

1 **Untangling the Annot sand fairway: structure and stratigraphy**
2 **of the Eastern Champsaur Basin (Eocene-Oligocene), French**
3 **Alps.**

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17 **ABSTRACT**

18 Early foredeep successions can yield insight on tectonic processes operating
19 adjacent to and ahead of fledgling orogenic belts but are commonly deformed
20 by the same orogens. We develop a workflow towards stratigraphic
21 understanding of these deformed basins, applied to the Eastern Champsaur
22 Basin of the French Alps. This contains a down-system correlative of the
23 southern-sourced (Eocene-Oligocene) Annot turbidites. These strata are
24 deformed by arrays of W-facing folds that developed beneath the Embrunais-
25 Ubaye tectonic allochthon. The folds vary in geometry through the
26 stratigraphic multilayer. Total shortening in the basin is around 4 km and the
27 restored (un-decompacted) stratal thickness exceeds 980m. The turbidites
28 are generally sand-rich and bed-sets can be correlated through the entire fold
29 train. The succession shows onlap and differential thickening indicating
30 deposition across palaeobathymetry that evolved during active basement
31 deformation, before being over-ridden by the allochthon. The sand system
32 originally continued over what is now the Ecrins basement massif that, while
33 contributing to basin floor structure, only served to confine and potentially

34 focus further sediment transport to the North. Deformation ahead of the main
35 Alpine orogen appears to have continued progressively, and the past
36 definition of distinct “phases” (“pre-“ and “post-Nummulitic”) is an artefact of
37 the stratigraphic record.

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39 Keywords: turbidites, structural restoration, foreland basin,

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41 Turbidite systems can offer important constraints on bathymetric continuity of
42 arrays of sedimentary basins (e.g. Smith 2004; Puigdefabregas *et al.* 2004)
43 and thus inform models of basin evolution. For ancient foredeep systems the
44 challenge is to unravel useful stratigraphic information from successions that
45 are involved in orogenic belts and then deformed. If stratigraphic information
46 can be extracted from deformed systems, the insights so gained can then not
47 only inform models of orogenic evolution but also impact on greater tectono-
48 stratigraphic understanding of turbidite systems. This is especially important
49 as many of the key ideas of turbidite sedimentation come from studies of
50 successions in syn-orogenic sedimentary basins. The aims of this paper are
51 two-fold. The first is to document a workflow for developing stratigraphic
52 knowledge from highly folded successions. The second is to provide a linked
53 stratigraphic and structural reconstruction of the Eastern Champsaur Basin –
54 a deformed part of the Eocene-Oligocene Annot Sandstone system of SE
55 France (the Grès d’Annot in Alpine literature: e.g. Sinclair 1997; Joseph &
56 Lomas 2004) that has become incorporated into the Western Alpine orogen.
57 While an extensive re-examination of the Annot system in the light of our
58 reconstructions lies beyond the scope of this paper, we do explore some
59 implications for the extent of the system, how crustal-scale deformation
60 migrates through time and how tectonic evolution is represented in the
61 stratigraphic record.

62

63 **Geological context and motivation**

64

65 The Annot system of SE France as a whole occupies a pivotal position
66 in turbidite research, with outcrops providing test-beds for models of deep-
67 water sedimentation and analogues for hydrocarbon reservoirs in the

68 subsurface. Bouma's pioneering descriptions of turbidite facies come from
69 proximal parts of the Annot system (e.g. Bouma 1959; Stanley & Bouma
70 1964). This turbidite system was deposited in a broadly north-south trending
71 foredeep basin, adjacent to the ancestral Alps and fed from eroding granitic
72 basement to the south (Maures-Esterel massif, Fig. 1a) together with Corsica
73 that, at the time, lay adjacent to the south coast of mainland France. Much of
74 the early research is cited by Joseph & Lomas (2004) and papers that follow
75 this introduction greatly expand upon these works. More recent studies
76 include Mulder *et al.* (2010), Etienne *et al.* (2012), Salles *et al.* (2014), and
77 Cunha *et al.* (2017).

78 Existing work has almost entirely focussed on sites that have
79 experienced rather little deformation. The system is known to continue into
80 more strongly deformed regions to the north and northeast (e.g. Sinclair
81 1997). By increasing understanding of the Annot system, insight is gained on
82 the continuity and early deformation of the Alpine "foreland". As subaqueous
83 gravity flows, turbidity currents seek low bathymetry. Therefore, the continuity
84 of sand fairways, the geological record of the routes taken by their causative
85 flows, provides evidence for relative bathymetry in front of the evolving Alpine
86 orogen. The various relationships between the turbidites and their substrate
87 chart not only the syndepositional deformation of the foredeep basin floor but
88 also the deformation that predated basin subsidence. This understanding is
89 important for evaluating the tectonic significance of unconformities associated
90 with orogens and their implications for evaluating the progression of
91 continental deformation. By increasing understanding of the Annot system,
92 insight is gained on the continuity and early deformation of the Alpine
93 "foreland".

94 Our case study here focusses on the south-eastern flank of the
95 external Alpine basement massif of the Ecrins (Fig. 1a). The Eocene-
96 Oligocene litho-stratigraphy of the area, in common with the main Annot
97 system (Fig 2; e.g. Joseph & Lomas 2004), consists of three main lithofacies
98 (Debelmas *et al.* 1980) that comprise the "Nummulitic trilogy". Collectively this
99 represents the foredeep megasequence developed during this early stage of
100 Alpine orogenesis. The megasequence oversteps a substrate of crystalline
101 basement, continuous with the neighbouring Ecrins massif, together with

102 remnant patches of the Triassic-Jurassic cover to the massif (e.g. Debelmas
103 *et al.* 1980). The eastern and southern margins to the basin are now buried
104 beneath the tectonic allochthon of the Embrunais-Ubaye thrust sheets.

105 The first part of the Nummulitic trilogy is the Nummulitic Limestone, a
106 shallow-water transgressive unit here of Priabonian age (Debelmas *et al.*
107 1980; Dumont *et al.* 2012). Locally, the Nummulitic Limestone is underlain by
108 conglomerates derived from the adjacent crystalline basement, interpreted to
109 be the fills to subaerially-incised palaeovalleys (e.g. Gupta 1997). The second
110 part of the Nummulitic trilogy are the Blue Marls (“Marnes à Globigérines” on
111 published maps, e.g. Debelmas *et al.* 1980). These calcareous mudstones
112 chart palaeo-bathymetries increasing to several hundred metres depth. The
113 final part of the trilogy are turbidites – the Champsaur Sandstone (“Grès du
114 Champsaur” on published maps, e.g. Debelmas *et al.* 1980). Locally, there is
115 a facies intermediate between these turbidites and the Blue Marls, the Brown
116 Marls (e.g. Stanbrook & Clark 2004). These are generally inferred to have
117 accumulated on local intra-basin highs from dispersed suspension clouds
118 associated with the turbidity currents that deposited the Champsaur (and main
119 Annot) Sandstone.

120 The Nummulitic trilogy is capped by a shaley olistostrome termed the
121 “Schistes à Blocs” that lies directly under the Embrunais-Ubaye thrust sheets.
122 It is classically inferred to have been shed off the advancing thrust sheets as
123 the foredeep basin was closed (Kerckhove 1969). Detailed sedimentological
124 investigation of the Schistes à Blocs in the Eastern Champsaur Basin is
125 strongly inhibited by the intense penetrative deformation within this unit, which
126 presumably reflects distributed shear deformation in the footwall to the
127 overlying thrust sheets. Consequently, it is generally difficult to establish
128 whether the unit represents a series of amalgamated debris flow deposits or is
129 the product of a single submarine mass-wasting event.

