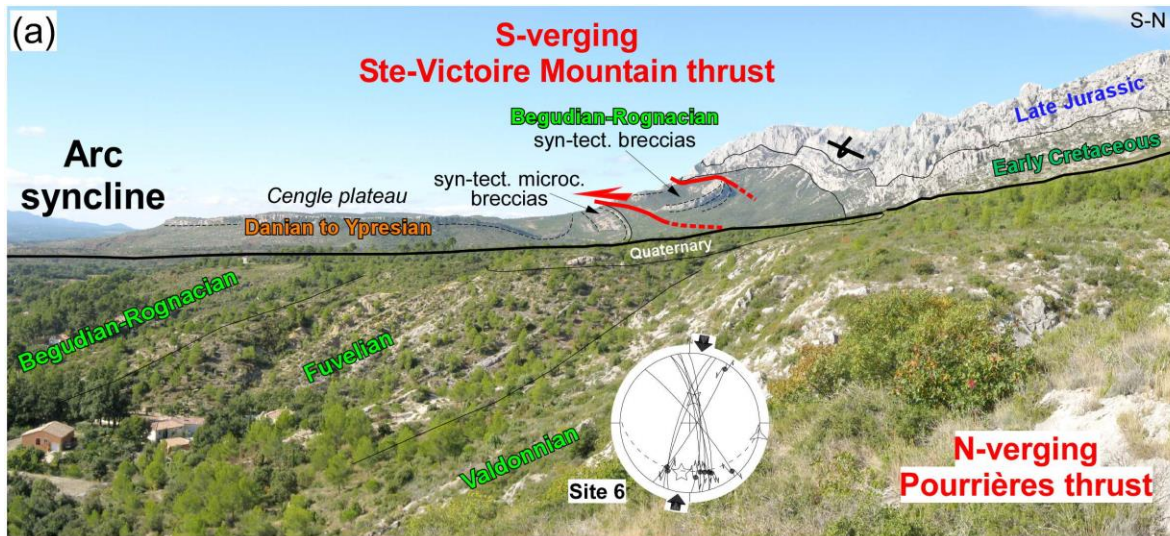
2397

2398 **Fig. 9.** (a) Panoramic view looking eastward of the Sainte-Baume Nappe and its analogous
 2399 units formed by the Pigièrre and Liquette slices (view looking eastward; 43.352896° ,
 2400 5.615643°). For location, see also Fig. 3. The nappe overthrusts northward Begudian-
 2401 Rognacian facies strata of the Villecroze zone tectonically separated from the Huveaune
 2402 Triassic anticlinorium high by the Huveaune fault. Oligocene conglomeratic strata of the La
 2403 Cauvine basin unconformably rest on the Triassic of the Huveaune high. Under the Sainte-
 2404 Baume Nappe, the La Lare half-window corresponds to a large autochthonous anticline armed

2405 of upper Jurassic and Cretaceous limestones. Geological data and structural features are from
 2406 Aubagne-Marseille 1:50,000 geological map, 3rd edition, 2018. (b) Panoramic view of the
 2407 southern Plan d'Aups half-window (view looking eastward; 43.312377°, 5.660989°). The
 2408 tight Plan d'Aups syncline located under the Pic de Bertagne slice and the Sainte-Baume
 2409 Nappe is cored by Fuvelian facies strata. The northern flank of the Plan d'Aups syncline is
 2410 truncated by the Grande Baume and Pin de Simon slices. Northward, an Oligocene normal
 2411 fault separates the collapsed Sainte-Baume Nappe from these slices. (c) Syn-tectonic breccias
 2412 (Begudian-Rognacian facies) embedding the Grande Baume upper Jurassic unit at the Source
 2413 de Cros (NW Plan d'Aups; view looking westward; 43.317931°, 5.689054°). The syn-tectonic
 2414 breccias occur on the northern flank of the Pic des Corbeaux anticline (location on Fig. 9d).
 2415 Photograph from Jean-Claude Tempier. (d) Zoom of the structural cross-section for more
 2416 clarity of the Sainte-Baume Nappe and Aurélien Mount thrust.



2417
 2418 **Fig. 10.** Panoramic view of the Aurélien Mount thrust overthrusting Rognacian facies strata
 2419 of the southern limb of Arc syncline (view looking eastward; 43.479551°, 5.73129°). For
 2420 location, see also Fig. 3. Striated faults measured in lower Jurassic strata are consistent with a
 2421 ~N-trending compression. Fault data and bedding (dashed line) have been projected in an
 2422 equal area stereogram, lower hemisphere. Geological data and structural features are from
 2423 Aix-en-Provence and Brignoles 1:50,000 geological maps.

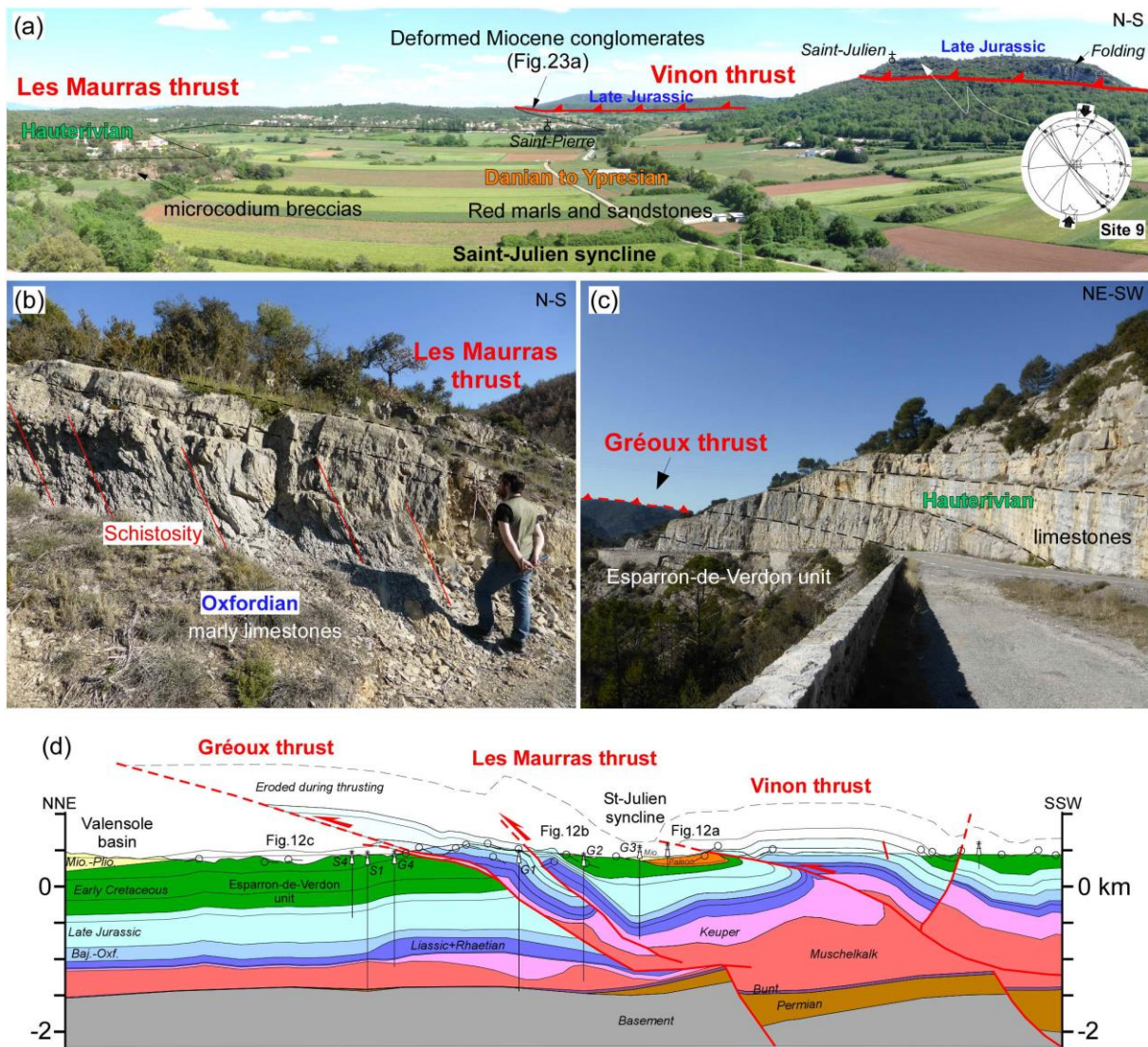


2424

2425 **Fig. 11.** (a) Panoramic view of the northern edge of the Arc syncline (view looking westward;

