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1 **THE SYN-RIFT TECTONO-STRATIGRAPHIC RECORD OF**  
2 **RIFTED MARGINS (PART 1): INSIGHTS FROM THE**  
3 **ALPINE TETHYS**

4  
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10  
11 **ABSTRACT**

12 The tectono-stratigraphic evolution of rift systems is at present poorly understood, especially  
13 the one preceding the onset of oceanic seafloor spreading. Improving our understanding of the  
14 complex, polyphase tectonic evolution of the fossil Alpine Tethys, one of the best calibrated  
15 magma-poor rift systems worldwide, offers great potential for the interpretation of the distal  
16 part of global rifted margins, which notoriously lack data. In this paper, we propose a tectono-  
17 stratigraphic model for the former Alpine Tethys, whose remnants are exposed in the Alps of  
18 Western Europe. We first review the historical evolution of some concepts grounding present  
19 knowledge on the Alpine- and global magma-poor rifted margins. Then we present a spatial  
20 and temporal template for the Alpine Tethys rift system using a “building block (BB)”  
21 approach. This new approach is powerful in that it can integrate high-resolution, structural,  
22 stratigraphic, thermochronological and petrological observations from dismembered outcrops  
23 into a margin-scale template compatible with the first-order seismic architecture defined at  
24 present-day, magma-poor rifted margins. The detailed analysis of the syn-rift sequence of the  
25 Alpine Tethys margins demonstrates that extension migrated and localized while evolving from  
26 the stretching to the necking phase. During hyperextension, rifting was asymmetric and  
27 controlled by in-sequence detachment faulting. Then, the rift system evolved back into a more  
28 symmetric, embryonic spreading system. With this contribution, we aim to allow readers  
29 without knowledge of the Alps to access to this unique “archive” preserving some of the world’s  
30 best-exposed rift structures. Recognising these structures is critical to understand the origin of  
31 some new concepts used to explain present-day, deep-water rifted margins, and to interpret and  
32 predict the tectono-stratigraphic evolution during advanced stages of rifting.

33  
34 **Highlights:**

35 Our “building block” approach enables us to upscale local outcrop observations into a global  
36 margin template

37 The syn-rift sequence of the Alpine Tethys margins is among the best calibrated worldwide

38 The observed rift migration - localization has far-reaching consequences for the interpretation  
39 of global rifted margins

40

41 **Keywords:** Alps, Alpine Tethys, tectono-stratigraphic evolution, polyphase rifting

## 42 INTRODUCTION

### 43 **Why revisit the syn-rift sequence of the Alpine Tethys?**

44 In the past two decades, our understanding of the architecture and evolution of rifted margins  
45 in general, and of the distal parts of magma-poor rifted margins in particular, have been greatly  
46 enhanced (Péron-Pinvidic and Manatschal, 2019). Field studies in the Alpine Tethys have  
47 played a key role in the development of new concepts. Indeed, there are few places where the  
48 syn-rift sequence of a conjugate, magma-poor rifted margin is so well exposed, including the  
49 most distal parts, and where rift-related structures have been examined in such detail. These  
50 studies, closely linked to the progress in seismic imaging and modelling of rifted margins, have  
51 enabled development and testing of many new concepts of rifted margin formation. They also  
52 helped understanding the latest stages of the rifting process (i.e., the breakup phase), the related  
53 tectono-stratigraphic evolution, as well as the resource potential in poorly calibrated distal  
54 margins. The spatial and temporal characterization of rifted margins is a prerequisite to generate  
55 better plate tectonic/geodynamic reconstructions, and to understand the formation of a plate  
56 boundary, a first-order plate tectonic process that remains yet little understood.

57 This paper focuses on the rift-related stratigraphy and tectonic evolution of the Jurassic Alpine  
58 Tethys rift system. Here we use the term “Alpine Tethys” to refer to the (proto-)oceanic  
59 domain(s) and related conjugate rifted margins separating the European-Iberian and African-  
60 Adriatic (micro)continents (Fig. 1a). We use the term rifting to refer to the process thinning the  
61 crust from its initial thickness ( $30 \pm 5$  km) to less than 30 km. We present an historical overview  
62 of the key observations and data that are at the origin of some recent concepts on rifted margin  
63 formation based on more than one century of research in one of the world’s best investigated  
64 orogens. As this review can obviously not include all details and refer to all studies published  
65 over more than a century, we focus on the more recent discoveries, in particular on the distal  
66 parts of the Alpine Tethys margins.