130 Although given the single designator “Champsaur Sandstone”, early
131 Oligocene siliciclastic sandstones along the southern flank of the Ecrins
132 massif have two distinct compositions and, by inference, provenances, lying in
133 distinct depocentres (Debelmas *et al.* 1980). Following our previous usage
134 (Vinnels *et al.* 2010; Butler 2017), we term these depocentres the Western
135 and Eastern Champsaur Basins. They are separated by a zone of

136 deformation generally termed the Selle Fault (e.g. Tricart 1981; Ford 1996
137 and others since). To date, sedimentological research in the region has
138 mainly focussed on the westerly-derived, volcanoclastic turbidites that outcrop
139 in the Western Champsaur Basin, with detailed descriptions not only of
140 depositional architectures (e.g. Brunt *et al.* 2007) but also of the discrete fold-
141 thrust structures that have deformed the successions (Butler & McCaffrey
142 2004). Separated from this western basin by the Selle Fault Zone, the Eastern
143 Champsaur Basin is significantly more deformed than its western counterpart
144 (Fig. 3). It is characterised by folds on various wavelengths and amplitudes
145 that generally have tight hinges and straight limbs, approximating to chevron
146 geometry. This style of deformation is typical of successions with a
147 pronounced planar mechanical anisotropy such as is provided by inter-
148 bedded sandstones and shales (Ramsay 1974). The folds face westwards
149 with axial surfaces inclined moderately to the east. The structures were
150 described by Kerckhove *et al.* (1978) and studied in detail by Bürgisser & Ford
151 (1998). These studies show that the folds developed beneath, and locally
152 incorporate, the Embrunais Ubaye thrust sheets which overran both the
153 Eastern and Western Champsaur Basins. The thrust sheets provided tectonic
154 burial sufficient for peak temperatures in Eastern Champsaur of c 300 °C
155 (Bellanger *et al.* 2015).

156 Provenance of the Eastern Champsaur Sandstone has been disputed.
157 Sinclair (1997) depicts their turbidites to be the down-system continuation of
158 the Annot system otherwise preserved in the SW Alps of Provence. However,
159 in their regional overview, Joseph & Lomas (2004) indicate a northern
160 provenance, sourced from the an elevated, proto-Ecrins massif. The
161 argument is resolved by Vinnels *et al.* (2010) who present palaeocurrent data
162 to show a southerly derivation and document that the sandstones of the
163 Eastern Champsaur basin have the same composition as the Annot
164 Sandstone in its type area to the south. Thus Sinclair's (1997) view is
165 supported. Vinnels *et al.* (2010) note that, although deflected by subtle basin-
166 floor topography, the causative turbidity currents continued over what is now
167 the uplifted Ecrins basement massif. However, Vinnels *et al.* only focused on
168 the lower stratigraphic intervals, establishing facies variations and palaeoflow
169 deflections along the basal onlap surface of turbidites onto the Blue Marls that

170 drape uplifted substrate within the basin. The current paper focusses
171 exclusively on the Eastern Champsaur Basin, building on our previous work
172 (Vinnels *et al.* 2010). The regional structural setting is discussed extensively
173 by Butler 2017). The stratigraphic relationships between the eastern and
174 western Champsaur Basins and the role of the Selle Fault in partitioning
175 distinct deep-water systems lie beyond the scope of the current paper and are
176 reserved for later contributions.

177

178 **A workflow**

179

180 Geologists working in mountain belts have long had to grapple with the
181 complexities of deformation when attempting to build stratigraphic knowledge.
182 However, the methods used for creating this understanding are rarely
183 documented, making it difficult to assess uncertainties in the larger-scale
184 basin reconstructions. Furthermore, there has been a reluctance for many
185 researchers investigating sedimentary basins associated with mountain belts
186 to examine their more-deformed components. In presenting a workflow here
187 we hope not only to promote research within these settings but also to
188 encourage clearer documentation of how deformed stratigraphic sections are
189 reconstructed.

190 The Champsaur area exhibits just under 2 km of topographic relief with
191 cliff sections that are of limited accessibility, attributes that are typical of much
192 of the Alpine mountain ranges and of young orogenic belts in general. A
193 virtue of the relief in our Champsaur study area is that major hillsides trend
194 east-west, perpendicular to the trend of fold axes. They therefore provide
195 ideal cross-sections, akin to in-lines in grids of seismic reflection data.
196 Accessible routes up these hillsides provide excellent vertical sections,
197 equivalent to boreholes for subsurface examples. Consequently, we apply a
198 workflow similar to that adopted for surface stratigraphic-structural mapping,
199 tying detailed stratigraphic observations from vertical sections (boreholes) to
200 remotely sensed cross-sections (in this case by direct observation from
201 opposite valley-sides, equivalent to seismic profiles).

202 The first decision lies in selecting the best vertical section from which to
203 build a stratigraphic succession. The west face of Le Piquet provides a strike

204 section, 3.5 km across, that runs from up from the base-Nummulitic
205 unconformity to a small klippe of Embrunais-Ubaye thrust sheets (Fig. 5a).
206 Only the upper part of this hillside provides continuous outcrop that is also
207 amenable to direct observation – and this yields stratigraphic logs (Fig. 5a).
208 Logged units include a c 380m interval that is sandstone-dominated (labelled
209 A6 on Fig. 5a), capped by Schistes à Blocs. However, the broad bed-set
210 characteristics of the underlying succession may be established by remote
211 observation. These include shale-dominated levels that are more recessive on
212 hillsides (A3 and A5 on Fig. 5a).

213 The boundaries between the bed-set units can be traced through the
214 hillsides on Le Piquet and across the Dourmillouse Valley onto the dip section
215 provided by its northern slopes (Fig. 5b). The upper part of the valley side,
216 including the ridge line, provides near continuous, if largely inaccessible,
217 outcrop within which the main folds can be identified. The fold axial traces can
218 be mapped down dip into the Dourmillouse Valley. Then the bed-set
219 stratigraphic units of the Champsaur sandstone can be traced through the fold
220 train. This section is moderately accessible and locally provides accessible
221 sections for detailed stratigraphic logging.

222 The axial traces of the folds provide the key linking structures to tie
223 adjacent transects. These can be traced over the ridge line bounding the
224 Dourmillouse Valley to the north and into the adjacent Fornel Valley (Fig. 4).
225 The northern side of the ridge-line is much steeper than its southern
226 counterpart so that accessibility is severely restricted. However, fold axial
227 traces may be readily traced through the cliffs and the bed-set stratigraphy
228 defined and similarly tracked through the fold array (Fig. 5c). Note that there is
229 some polyharmonic folding but the axial surfaces for the main fold closures
230 may be traced from the ridge-line down to the Blue Marls.

231

232 **The bed-set stratigraphy**

233

234 Following the workflow laid out above it is possible to map out not only
235 the axial traces of the principal folds but also the bed-set stratigraphic units
236 within the Champsaur sandstone (Fig. 4). Although structural deformation
237 hinders identification of individual beds across the whole basin, thicker

238 intervals can be laterally traced continuously for several kilometres and
239 probably extend much further. Vertical stratigraphic sections display
240 significant heterogeneity in sandstone to shale ratio. These are well-illustrated
241 by the cliff sections of Le Piquet (Fig. 5a),

242 As noted above, it is the variety in sandstone-shale abundance that
243 provides a framework, on the bed-set scale, for dividing the basin stratigraphy
244 into mappable units (Fig. 5). These units are illustrated on arrays of composite
245 logs (Fig. 6). Principally, our sedimentary logs were measured from the fold
246 limbs to avoid hinge areas, where beds show significant layer parallel
247 shortening strain manifested by cleavage, and tracts of strongly overturned
248 strata where bed thinning is likely. Strains in both of these structural settings
249 are localised in the shales – sandstone beds retain unmodified sedimentary
250 structures suggesting penetrative strains in these units are very low. We
251 estimate an uncertainty of less than 10% on the normally-compacted
252 thickness of sandstones and perhaps up to 20% for the shales on the values
253 reported in Fig. 6, as we cannot rule out deformation even in the normal limbs
254 of folds. The bed-set sequences may still be identified regardless of strain
255 variations and we now use these to erect a basin-wide lithostratigraphy.

256 The Champsaur Sandstone onlaps its substrate towards the western
257 basin margin (Vinnels *et al.* 2010 and references therein): the oldest
258 preserved parts of the sandstone sequence are only present on the vegetated
259 lower slopes of the far eastern exposures of the basin. In the Narreyroux
260 valley (log location 4 on Fig. 4), a 670m thick section of Champsaur
261 Sandstone (A4 on Fig. 6) is underlain to the east by considerably more
262 stratigraphy but precise correlation has proven elusive. The higher parts of
263 the stratigraphy are more tractable. Two levels are especially sandstone-poor
264 and form key markers (A3 and A5 in Fig. 6). The lower of these two shaley
265 intervals is approximately 50m thick and contains a distinctive couplet of 2 m
266 thick sandstone beds that can be found in all sections that include this
267 stratigraphic interval. There are a few thin sandstone beds present in the
268 higher shale-dominated interval (A5 in Fig. 4) but these are laterally
269 discontinuous. Between the two shaley sections lies an interval with
270 abundant thick sandstone beds (A4 in Fig 4). The section below the lower
271 shaley interval also has higher sandstone content; its upper portion contains

272 beds that are 3-5 m thick while the lower portion is bedded on the 50 cm – 1
273 m scale. These distinctions provide confirmations of stratigraphic correlations
274 to elsewhere within the Eastern Champsaur district.