2426 43.529058°, 5.676583°). For location, see also Fig. 3. Westward, the syncline is thrust by

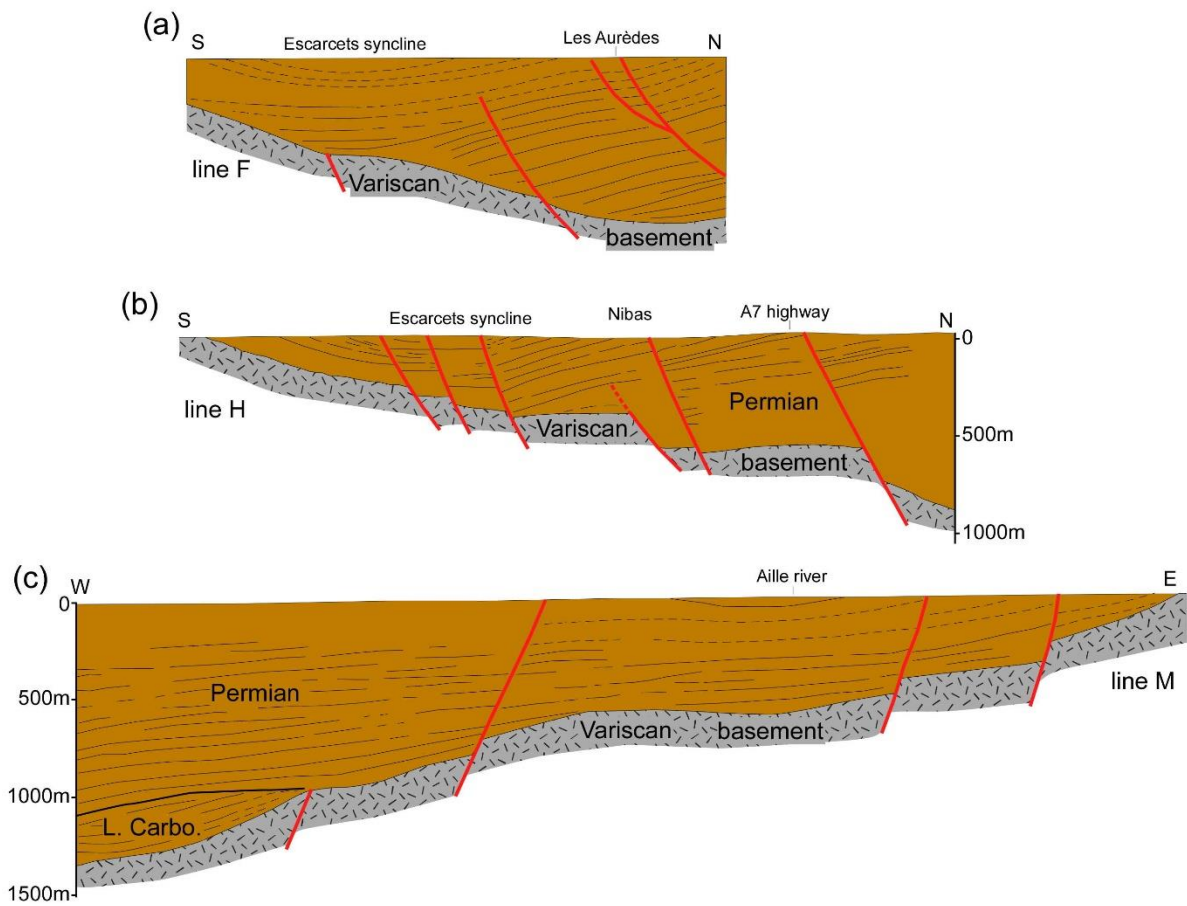
2427 the S-verging Sainte-Victoire Mountain thrust system, including Begudian-Rognacian and
2428 Danian microcodium syn-tectonic breccias. Eastward, the upper Cretaceous foreland rocks of
2429 the Arc syncline pinch-out on lower Cretaceous limestones. The syncline is transported
2430 northward onto the Pourrières thrust. (b) Panoramic view of the S-verging Sainte-Victoire
2431 Mountain thrust system with growth strata in Danian microcodium breccias (modified from
2432 Espurt et al. (2012); view looking eastward; 43.524806°, 5.569858°). For location, see also
2433 Fig. 3. (c) Danian microcodium breccias on the Pallières unit thrusting Ypresian Bluish Sands
2434 Formation (43.590979°, 5.837873°). (d) Details of the microcodium breccias including
2435 Mesozoic limestone clasts corroded by microcodium. Striated faults measured in Valdonnian
2436 limestones and Danian microcodium breccias are consistent with a NNE-trending
2437 compression. Fault data and bedding (dashed line) have been projected in an equal area
2438 stereogram, lower hemisphere.



2439

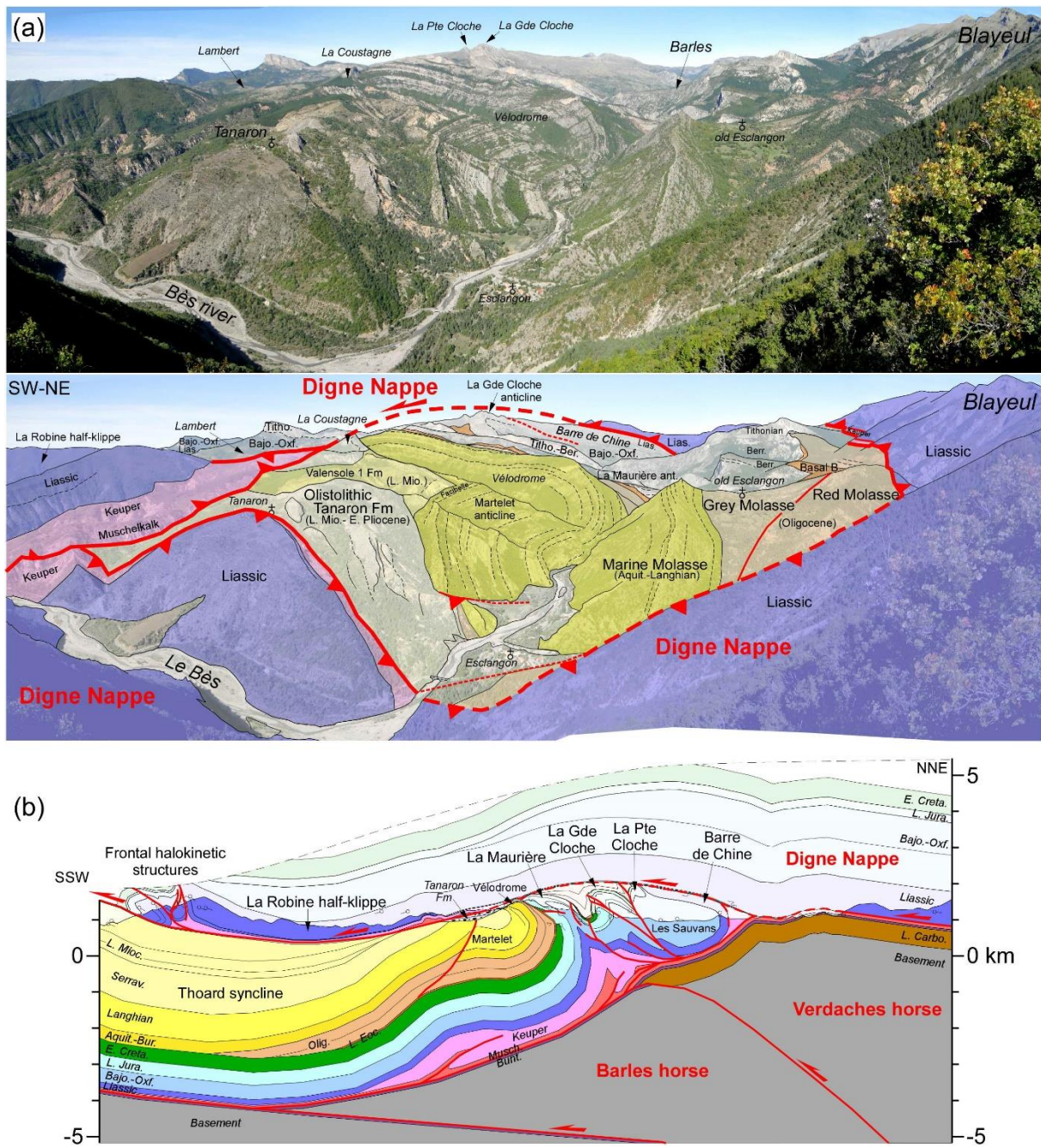
2440 **Fig. 12.** Field observations along the Vinon, Les Maurras and Gréoux thrusts. (a) Panoramic
 2441 view of the Vinon thrust overlying Paleocene strata in the Saint-Julien syncline (view looking
 2442 eastward; 43.702067° , 5.899911°). For location of the Saint-Julien syncline, see also Fig. 3.
 2443 The activity of this thrust was recorded at least by Danian microcodium breccias outcropping
 2444 in the norther limb of the Saint-Julien syncline. Striated faults measured in upper Jurassic
 2445 limestones of the Vinon thrust are consistent with a NNE-trending compression. Fault data
 2446 and bedding (dashed line) have been projected in an equal area stereogram, lower hemisphere.
 2447 Eastward, the Vinon thrust is covered by Miocene conglomerates with sheared and striated
 2448 pebbles (see Fig. 23a). Geological data and structural features are from Tavernes 1.50 000