275 The lower shaley interval (A3) can be readily correlated between the
276 Narreyroux, Fournel and Dourmillouse valleys (Fig. 4) to parts of the study
277 area where more complete sections of the underlying sandstone-rich intervals
278 are exposed. In these places we can separate a thicker-bedded upper
279 component from a thinner-bedded lower component (A2 and A1 respectively
280 in Fig. 6). The thicker-bedded interval (A2) is about 40 m thick. The logs from
281 the Fournel and Dourmillouse valleys (Fig. 6) only demonstrate about 60 m of
282 this lower portion. However, in the Narreyroux valley it achieves a thickness in
283 excess of 150 m with a further 200 m represented by the poorly exposed and
284 deformed terrain deeper in the section.

285 The upper shaley horizon defined in the Narreyroux valley (A5 on Fig.
286 6) can be correlated across the Dourmillouse valley (Fig. 4) to the cliffs of Le
287 Piquet. This is the “Thick Shale” of Vinnels *et al.* (2010). Overlying this, 380
288 m of sandstone-rich turbidies (A6 on Fig. 4) are capped by the “Schistes a’
289 Blocs” that directly underlie the Embrunais-Ubaye thrust sheets. The
290 sandstone rich interval (A6) can be traced through the fold structures of the
291 upper Dourmillouse and Fournel Valleys, and constitute the upper 60m of the
292 logged section in the Narreyroux valley (Fig. 4). In the upper Fournel Valley
293 the thickness of this interval exceeds 400 m (log 5 on Fig. 6). Individual beds
294 may be traced through folds and across the ridge line between the upper
295 parts of the Dourmillouse and Fournel valleys. For example, a prominent
296 shale interval (5-15 m thick; X on logs 5 and 6 on Fig. 6), at the base of
297 interval A6b) can be traced between sections.

298 The Eastern Champsaur Basin continues over the watershed at the
299 head of the Dourmillouse Valley into drainage of the Champoleon Valley (Fig.
300 4). Individual beds may be traced along the ridge line (Pointe des Estaris -
301 Pointe des Pisses: log locations 8 and 9, Fig. 4) to complete the upper
302 stratigraphy of this part of the Champsaur Sandstone up to the Schistes à
303 Blocs (Fig. 6).

304 By combining the logs, we estimate that about 950m of Champsaur
305 Sandstone stratigraphy (A6 – A7) overlies the prominent upper shaley horizon

306 (A5) in the ground between the upper Dourmillouse valley and Pointe des
307 Pisses (Fig. 4). The equivalent interval at Le Piquet is just 380m thick (log 1
308 on Fig. 6). Both sections are capped by the Schistes à Blocs so the thickness
309 difference is not due to post-depositional tectonic truncation. The difference is
310 also too great to be due to variations in distributed deformation or differential
311 compaction and there are significant differences in the bed-set stratigraphy
312 between these locations. We interpret the thickness variation to result from
313 differential stratigraphic growth across the Eastern Champsaur Basin.

314

315 Depositional architecture and growth

316 Although folding permits construction of substantial stratigraphic
317 sections and deformation has not been sufficient to prevent individual bed-
318 tracing through the high Alpine landscape, it does inhibit the recognition of
319 elements of depositional architecture that might be expressed at the km-scale
320 (such as the development of compensating turbidite lobes or, depending on
321 the scale, of erosional or constructional channels). In the Western Champsaur
322 basin, adjacent to our study area (but forming a distinct turbidite system),
323 extensive deep incision and former submarine canyons have been described
324 (e.g. Brunt *et al.* 2007). However, there is no evidence of incision within the
325 Eastern Champsaur Basin, beyond local scours (and associated layers of
326 mudstone rip-up clasts) and bed-scale amalgamation, nor of channel-levee
327 development. The 4-5 km strike section on the western face of Le Piquet
328 (located on Fig. 4, illustrated on Fig. 5a) reveals very low-angle discordances
329 within parts of the upper sandstone interval (A6) consistent with an expansion
330 of stratigraphic thickness from north to south (Vinnels *et al.* 2010). Individual
331 beds appear to thin northwards while bed-sets abut onto underlying beds –
332 relationships we interpret as internal onlap. Collectively these relationships
333 are plausibly interpreted as representing syn-depositional tilting within the
334 Eastern Champsaur Basin elevating Le Piquet area relative to the Pointe des
335 Pisses area during deposition of these upper sandstone intervals. Vinnels *et*
336 *al.* (2010) suggest that these thickness changes reflect differential growth of
337 an anticline in the basin floor. We infer that the intrabasinal slopes during
338 deposition of at least this part of the Champsaur Sandstone were relatively
339 gentle, there being no evidence for significant mass-wasting within the

340 succession. Therefore, evolving basin floor structures were largely swamped
341 by concurrent sedimentation.

342

343 **The post-depositional structure of the basin**

344

345 The stratigraphic template of mappable bed-sets permits full correlation of
346 units through the fold systems. These correlations are illustrated on our
347 summary map (Fig. 4) and underpin two cross-sections through the Eastern
348 Champsaur Basin (Fig. 7). The sections are necessarily simplified at the scale
349 reproduced here, and no attempt is made here to illustrate detail of the
350 thickness changes in the younger units (A6-A7), nor the precise geometry of
351 polyharmonic folding deeper in the stratigraphic pile (A1).

352 The two cross-sections (Fig. 7) show a similar structural style: a train of
353 asymmetric, west-facing folds with axial surfaces dipping towards the east.
354 The folds detach downwards into the Blue Marls (Bürgisser & Ford 1998).
355 Thrust faults are rare, show bed offsets of a few metres and are restricted to
356 fold hinge zones. The attitude of the fold axial surfaces and fold interlimb
357 angles vary through the region and, according to Bürgisser & Ford (1998),
358 these variations reflect heterogeneous simple shear strains distributed into the
359 footwall of the Embrunais-Ubaye thrust sheets. Folds occur in clusters,
360 separated by segments where stratigraphic sections are only weakly
361 deformed. Within the clusters the folds have wavelengths of 200-500m and
362 amplitudes of around 200m. The larger folds are broadly harmonic, with axial
363 surfaces that can be traced through the visible succession of the Champsaur
364 Sandstone. Prime examples are those folds that cross the ridge-line on the
365 north side of the Dourmillouse valley near the summit of Pic Felix Neff (Figs
366 5b,c, 7). The distribution of these larger folds is consistent with a model that
367 considers folds as ductile equivalents to imbricate thrusts (e.g. Pfiffner 1985;
368 Butler 1992). Qualitatively, the general form of the folds appears to be
369 influenced, at least in part, by variations in bed-thickness in the Champsaur
370 Sandstone. The thicker, sandstone rich sequences (e.g. Fig. 2a; A6 on Figs 4
371 and 5) appear to act as “control units”, in the sense of Price & Cosgrove
372 (1990), determining the wavelength of the main fold sets (Fig. 8). The older,
373 thinner-bedded units (Fig. 2; A2-A4 on Figs 4 and 5) deform poly-harmonically

374 with respect to the younger, thick-bedded stratigraphy (e.g. unit A6), and are
375 characterised by folding at short-wavelength. Together with this layer-
376 dependence, the location of folds in the Champsaur Sandstone may relate to
377 variations in the properties of the basal detachment zone in the Blue Marls, as
378 is well-known in other detachment systems (e.g. Cotton & Koyi 2000 and
379 many others). Less effective detachment would promote folding in the
380 overlying strata. These aspects of the deformation await more detailed
381 investigations.

382 The cross-sections (Fig. 7) demonstrate the regional discordance
383 between the lithostratigraphic units in the Champsaur Sandstone and the Blue
384 Marls. As noted by Vinnels *et al.* (2010, but well-known informally), this
385 represents a regional onlap surface, albeit modified by the tectonic
386 detachment. It charts deformation of the basin floor during the accumulation of
387 the Nummulitic trilogy. Vinnels *et al.* (2010) treat the onlap angle qualitatively.
388 To consider the angle of this onlap in profile, and to establish an estimate of
389 tectonic shortening in the basin, we present a balanced and restored section
390 (Fig. 8) for the Dourmillouse transect (Fig. 7b).