2449 geological map. (b) Sheared middle Jurassic marly limestones in the hanging wall of the Les
 2450 Maurras thrust (43.737375° , 5.874535°). (c) N-dipping Hauterivian limestones in the footwall
 2451 of the Gréoux thrust (43.764164° , 5.928908°). (d) Zoom of the structural cross-section for
 2452 more clarity of the Vinon, Les Maurras and Gréoux thrusts.



2453

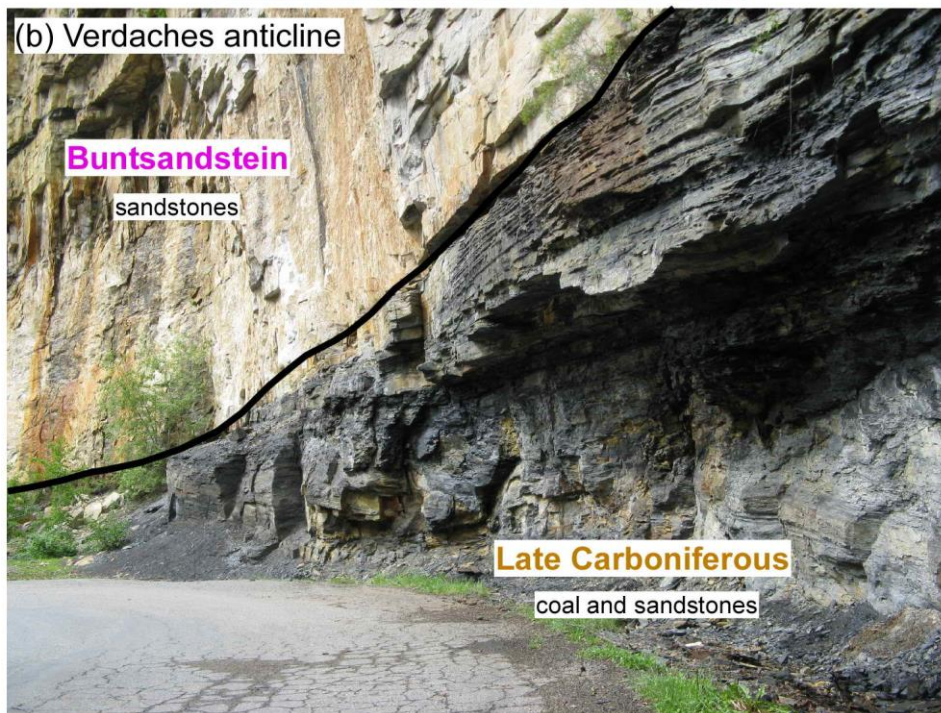
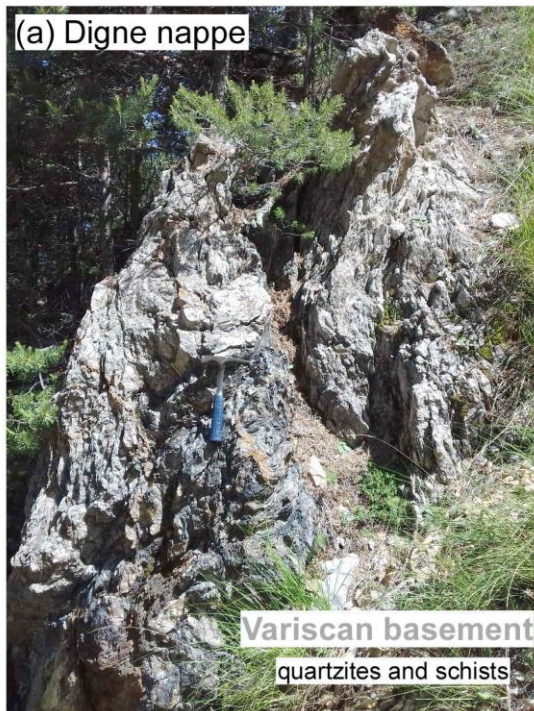
2454 **Fig. 13.** Interpreted cross-sections documenting the structure of the Permian extensional basin
 2455 of Luc along the seismic lines F (a), H (b) and M (c). Modified from Baudemont (1985). For
 2456 locations, see Fig. 3.



2457

2458 **Fig. 14.** Structures of Barles half-window and Digne Nappe. (a) Panoramic view of the Barles
 2459 half-window and folded Digne Nappe from l'Escuichière (view looking NW-ward;
 2460 44.190201°, 6.288481°). For location, see also Fig. 3. The cross-section trace is in the second
 2461 plan of this panoramic view. Along the Bès valley, the Digne Nappe overthrusts SW-ward the
 2462 Barles fold system (La Maurière, La Grande Cloche, La Petite Cloche and Barre de Chine
 2463 flap) and the Vélodrome structure and Martelet anticline, corresponding to the northern limb

2464 of the Thoard syncline. The Miocene strata of the Vélodrome structure (Marine Molasse and
2465 conglomerates of the Valensole 1 Formation) show sedimentary wedging. These strata are
2466 unconformably covered by the upper Miocene-lower Pliocene olistolithic Tanaron Formation,
2467 a chaotic melange of large middle Jurassic olistoliths with evaporites (Lambert locality),
2468 upper Jurassic-Berriasian limestone (e.g., La Coustagne) and Eocene-Miocene olistoliths
2469 included in red marls, sandstones and conglomerates beds. Geological data and structural
2470 features are from La Javie 1:50,000 geological map. (b) Zoom of the structural cross-section
2471 for more clarity of the more clarity of the Barles half-window and Digne Nappe.