391 The restoration assumes that the sand-rich parts of the succession
392 deformed largely by concentric folding, without appreciable changes in bed
393 thickness by distortional strain. To track this assumption, we chart the
394 thicknesses of bed-sets through the cross-section. These thicknesses are
395 reported on both the final state and restored cross-sections (thin bed-
396 perpendicular rulings). With the assumption of concentric folding, restorations
397 can be performed simply by restoring the length of strata, conserving the
398 sinuous lengths measured on the final state section. This is line-length
399 restoration, as proposed by Dahlstrom (1969). Note that in this analysis we do
400 not consider the impact of broad arching of the enveloping surfaces of the
401 folds with in the Champsaur Sandstone. This longer-wavelength deformation
402 couples with the underlying basement and it is the architecture of the basin fill
403 that we wish to reconstruct.

404 The graphical restoration (Fig. 8b) is hung from the younger
405 stratigraphic levels (A6), the top of which forms a horizontal datum, or “target
406 horizon”, for determining the progress of deformation. This interval, together
407 with those below, retain bed-length and bed-thickness from the deformed

408 state. A complete restoration might be expected to yield strata that are planar
409 and parallel – features that are not evident here. There are various possible
410 explanations for these discrepancies. First, distortional strains, especially
411 within the mudstone-rich intervals, have not been considered in our method,
412 which assumes ideally concentric folding. Certainly, these units are cleaved,
413 implying that they have experienced such distortions. There may also be
414 errors in the drafting of the cross-section, unavoidable given variations in
415 outcrop quality and uncertainties in the precise position of stratigraphic
416 boundaries. Errors arising from these effects amplify away from the “target
417 horizon”, creating significant short-wavelength relief on the base of section.
418 These are regarded here to be restoration artefacts (selected examples
419 identified in in Fig. 8b), as the top-basement surface is readily mapped and
420 shows no such abrupt changes in relief where observed in the field.

421 Notwithstanding the short-wavelength artefacts, the restoration (Fig.
422 8b) can be used to calculate the angular relationship between the depositional
423 units in the Champsaur Sandstone and the underlying Blue Marls. The
424 restoration implies onlap of 980m of Champsaur Sandstone over 9.4 km –
425 implying that the average angle of the onlap slope was 6 degrees. Note that
426 this angle would have been significantly higher at the time of deposition as we
427 have not allowed for vertical compaction during burial.

428 Tectonic shortening for the upper stratal levels (A6) is the difference in
429 length between deformed and undeformed state, 4 km. However, the
430 mismatches in restored lengths of the underlying strata indicate that the
431 section is not balanced (the bed-lengths and implied longitudinal strain are not
432 equal in layers, so that the restoration is not complete). We resolve this by
433 inferring significant “top-to-the-west” shear distributed through the Champsaur
434 Sandstone, as invoked by Bürgisser & Ford (1998). Variations in layer-parallel
435 shortening can account for the open folding of underlying sandstone units and
436 in the substrate.

437

438 **Discussion – implications for tectonostratigraphic evolution**

439

440 A palaeogeographic sketch is provided (Fig. 9), extending that of Joseph &
441 Lomas 2004, their Fig. 5) which integrates our findings from the Eastern

442 Champsaur Basin into a semi-regional context. Following our earlier work
443 (Vinnels *et al.* 2010), we consider its turbiditic basin fill to be a down-system
444 continuation of the Annot sand fairway. We show the main Annot turbidite
445 system to be derived from a southern source area and then routed along the
446 sinuous Provencal basins defined by the underlying deformation within the
447 broad foredeep basin (e.g. Salles *et al.* 2012). This system is distinct from a
448 secondary turbidite system that feeds the Western Champsaur Basin.

449

450 Down-system continuity of the Annot Sand Fairway

451 The total (compacted) stratal thickness of the Champsaur Sandstone in
452 the Western Champsaur Basin approaches 1km. Stratal patterns together
453 with onlap at the base and within the Champsaur Sandstone indicate active
454 deformation of the underlying basement during deposition. We propose that it
455 is this deformation that provided the basin floor structure that served to
456 confine and guide turbidity currents, facilitating sediment transport further
457 down-system. The high proportion of coarse sandstone compared to finer-
458 grained fractions, coupled with palaeocurrent data (Vinnels *et al.* 2010)
459 implies that there was significant bypass through the Eastern Champsaur
460 Basin. The Annot sand fairway therefore continued north from Champsaur,
461 with no ponding behind the ancestral Ecrins massif (cf. Joseph & Lomas
462 2004; Fig. 9); turbidity currents are inferred here to have transited the region
463 and fed down-system depocentres (Fig. 9). Certainly, the area of the Ecrins
464 massif did not form a bathymetric high during deposition of the Annot system
465 (cf. Ford *et al.*, 1999) – it must have been deeper than the upstream Annot
466 fairway. Future studies should address the sedimentology of the likely
467 correlatives of the Annot turbidites further down-system, in the strongly
468 deformed sections along the eastern part of the Ecrins and north in the
469 Aiguilles d'Arves (Fig. 1).

470

471 Eastern confinement of the Annot system

472 For Joseph & Lomas (2004, their fig.5) and references therein) the
473 eastern edge of the Annot sand system was confined by the Embrunais-
474 Ubaye thrust sheets, an exotic tectonic allochthon emplaced from the internal
475 Alpine domain onto the foreland domain (see also Ford *et al.* 2006). The

476 sheets were sufficiently thick to have buried the southern Ecrins district so
477 that it reached temperatures of c 300°C (Bellanger *et al.* 2015). Presumably
478 they created significant relief within the foredeep. Olistostromes were shed
479 from this relief during thrust sheet emplacement, manifest as the Schistes à
480 Blocs (e.g. Kerckhove 1969) which cap the turbidites in the Eastern
481 Champsaur Basin and elsewhere in the Annot system. Ford & Lickorish
482 (2004) suggest that it was the emplacement of the Embrunais-Ubaye thrust
483 sheets that closed off the Annot basin system. The implication is that this was
484 a progressive process, with the advance of thrust sheets gradually restricting
485 the pathways available for turbidity currents. However, the relationship
486 between the olistostomes and the Champsaur Sandstone is not consistent
487 with this implication. Muddy debris flows, such as carried the Schistes à
488 Blocs, would be expected to have had significant run-out distances, potentially
489 forming obstructions within the main flow paths for the Annot turbidity
490 currents. Deposits accumulating from axial flows in the case of the turbidity
491 currents, and from coeval basin flank-derived olistostromes, should be
492 interleaved. No such interleaving has been recognised by us in the Eastern
493 Champsaur Basin, nor elsewhere in the Annot System. Therefore, we deduce
494 that the olistostromes and Annot turbidites are not coeval – rather that the
495 Schistes à Blocs forms a distinct, younger depositional unit that overran (and
496 entrained) dark-shales. The corollary is that the flux of Annot turbidity currents
497 had terminated for reasons unrelated to the emplacement of the Embrunais-
498 Ubaye thrust sheets and that some other feature must have provided the
499 lateral (eastern) confinement to Annot flows.

500 If lateral confinement was not provided by the tectonic allochthon of the
501 Embrunais-Ubaye thrust sheets, then presumably it was provided by weakly
502 inverted parts of the ancestral rifted margin of Europe adjacent to the suture
503 of the closed Ligurian Tethys seaway. These may have formed additional
504 basement ridges, equivalent to the Ecrins but now buried beneath Alpine
505 thrust sheets. Alternatively, the Annot system may have abutted against
506 deformed Briançonnais and Sub-Briançonnais units, as illustrated on Fig. 9.
507 Both units contain sedimentary successions that are broadly time-equivalent
508 to the Annot turbidites – the so-called “Flysch Noir”. This is a succession of
509 pelites and thin, fine sandstones of mid-late Eocene age (e.g. Debelmas

510 1989). We tentatively suggest that the Flysch Noir represents an early part of
511 the Annot turbidite system and that its pathway was elevated as its substrate
512 of Brianconnais and Sub-Brianconnais units became deformed. Note that, by
513 the early Oligocene, the eastern edge of the Brianconnais in this sector of the
514 Alps had been incorporated into the Alpine orogen and was experiencing blue
515 schist metamorphism, and maybe even greater burial (e.g. Michard *et al.*
516 2004; Dumont *et al.* 2012). Regardless of the nature of the eastern retaining
517 margin to the Annot System, there is no documented evidence that it provided
518 detritus into the sand fairway, until the emplacement olistostromes of the
519 Schists à Blocs from the advancing Embrunais-Ubaye thrust sheets.