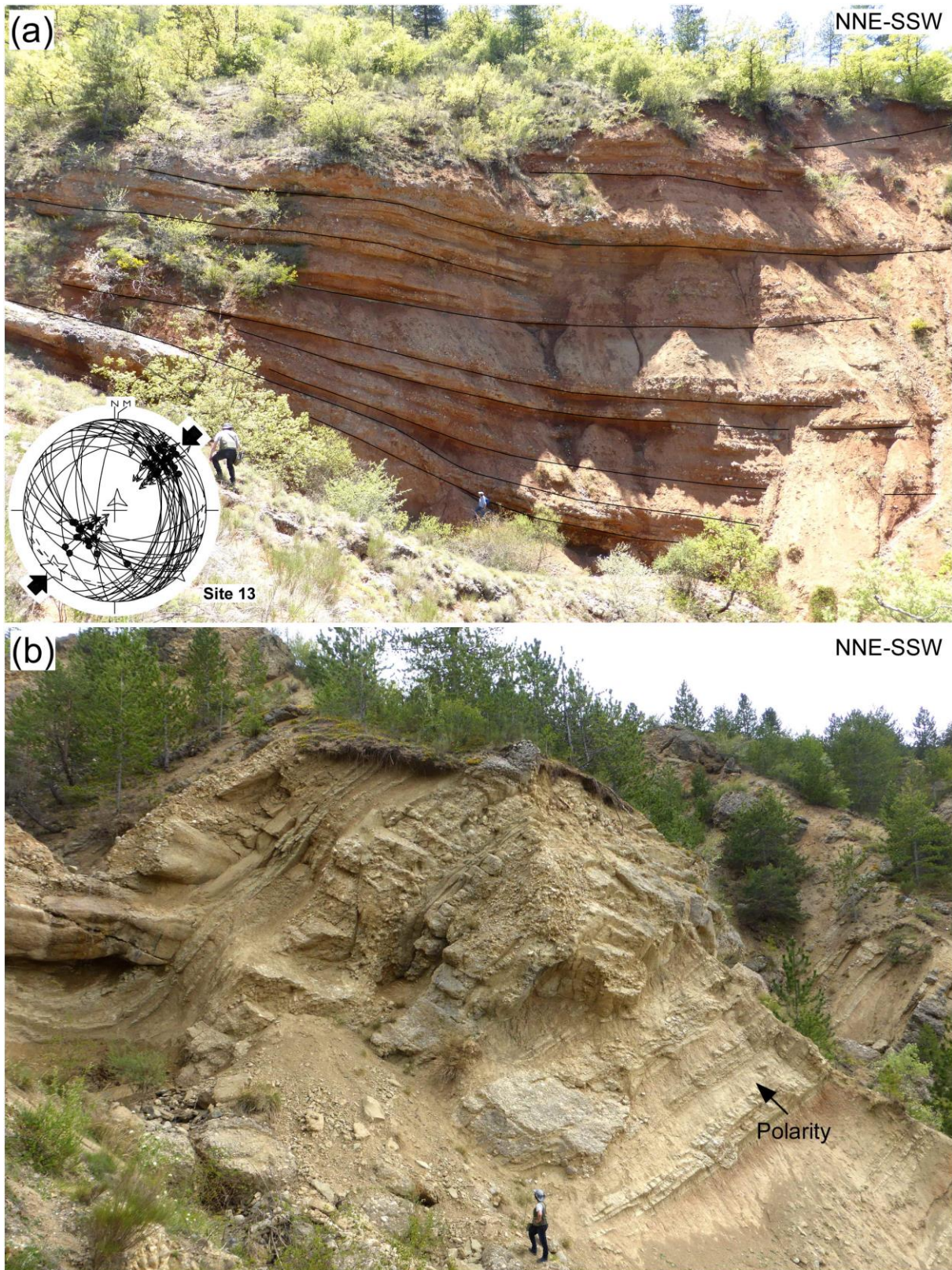
2472

2473 **Fig. 15.** (a) Tectonic slice of Variscan basement at the base of the Digne Nappe (Guiomar,
2474 1990) (44.263578°, 6.293548°). (b) Upper Carboniferous coal and sandstones unconformably
2475 covered by lower Triassic sandstones in the Verdaches basement anticline (44.273110°,
2476 6.299909°).



2477

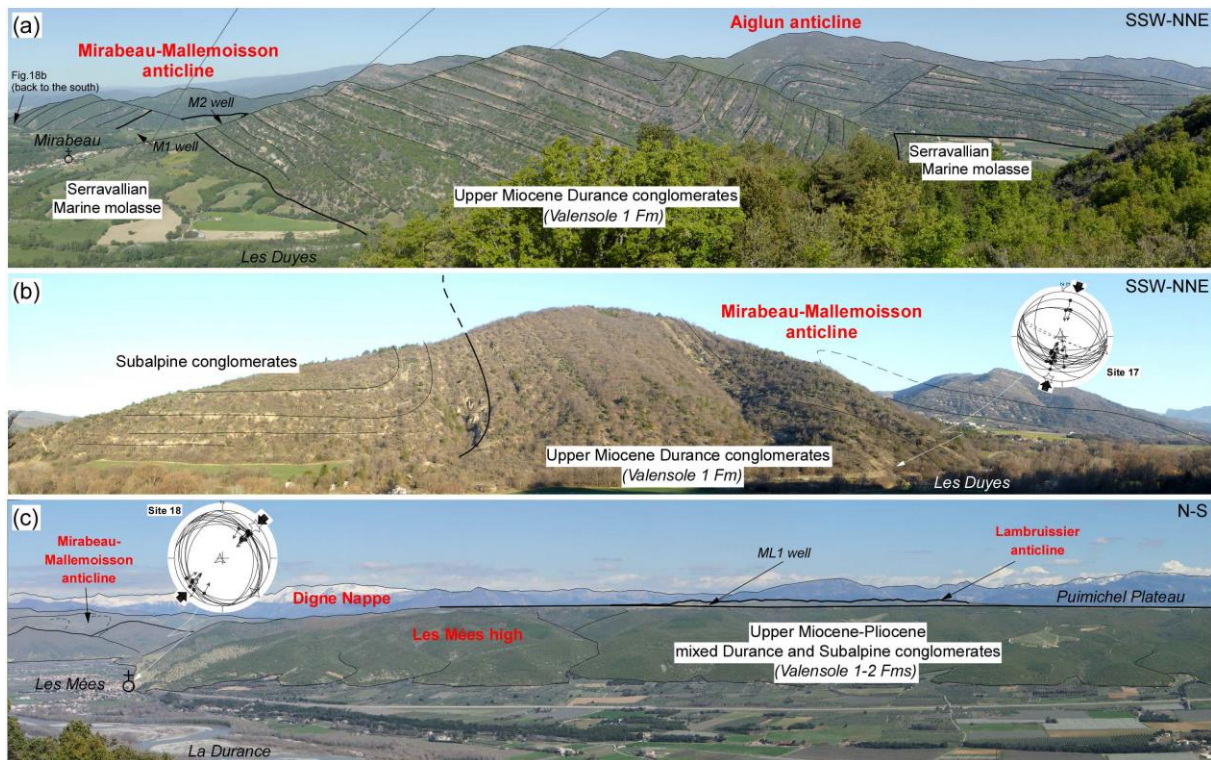
2478 **Fig. 16.** Outcrops of upper Eocene-lower Oligocene Basal Breccia Formation in the erosional
 2479 Barles half-window. (a) Hauterivian strata unconformably covered by the Basal Breccia
 2480 Formation on the southern limb of the La Grande Cloche anticline (44.228736° , 6.287899°).
 2481 (b) Fan-shaped geometry in the Basal Breccia Formation in the northern limb of the Feissal
 2482 syncline (44.246639° , 6.201392°). Striated faults are consistent with a syn-sedimentary NNE-
 2483 trending compression. Fault data and bedding (dashed line) have been projected in an equal
 2484 area stereogram, lower hemisphere. (c) Details of the breccias including Mesozoic limestone
 2485 clasts (Le Pas de Terre Rouge, southern limb of the La Grande Cloche anticline; 44.232062° ,
 2486 6.229132°).



2487

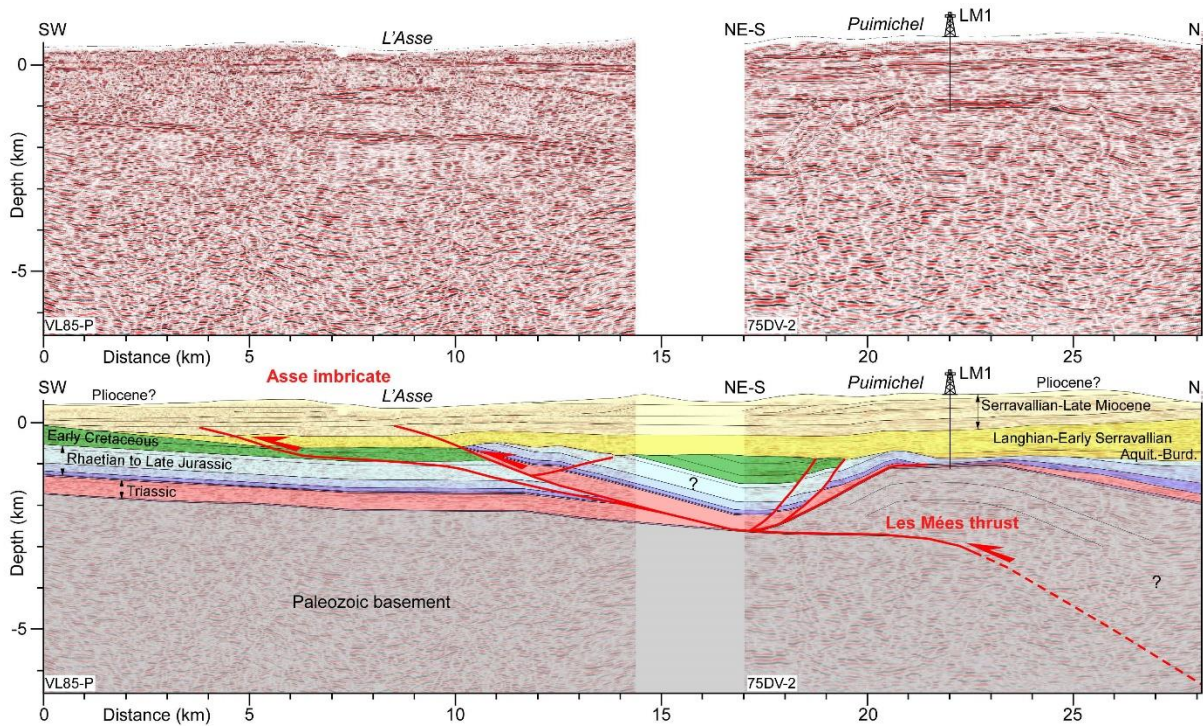
2488 **Fig. 17.** Upper Miocene (Serravallian-Tortonian) Subalpine conglomerate accumulations in
 2489 the Thoard unit along the Duyes valley west of the La Robine half-klippe (Fig. 3). (a) Growth
 2490 strata near Thoard town (44.164978°, 6.151822°). Striated faults in these Subalpine