520

521 Structuring of the Eastern Champsaur Basin and the evolution of crustal 522 shortening

523 The onlap of Champsaur Sandstone onto a tilted succession of Blue
524 Marls and Nummulitic Limestone indicate that the floor to the Eastern
525 Champsaur basin was actively deforming during deposition (Vinnels *et al.*
526 2010). Our studies confirm our earlier work and extend it – stratigraphic
527 thickness variations across the Eastern Champsaur Basin (Fig. 6) strongly
528 suggest that open folding continued to influence sand accumulation. However,
529 sedimentation generally swamped the folds, with turbidites overstepping the
530 anticline crests. These folds are cored by crystalline basement which outcrops
531 in the Dourmillouse and Fournel valleys (Fig. 4) and it is this deep-rooting
532 deformation style that is illustrated on Fig. 9. Basement-rooting deformation is
533 also shown to provide the structural barrier that separated the distinct turbidite
534 systems that characterise the Eastern and Western Champsaur Basins. Folds
535 and thrusts in the adjacent crystalline basement now forming the Ecrins
536 massif (located on Fig. 1a) are generally interpreted as resulting from tectonic
537 inversion of half-graben inherited from Jurassic-aged continental rifting (e.g.
538 Gillcrist *et al.* 1987; Dumont *et al.* 2008).

539 In Provencal basins (Fig. 9), active deformation during deposition of the
540 Annot turbidites has been characterised as “wedge-top” (Salles *et al.* 2012
541 and references therein). The term harks back to idealised conceptualisations
542 of orogenic margins that are divided into a foreland basin and an adjacent
543 thin-skinned thrust system (e.g. De Celles & Giles 1996). In these models,

544 deposition can occur in the yet-to-deform foreland basin and in small basins
545 perched on the advancing thrust system (or wedge). However, such
546 distinctions appear inappropriate to us for the Champsaur district as
547 deformation of the basin floor during the turbidite deposition was thick-
548 skinned. Concepts of thrust wedges are unlikely to be directly relevant to
549 inversion tectonic systems such as the Ecrins or the initial deformation of the
550 Briançonnais as illustrated on Fig. 9.

551 In Provence, two distinct phases of folding are recognised – one
552 inferred to result from north-south contraction and traditionally related to
553 “Pyrenean-Provençal” orogeny (early Tertiary), and a subsequent SW-
554 directed thrust system (e.g. Siddans 1979; de Graciansky *et al.* 2011).
555 Likewise, deformation in the crystalline basement and the structural evolution
556 of the Ecrins basement massif has long been considered result from
557 punctuated tectonic activity - i.e., two distinct episodes of Alpine deformation,
558 separated by the unconformity that underlies the transgressive Nummulitic
559 Limestone (e.g. Gidon 1979). Ford (1996) correlates these with the two
560 tectonic episodes of Provence. Gupta & Allen (2000) argue that the southern
561 Ecrins was structured by folds and uplifted fault blocks to modulate the
562 Nummulitic transgression. It is a deduction that is consistent with the
563 punctuated model for tectonics in this part of the Alps. However, it seems
564 unlikely that initiation of a second episode of tectonic activity simply coincided
565 with the transgression of the Nummulitic megasequence across the
566 basement.

567 Recent radiometric dating of thrust zones in the Ecrins massif
568 (Bellanger *et al.* 2015) implies that deformation in the basement straddled
569 deposition of the post-Nummulitic successions of the Eastern Champsaur
570 Basin. Using these data, Butler (2017) proposes that crustal shortening in the
571 Ecrins and, by inference, beneath the Eastern Champsaur Basin was
572 continuous in time. Where the instantaneous syn-orogenic surface is sub-
573 areal, deformation is accompanied by erosion, marked by exhumation and
574 cooling of the basement. Where the instantaneous syn-orogenic surface
575 moves to below sea-level, deformation is accompanied by deposition. The
576 transition between these states is marked by transgression and its
577 diachroneity reflected by the sub-Nummulitic unconformity.

578 Dividing deformation episodes on the basis of an unconformity has a
579 long tradition in orogenic geology (see discussion in Gray *et al.* 1997).
580 However, is the interplay between uplift due to crustal shortening, with
581 concomitant erosion, and subsidence due to long-wavelength orogenic
582 loading of the lithosphere, with concomitant deposition, that defines the
583 stratigraphic record of orogenesis. Focussing on unconformities, with their
584 implications for missing time-sections, can create the false impression that
585 deformation was punctuated. It is our proposal that the crust beneath the
586 Champsaur district experienced deformation that was essentially continuous
587 over geological time, through the late Eocene and into the Oligocene.

588 The arrival of the Embrunais-Ubaye Thrust Sheets in the late Rupelian
589 (Dumont *et al.* 2008) marked a transition in the tectonic style to thin-skinned
590 shearing. Deformation of the turbidites of the Eastern Champsaur Basin,
591 dominantly by folding, happened beneath this over-riding allochthon.
592 Shortening within the basin was approximately 4 km with differential
593 shortening through the Champsaur Sandstone stratigraphy implying
594 significant simple shear distributed penetratively through the multilayer
595 (Bürgisser & Ford 1998). It is likely that thick-skinned deformation in the
596 underlying crust continued, eventually arching the Embrunais-Ubaye thrust
597 sheets and uplifting the adjacent Ecrins massif. The challenge facing further
598 investigations of syn-tectonic sedimentation lies in unravelling deformed
599 sedimentary basins to reveal their stratigraphy. These are endeavours that we
600 hope will be encouraged by our study of the eastern Champsaur Basin.

601

602 **Conclusions**

603

604 Linked stratigraphic, sedimentological and structural studies in the Eastern
605 Champsaur Basin, SE France, reveal:

606 1 – The Champsaur Sandstones of the basin represent a continuation of the
607 late Eocene-early Oligocene Annot system. Therefore, the well-known type
608 locations in Provence represent only a small part of this system and their
609 causative turbidity currents continued up to, and presumably over and beyond
610 what is now the Ecrins basement massif. There is no evidence of flow
611 ponding in the Eastern Champsaur Basin, indeed bed character suggests

612 substantial sediment bypass continuing into more distal settings. These
613 locations await further study.

614 2 – The continuity of sand fairways around the orogen can only arise where
615 foredeep axial slopes and hence palaeobathymetry dipped monotonically in
616 the direction of sediment dispersal. Therefore, during the early Oligocene, the
617 Alpine foredeep became progressively deeper clockwise around the western
618 Alpine arc in SE France. This implies greater tectonic subsidence, perhaps
619 coupled with the building of depositional gradients in this direction, during the
620 late Palaeogene. The proto-Ecrins basement massif did not lie at shallow
621 bathymetric levels as suggested in some Alpine syntheses but, during the
622 early Oligocene, lay under water depths significantly deeper than the
623 Provençal sector of the foredeep.

624 3 – The floor to the foredeep basin within which the Annot turbidites were
625 deposited was actively deforming. This has been inferred elsewhere for the
626 type localities of the Annot system to the south of Champsaur where the
627 foredeep megasequence rests on a gently folded substrate of Mesozoic
628 strata. For the Eastern Champsaur, Mesozoic strata have been largely eroded
629 so that the foredeep megasequence unconformably overlies basement. Small
630 pockets of Triassic and Lower Jurassic strata are preserved, folded and thrust
631 into basement beneath this unconformity, testifying to significant crustal
632 deformation predating the deposition of the Nummulitic Limestone. We
633 interpret the pre-Nummulitic deformation to form simply an early part of crustal
634 shortening that progressed steadily through the late Eocene and early
635 Oligocene.

636 4 – The Champsaur Sandstone succession is capped by the Schistes à Blocs,
637 an olistostrome encased in dark shales, a harbinger of the advancing
638 Embrunais-Ubaye thrust sheets. Notwithstanding the likely run-out distance of
639 the causative debris flows ahead of the advancing thrust sheet, this
640 olistostromal formation is not interbedded with the Champsaur Sandstone.
641 Presumably these flows arrived on a basin floor that had largely been starved
642 of sand supply and were not themselves the cause of this starvation.