2491 conglomerates are consistent with a NE-trending Alpine compression. The conjugated and
 2492 tilted reverse faults of the stereogram confirm the tectonic origin of the sedimentary fan. Fault
 2493 data and bedding (dashed line) have been projected in an equal area stereogram, lower
 2494 hemisphere. (b) Base of the syn-tectonic Subalpine conglomerates in the Bramefan-Mauduech
 2495 ravine, high Duyes valley (44.181431° , 6.174494°).

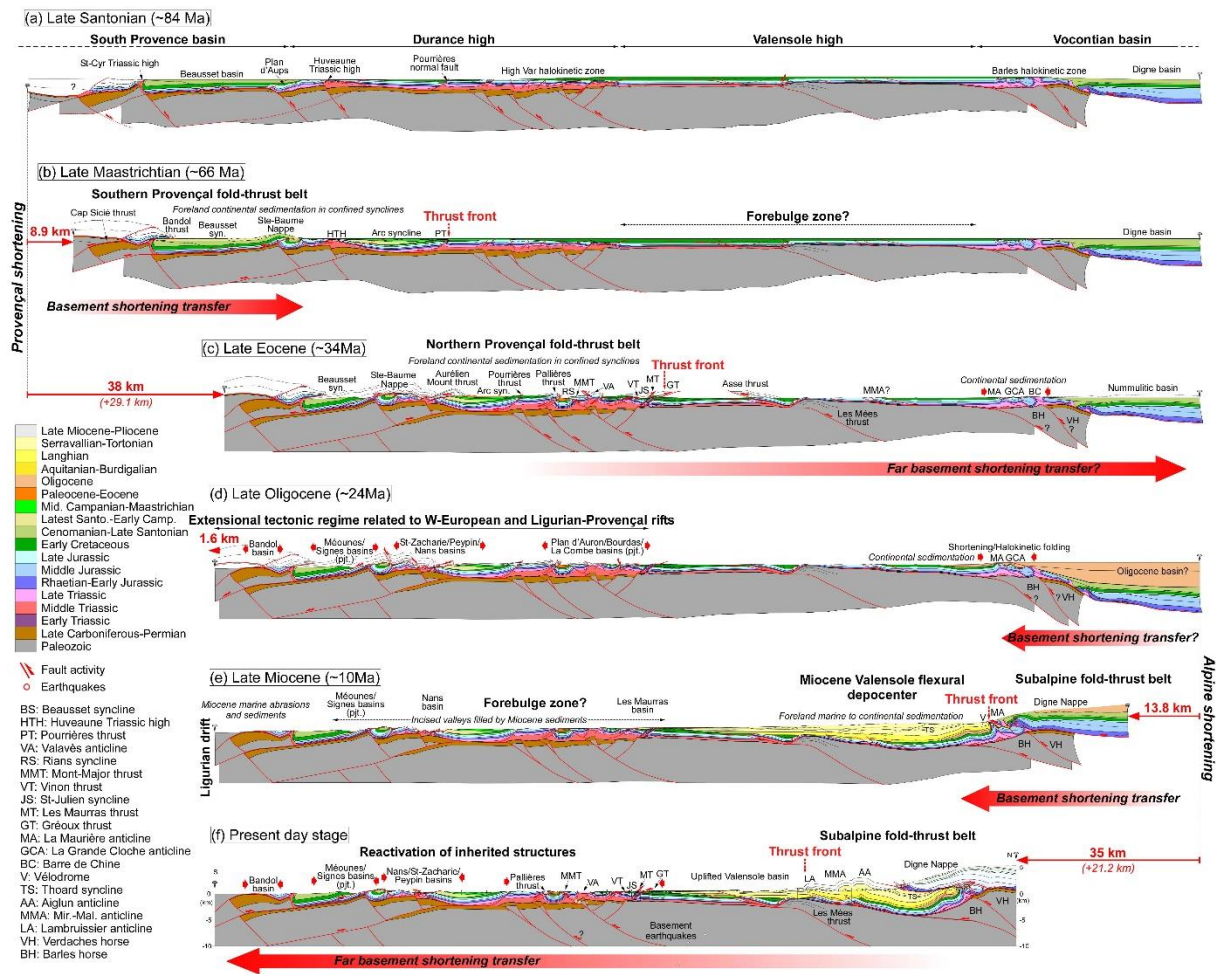


2496
 2497 **Fig. 18.** (a-b) Panoramic views of the Aiglun and Mirabeau-Mallemoisson fault-propagation
 2498 folds armed by upper Miocene conglomerates (Valensole 1 Formation) along the Duyes
 2499 valley (views looking westward; 44.066856° , 6.121088° ; 44.039937° , 6.103457°). For
 2500 location, see also Fig. 3. (c) Panoramic view of the southern Valensole foreland basin (view
 2501 looking NE-ward; 43.994556° , 5.906601°). Here, the basin is filled by upper Miocene-
 2502 Pliocene conglomerates (Valensole 1-2 Formations), topped by slightly S-dipping surface of
 2503 the Puimichel plateau. Northward, the sedimentary pile is deformed by the Lambruissier and
 2504 Mirabeau-Mallemoisson anticlines. Striated faults measured in conglomerate beds of the
 2505 southern limb of the Mirabeau-Mallemoisson anticline and close to the Les Mées town are

2506 consistent with a NE-trending compression. Fault data and bedding (dashed line) have been
 2507 projected in an equal area stereogram, lower hemisphere.