643 5 - Deformation continued after the deposition of the sedimentary fill to the
644 Eastern Champsaur Basin, with c. 4 km of shortening accommodated in the

645 footwall to the over-riding the tectonic allochthon represented by the
646 Embrunais-Ubaye thrust sheets.

647

648 Our study demonstrates that even in deformed parts of the Alpine orogen it is
649 still possible to establish detailed stratigraphies in turbidites, thereby opening
650 opportunities to extend tectono-stratigraphic investigations elsewhere in this
651 and other orogens. Such work may improve understanding not only of the
652 turbidite systems and their use as analogues for modern deep-water systems
653 but also of the tectonic relationships between orogens and their “forelands”.

654

655

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667

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818

819

820 **Figures**

821

822 Fig. 1. a) Location map of the Eastern Champsaur Basin (boxed area is Fig 4)
823 in the context of the western Alps (modified after Schmid *et al.* 2004). b) the
824 continuity of the Annot sand system in SE France, in the context of other
825 Eocene-Oligocene turbidite units in the external Western Alps (modified after
826 Joseph & Lomas 2004). The grey area represents thrust sheets, including the
827 Embrunais-Ubaye and lateral equivalents that have over-ridden the Annot
828 sand system, which are shown as inferred at depth. SF – Selle Fault.

829

830 Fig. 2. Idealised Nummulitic stratigraphy (modified after Joseph & Lomas
831 2004)

832

833 Fig. 3. Fold geometries, all of which face west, in various parts of the
834 Champsaur Sandstone (see Fig. 4 for place names). a) looking NNW onto Pic
835 Felix Neff, to illustrate the broadly harmonic and relatively long-wavelength
836 folding of the upper stratigraphic packages (unit A6 on Figs 4 and 6). b)
837 looking SE onto polyharmonic folding in sandstone units (unit A3 on Figs 4
838 and 6) directly above the regional detachment zone of the Blue Marls. The
839 visible cliff height is 100m. c) looking SSW onto chevron-style folding in the
840 lower units of the thinly bedded Champsaur sandstone (Unit A1 on Figs 4 and
841 6). The bottom of the image is at an altitude of c 2200m, the skyline reaches
842 up to 3085m.

843

844 Fig. 4. Geological map of the Eastern Champsaur Basin showing the
845 distribution and structure of the mappable units within the Champsaur
846 Sandstone as defined by this study (Fig. 4). The location of lines of section in
847 Fig 5a, b are annotated (X-X' and Y-Y' respectively), along with logged
848 stratigraphic sections (circled numbers, 1-9; Fig. 6) and field photographs of
849 Fig. 5.

850

851 Fig. 5. Selected key panoramas that serve to trace the bed-set stratigraphy of
852 the Champsaur Sandstone, with varying accessibility. The viewpoints are
853 shown on Fig. 4. Stratigraphic boundaries are shown as yellow lines while

854 fold axial traces are shown by red dashed lines. a) illustrates the west face of
855 Le Piquet, a strike-section, together with the location and content of two
856 sedimentary logs (1 and 2 on Fig. 4) that characterise the upper part of the
857 Champsaur Sandstone. b) Is an oblique view of the north side of the
858 Dourmillouse Valley, which provides a dip-section through the fold belt. This
859 hillside is partly accessible permitting ground-truthing of units and completion
860 of measured sections. c) illustrates part of the steep south side of the Fournel
861 Valley. This provides exceptional exposures for the stratigraphy and folds but
862 has limited accessibility.

863

864 Fig. 6. A selection of lithostratigraphic logs with defined mappable units (A1 –
865 A6-A7), correlated through the basin. The locations are shown on Fig. 4.
866 These sections were constructed with direct bed-measurement and field
867 observations – their location being restricted by accessibility in the terrain.

868

869 Fig. 7. Cross-sections through the Eastern Champsaur Basin. The lines of
870 section are shown in Fig. 2. a) the northern basin; b) the Dourmillouse
871 transect.

872

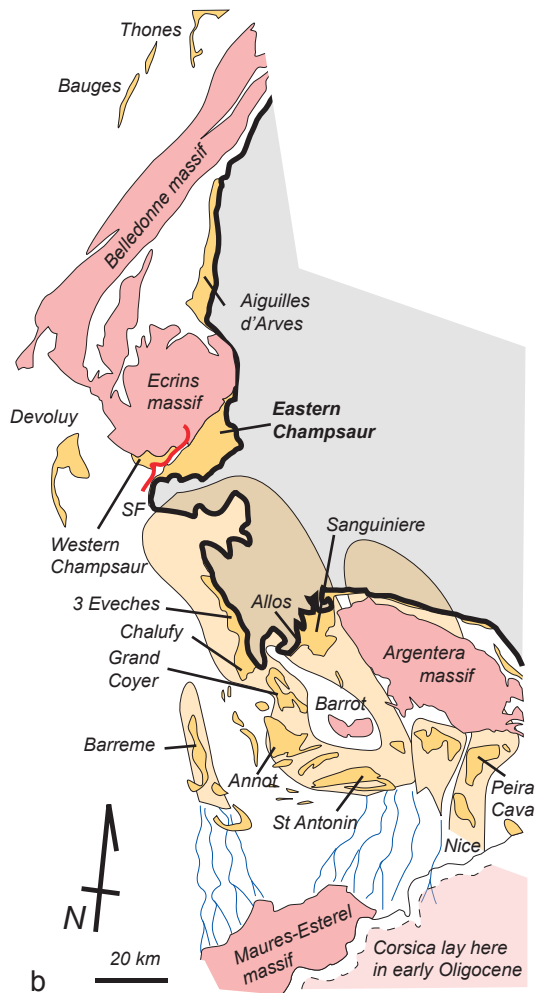
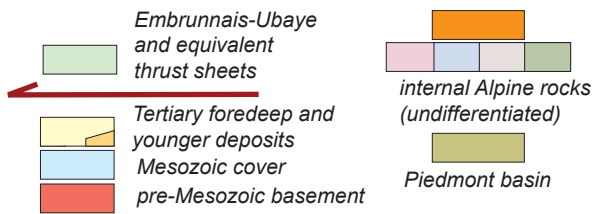
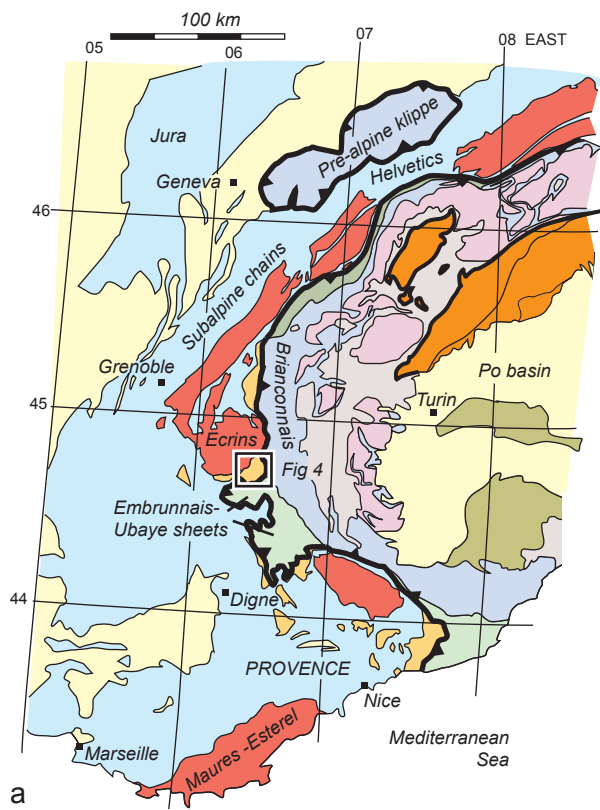
873 Fig. 8. A line-length restoration of the Dourmillouse cross-section through the
874 basin. The thicknesses of bed-sets are shown (intra-formation rulings), and
875 retained between the final state section (a) and its restored version (b).
876 Selected bed-set levels are illustrated (in colour) together with the top
877 basement (in red).

878

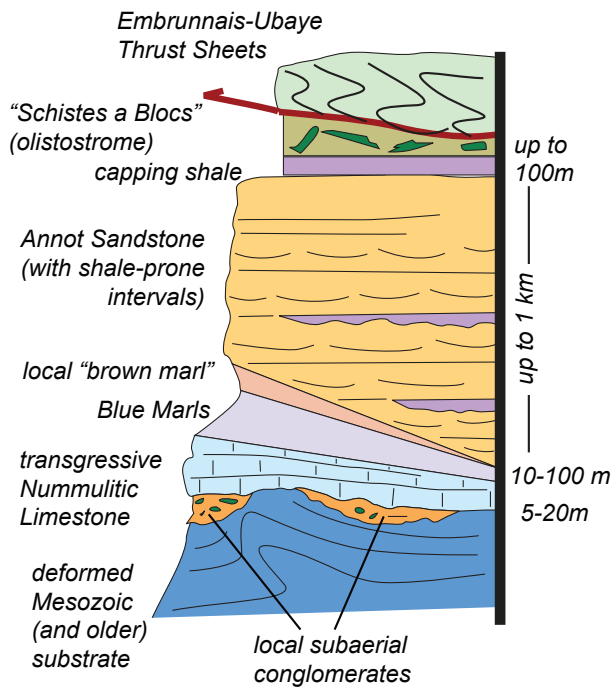
879 Fig. 9. A schematic representation of the extent of the Annot turbidite system,
880 connecting its well-studied proximal areas in Provence to the almost un-
881 studied, more distal regions that include the Aiguilles d'Arves basin, via the
882 Eastern Champsaur Basin described here. The diagram is inspired by an
883 equivalent perspective provided by Joseph & Lomas (2004). Evidence for
884 deformation and its timing beneath the sand fairway within the Ecrins and
885 beneath the Aiguilles d'Arves sector is discussed by Butler (2017).

886

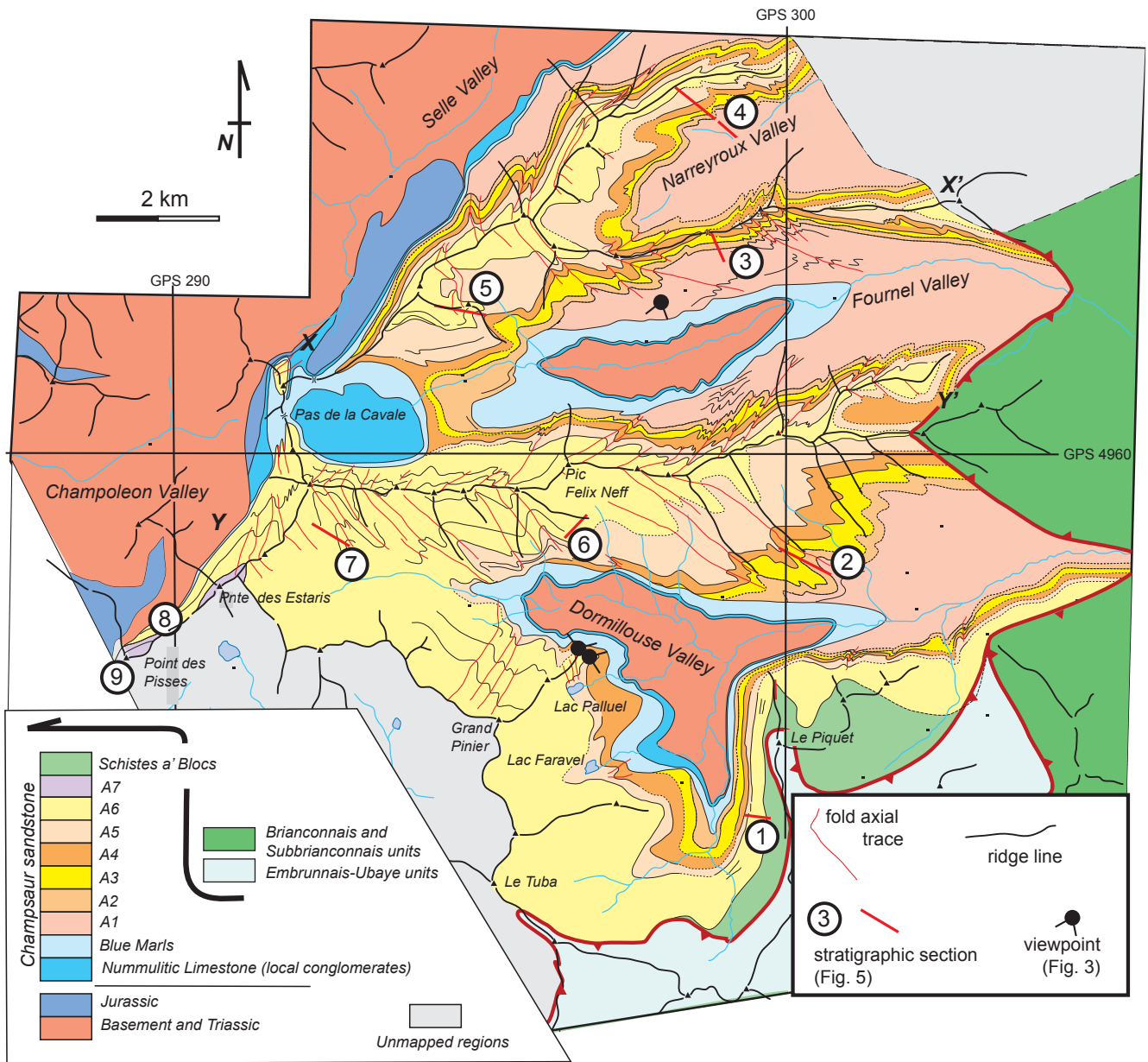
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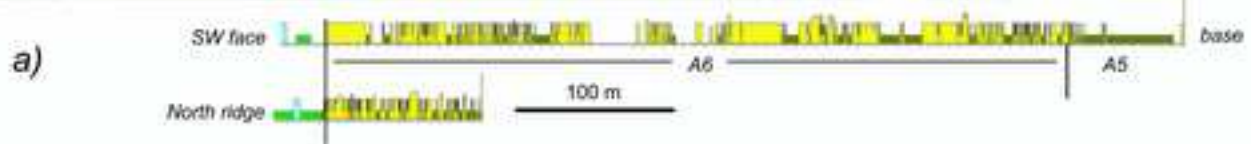
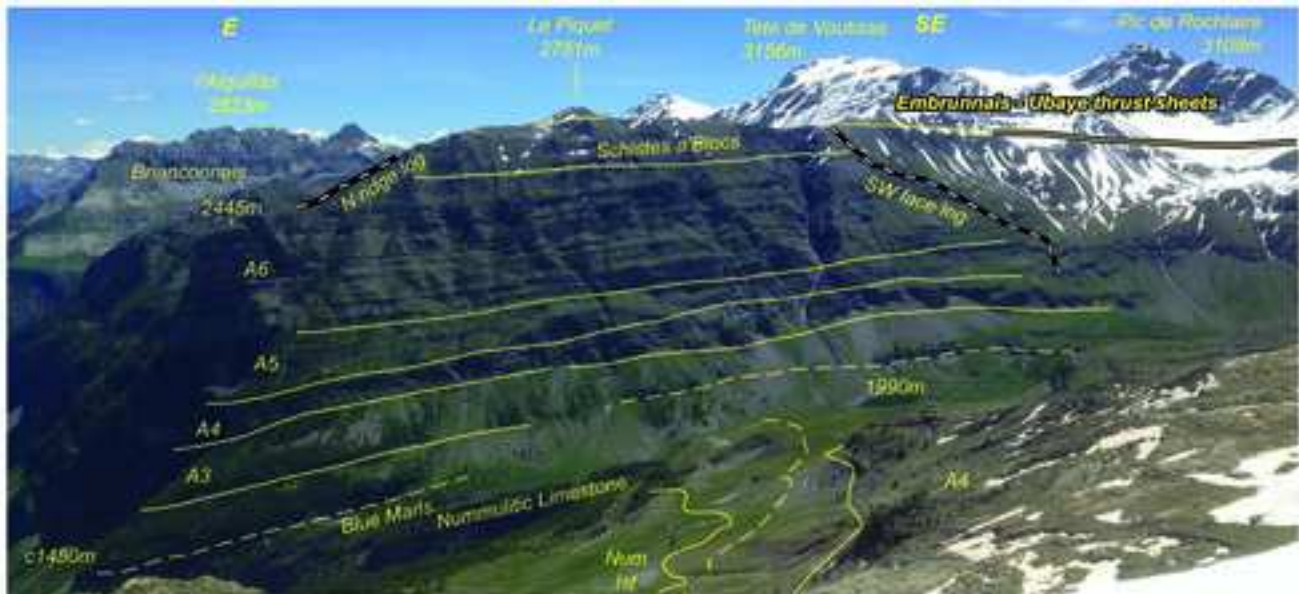