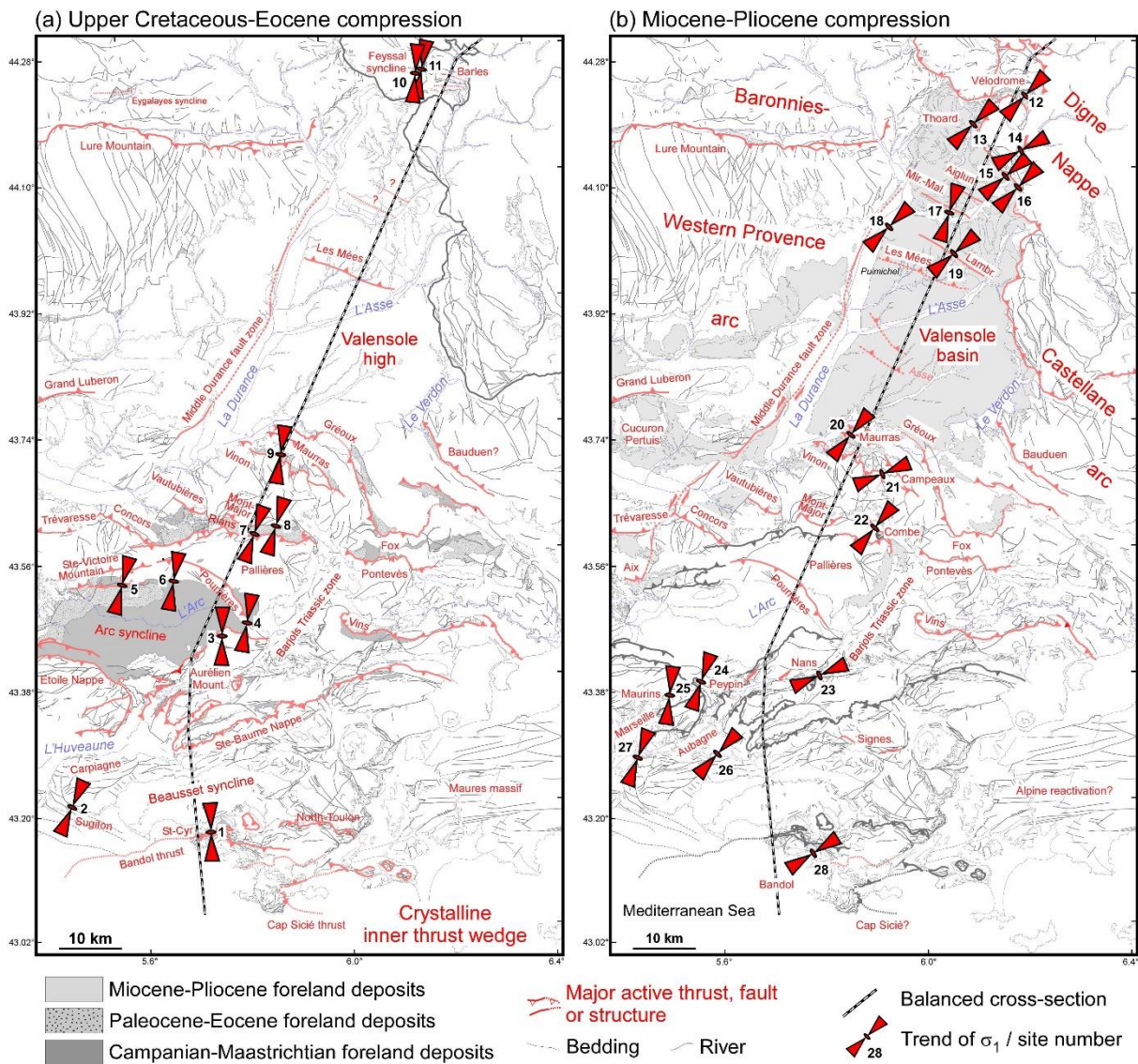
2508
 2509 **Fig. 19.** Structural interpretation of southern and central part of the Valensole foreland basin
 2510 using reprocessed seismic profiles VL85-P and 75DV-2 calibrated using the Les Mées-1 well.
 2511 For locations, see Fig. 3. The interpretation of the seismic profiles shows Provençal
 2512 compressional structures (Les Mées basement thrust and upper Asse thrusts) eroded and
 2513 sealed by Miocene strata. These thrusts have been slightly reactivation by the Alpine
 2514 compression.



2515

2516 **Fig. 20.** Tectonic evolution of the superimposed Provençal and southwestern Subalpine thrust
 2517 wedges. (a) Late Santonian restoration showing the structural architecture of the South
 2518 Provence basin (Beausset basin) and Vocontian basin (Digne basin) separated by the
 2519 Durance-Valensole high system. (b) Restoration of Provençal belt in Late Maastrichtian with
 2520 activation of the Bandol thrust and Sainte-Baume Nappe, and sedimentary infill of the Arc
 2521 depocenter. (c) Restoration of the Provençal belt in Late Eocene with northward transmission
 2522 of the shortening through the northern part of the Provençal zone, Valensole high, and up to
 2523 the Barles area. (d) Late Oligocene stage illustrating the deformation of the southern
 2524 Provençal wedges by extensional faulting and initiation of the Alpine shortening in the Barles
 2525 area to the north. (e) Growth of the Barles triangle zone during Miocene and concomitant
 2526 sedimentary infill of the Valensole foreland basin. (f) Late Miocene-Pliocene emplacement of

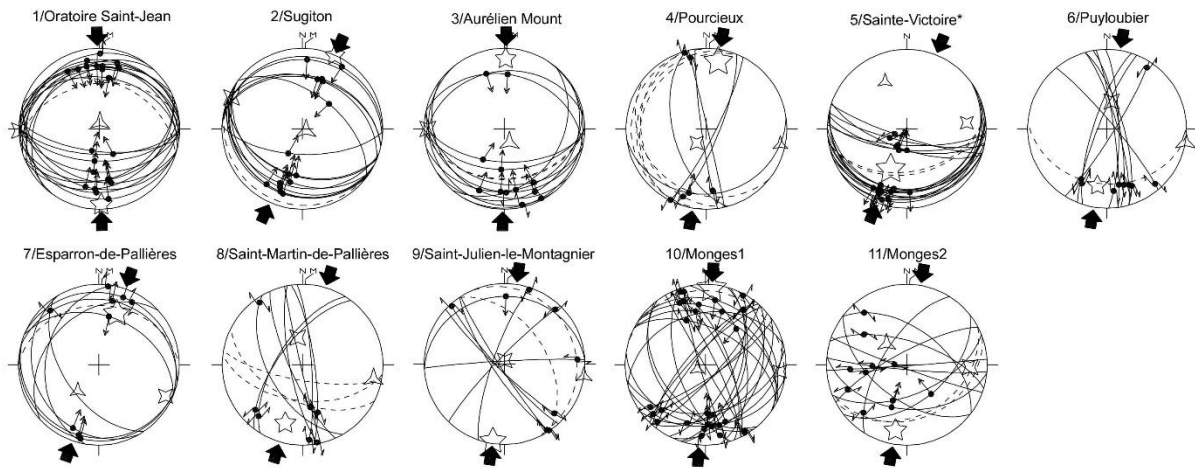
2527 the Digne Nappe and southward propagation of the Subalpine thrust front into the Valensole
 2528 foreland basin in cover and up to Mediterranean coast (Bandol basin) through the basement.



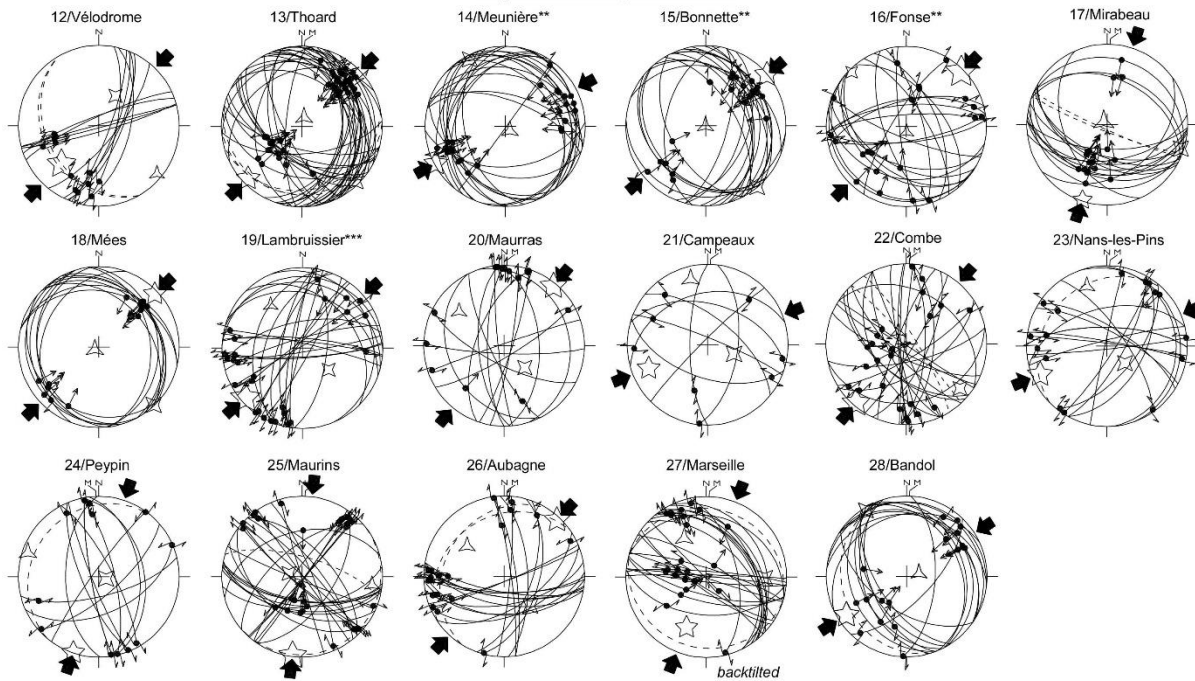
2529

2530 **Fig. 21.** Reconstruction of compression directions (trend of σ_1 is shown by red double arrow).
 2531 (a) Provençal compressional stress field. (b) Alpine compressional stress field. Site numbers,
 2532 fault slip data and computed principal stress axes are shown in Fig. 22 and Table 1. The
 2533 names of the major active structures are indicated in red. Thick dashed black and white line
 2534 shows the trace of the balanced and restored cross-section of Fig. 6.