b





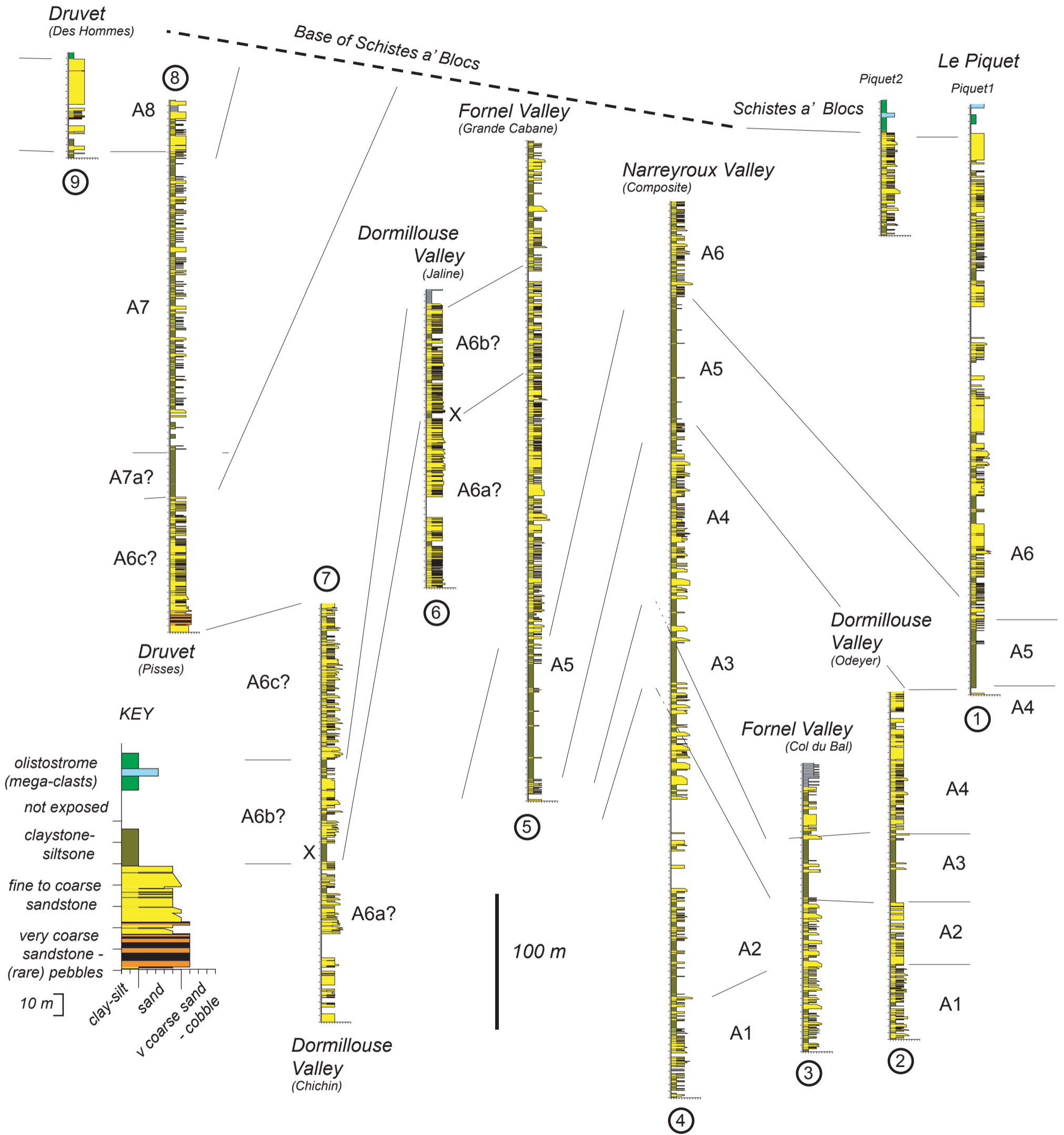


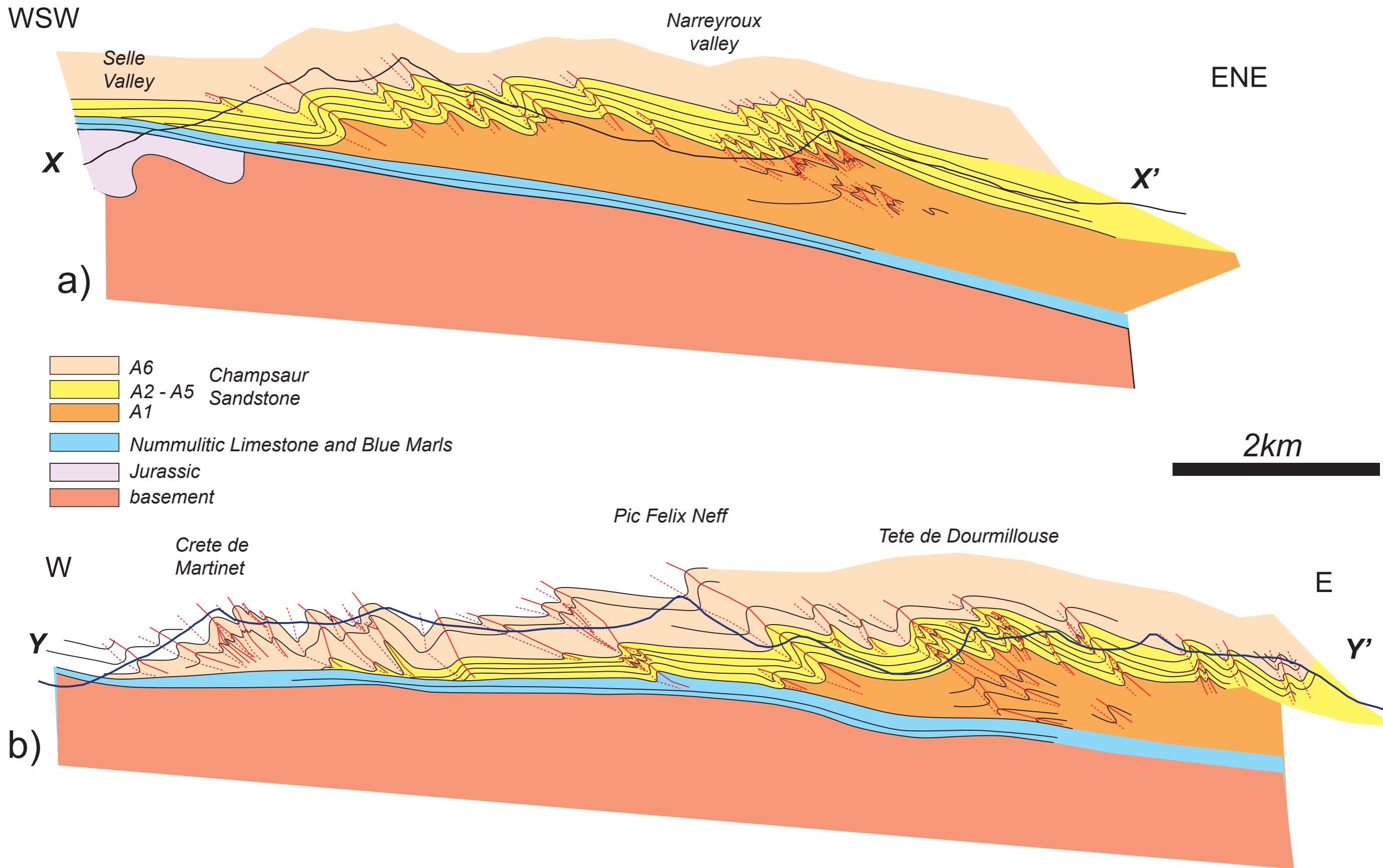


b)



c)







Provencal Basins

Corsica

Maures-Esterel

Eastern
Champsaur
Basin

low-relief
intrabasin
high

Western
Champsaur
Basin

weakly-developed
thrust belt

c 200 km

sea level

sea level



proto -

Ecrins Massif

outfall beyond
proto- Ecrins massif
in canyons ?

Aiguilles d'Arves
basin

Sub-Nummulitic unconformity
truncates folds in Jurassic shales

Inverted basins of
Brianconnais and
Sub-Brianconnais

Dauphinois-Ultra-Dauphinois domains