Provençal compression



Alpine compression



2535

2536 **Fig. 22.** Fault data and principal computed stress axes (see Angelier (1990) for more details).

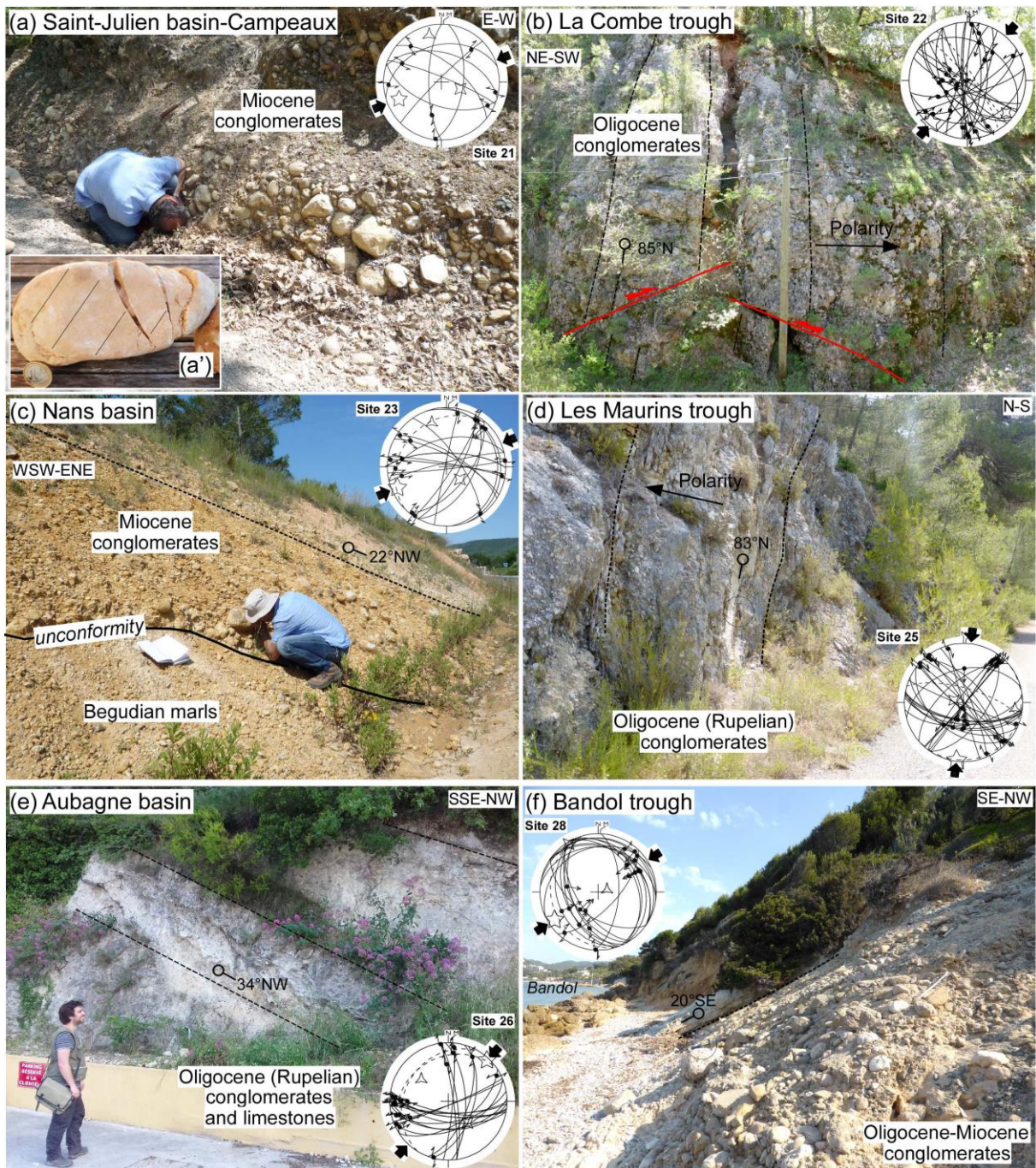
2537 Principal stress axes (σ_1 , σ_2 , and σ_3) are symbolized by five-branch, four-branch, and three-

2538 branch stars, respectively. Trend of σ_1 is shown by arrows. For locations, see Fig. 21. Data are

2539 projected in an equal area stereogram, lower hemisphere (see Table 1). Dashed line

2540 corresponds to the projection of bedding dip in the stereogram. *Espurt et al. (2012).

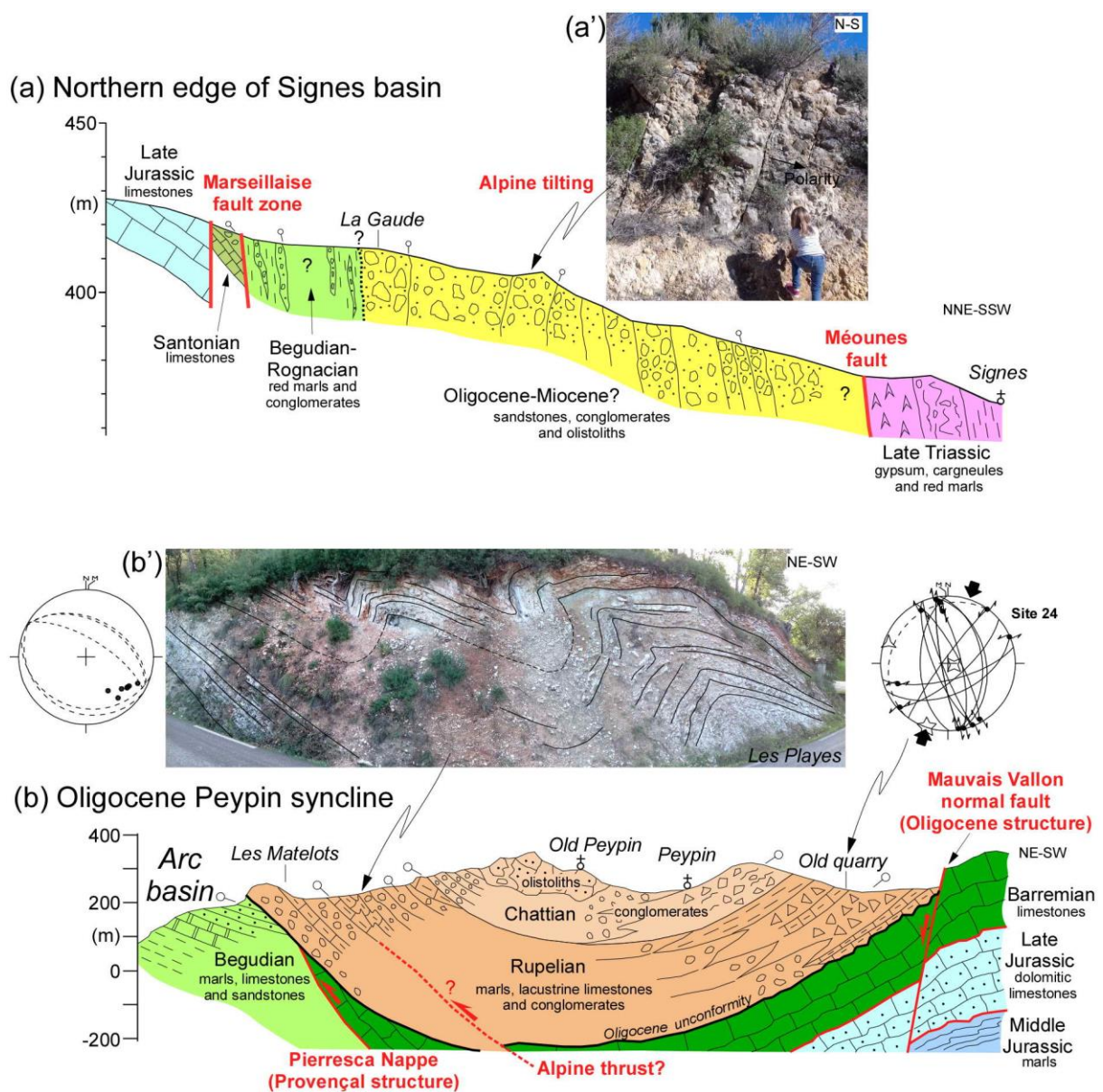
2541 **Hippolyte et al. (2012). ***Hippolyte and Dumont (2000).



2542

2543 **Fig. 23.** Field evidence for Alpine contractional deformations in the Provençal zone. For
 2544 location, see Fig. 2. (a) Striated and sheared Miocene conglomerates above the Vinon thrust,
 2545 eastern edge of the Saint-Julien basin-Campeaux (43.677722°, 5.946401°; Figs. 3 and 12a).
 2546 (a') Zoom on a sheared and striated (thin black lines) limestone pebble. (b) Overturned
 2547 Oligocene strata in the La Combe trough, eastern edge of the Rians basin (43.600228°,
 2548 5.927771°; modified from Espurt et al., 2019b). (c) Tilted Miocene strata in the Nans basin,

2549 southern edge of the Barjols Triassic zone (43.392411°, 5.798388°). (d) Tilted Oligocene
 2550 (Rupelian) conglomerate beds in the Les Maurins trough (43.363886°, 5.512575°). (e) NW-
 2551 dipping Oligocene (Rupelian) conglomerate and limestone beds in the southern edge of the
 2552 Aubagne basin (43.282596°, 5.597940°). (f) SE-dipping Oligocene-Miocene conglomerate
 2553 beds in the Bandol trough (43.133639°, 5.779645°). Striated faults in these Oligocene and
 2554 Miocene strata are consistent with a ~NE-trending Alpine compression. Fault data and
 2555 bedding (dashed line) have been projected in an equal area stereogram, lower hemisphere.



2556

2557 **Fig. 24.** Surficial cross-sections and field evidence showing Alpine contractional
2558 deformations in the southern part of the Provençal domain. For location, see Fig. 3. (a)
2559 Surface cross-section across the northern part of the Signes basin (43.295733°, 5.867607°).
2560 (a') Photograph of overturned beds of Oligocene-Miocene conglomerates is shown. (b)
2561 Surface cross-section of the Peypin syncline (43.381589°, 5.575817°). Photograph of SE-
2562 trending folds in Oligocene limestone and conglomerate beds on the northern limb of the
2563 Peypin syncline is shown (43.386835°, 5.603627°). Striated faults and fold axis in these
2564 Oligocene strata are consistent with a NE-trending Alpine compression. Fault data, fold axis
2565 (point) and bedding (dashed line) have been projected in equal area stereograms, lower
2566 hemisphere.

Site number/name	Latitude (m)	Longitude (m)	Elevation (masl)	Age/Lithology	Number of faults	σ_1		σ_2		σ_3		Φ	AN	RU	Stress regime
						Tre nd	Plun ge	Tre nd	Plun ge	Tre nd	Plun ge				
1/Oratoire Saint-Jean	43.166772	5.719908	139	Santonian/Limestones	22	179	6	269	1	5	84	0.4 48	10	30	reverse
2/Sugiton	43.217049	5.450640	208	Berriasian/Limestones	15	24	0	294	4	115	86	0.5 59	6	27	reverse
3/Aurélien Mount	43.455585	5.775748	631	Early Jurassic/Limestones	12	1	14	270	5	160	75	0.5 24	14	34	reverse
4/Pourcieux	43.475319	5.796727	351	Late Jurassic/Limestones	12	1	14	270	5	160	75	0.5 24	14	33	strike-slip
5/Sainte-Victoire*	43.524806	5.569858	334	Danian/Conglomerates	17	202	45	284	25	335	35	0.0 8	8	31	reverse
6/Puylobier	43.528522	5.675310	396	Early Campanian/Limestones	9	190	28	10	62	100	0	0.5 69	8	36	reverse
7/Esparron-de-Pallières	43.590707	5.838716	495	Danian/Conglomerates	9	19	33	115	9	219	55	0.2 07	9	39	strike-slip reverse
8/Saint-Martin-de-Pallières	43.601974	5.887519	400	Danian/Conglomerates	8	197	25	350	62	101	11	0.4 89	14	36	strike-slip
9/Saint-Julien-le-Montagnier	43.692550	5.905738	565	Late Jurassic/Limestones	8	189	3	6	87	99	0	0.2 38	10	32	strike-slip reverse
10/Monges1	44.242379	6.192825	1450	Late Eocene-Oligocene/Conglomerates	29	5	5	95	8	243	80	0.1 06	10	38	strike-slip reverse
11/Monges2	44.246639	6.201392	1535	Late Eocene-Oligocene/Conglomerates	11	190	18	92	22	316	60	0.4 18	17	44	strike-slip reverse
12/Vélodrome	44.203386	6.256325	1191	Late Miocene/Conglomerates	12	224	34	29	55	130	7	0.1 13	9	33	strike-slip
13/Thoard	44.171178	6.151978	785	Late Miocene/Conglomerates	38	226	8	135	5	14	81	0.3 56	8	24	reverse
14/Meunières**	44.123561	6.245412	866	Late Pliocene/Conglomerates	25	242	2	332	6	134	83	0.3 4	10	38	strike-slip reverse
15/Bonnette**	44.093349	6.221008	767	Late Pliocene/Conglomerates	21	50	0	140	0	249	90	0.2 9	10	32	strike-slip reverse
16/Fonse**	44.072696	6.249599	1050	Late Pliocene/Conglomerates	21	45	4	315	4	177	85	0.0 1	11	34	strike-slip reverse
17/Mirabeau	44.043538	6.105934	536	Late Miocene/Conglomerates	17	192	2	108	4	309	85	0.6 05	7	18	reverse
18/Mées	44.030339	5.980524	424	Late Miocene/Conglomerates	15	46	5	137	1	238	85	0.4 07	8	36	reverse
19/Lambruissier***	43.981107	6.105508	845	Late Miocene/Conglomerates	26	229	6	130	55	323	34	0.0 83	10	39	strike-slip reverse
20/Maurras	43.733025	5.891955	458	Miocene/Conglomerates	14	40	4	137	59	307	90	0.2 23	11	35	strike-slip reverse
21/Campeaux	43.677643	5.946504	522	Miocene/Conglomerates	8	248	22	109	62	345	17	0.3 93	15	37	strike-slip
22/Combe	43.600039	5.927321	337	Oligocene/Conglomerates	23	221	4	129	15	327	75	0.1 62	17	45	strike-slip reverse
23/Nans-les-Pins	43.392464	5.798273	369	Miocene/Conglomerates	8	251	18	54	71	159	5	0.4 34	11	36	strike-slip
23/Peypin	43.381959	5.573757	303	Oligocene/Conglomerates	12	199	1	105	82	289	8	0.3 89	11	29	strike-slip
25/Maurins	43.366172	5.506941	300	Oligocene/Conglomerates	27	187	2	283	76	97	14	0.1 7	11	30	strike-slip reverse
26/Aubagne	43.282650	5.597922	114	Oligocene/Conglomerates	19	43	4	138	49	309	40	0.2 88	13	30	strike-slip
27/Marseille	43.279588	5.444457	97	Oligocene/Conglomerates	23	201	26	82	44	311	35	0.3 24	13	32	strike-slip reverse

2567 **Table 1.** Paleo-stress tensors computed from fault slip data. See location of sites in Fig. 21.
 2568 Fault data and principal computed stress axes are shown in Fig. 22. σ_1 , σ_2 , σ_3 : maximum,
 2569 intermediary, and minimum principal stress axis, respectively. $\Phi=(\sigma_2-\sigma_3)/(\sigma_1-\sigma_3)$.
 2570 ANG=average angle between computed shear stress and observed slickenside lineation (in
 2571 degrees). RUP=quality estimator ($0\leq RUP\leq 200$) taking into account the relative magnitude of
 2572 the shear stress on fault planes (see Angelier (1990) for more details). *Espurt et al. (2012).
 2573 **Hippolyte et al. (2012). ***Hippolyte and Dumont (2000).