### 67 **Geological and paleo-geographical setting**

68 From a broad-scale perspective, the Alpine Tethys formed contemporaneously with the  
69 subduction of the Paleo- and/or Neo-Tethys Ocean(s) to the east, and with the opening of the  
70 Central Atlantic to the west (Fig. 1b). Aside from the Alpine Tethys, other extensional basins  
71 such as the *Meliata-Vardar* and the *Pyrenean-Bay of Biscay* (Fig. 1a) formed between the  
72 Paleo- and/or Neo-Tethys and Atlantic domains, from the Late Triassic to Cretaceous (Handy  
73 et al., 2010). These (proto-) oceanic domains developed successively westwards, resulting in a  
74 complex, so-called *multi-stage*, plate-kinematic framework that will be discussed later.

75 The Alpine Tethys evolution is part of a *Wilson Cycle* that may have been incomplete, in the  
76 sense that neither seafloor spreading, nor subduction reached a steady-state evolution (Chenin  
77 et al., 2018a). It initiated at the end of the late- to post-orogenic collapse of the Variscan Orogen,  
78 which led to peneplanation in Western Europe, as recorded in the widespread Late  
79 Carboniferous to Permian depocenters. The stratigraphic development of these depocenters and  
80 their link to the Permian and Triassic volcanics and thick Triassic platforms observed in the

81 Adriatic and Briançonnais domains are ill constrained. At a super-regional scale, the Alpine  
82 Tethys rifting initiated in the Late Permo-Triassic in the Dinarides, Hellenides, Turkey, Iran  
83 and Oman (e.g., Stampfli, 1990 ; Robertson & Searle, 1990). In the Vardar in Greece, rifted  
84 margins and fringing MORB crust is dated following the Permo-Triassic rift phase (van  
85 Hinsbergen et al., 2020). Rifting localized in the Alpine Tethys during the so-called *necking*  
86 *phase* in the late Sinemurian/Pliensbachian ( $185 \pm 5$  Ma) (Mohn et al., 2010). This phase  
87 corresponded to both major crustal thinning and the formation of a segmented rift system  
88 comprising three basins, namely from southwest to northeast, the Ligurian, Piemonte and Valais  
89 basins (Fig. 1a). These basins merged in a continuous, proto-oceanic domain, linking the  
90 Central Atlantic with the Tethys realm (Baumgartner, 2013) during the subsequent phase of  
91 mantle exhumation. Mantle exhumation initiated during the Callovian/Bathonian ( $165 \pm 5$  Ma)  
92 (Ribes et al., 2019) and was associated with incipient magmatic activity. Whether the Alpine  
93 Tethys eventually developed into a mature, steady-state spreading system remains debated. At  
94 present, observations and data do not support the existence of a mature Penrose crust in the  
95 Alpine Tethys domain, but that it existed cannot be excluded. The nature and width of the  
96 Alpine Tethys oceanic or proto-oceanic domain have direct implications on a second debate  
97 related to the nature of the subsequent Late Cretaceous to early Cenozoic Alpine subduction.  
98 While McCarthy et al. (2020) suggested that the Alpine Tethys was closed via an Ampferer-  
99 type subduction (i.e., closure of narrow hyperextended rift basins with immature subduction  
100 zones and without formation of Arc magmatic activity), Handy et al. (2010) proposed a more  
101 classical model involving subduction of a wider and mature oceanic domain. Whatever the  
102 nature of the oceanic domain(s) and subduction in the Alps, the fact is that there is little evidence  
103 for a mature oceanic domain and Pacific-type subduction. Indeed, most of the rocks exposed in  
104 the Alps are derived from the former Tethyan margins (Fig. 1c).

105 At present, remnants of the Alpine Tethys margins are scattered along the Alpine belt, over an  
106 area that extends from the Carpathians to the Apennines and further southwest in the Tell, Betics  
107 and Rif regions of northern Africa. In this paper we focus on examples that are assigned to the  
108 Adriatic margin and exposed in the Austroalpine and south Penninic nappes of southeastern  
109 Switzerland and northern Italy; and on the other hand, on the former European margin exposed  
110 in the Western and Central Alps of W-Switzerland, SE France and NW Italy (Figs. 1c and 1d).  
111 The northeastern and southwestern terminations of the Alpine Tethys (Fig. 1a) are less well  
112 understood due either to a more complex rift evolution or a stronger overprint by the later,  
113 subduction-related, back-arc extension (Pannonian Basin, Gulf of Lyon and Tyrrhenian Sea;  
114 Schettino and Turco, 2011). The link between the rift structures and the later Alpine  
115 reactivation, as well as how remnants of the former rifted margin were emplaced and partly  
116 preserved in the Alpine nappe stack is not directly addressed in this study, and we refer to the  
117 studies by Mohn et al. (2011); Beltrando et al. (2014) and Epin et al. (2017) for further  
118 discussion. A review discussing the role of rift inheritance in building collisional orogens can  
119 be found in Manatschal et al. (2021).

120

121 A major, so far unresolved problem in describing the Alpine Tethys is linked to the difficulty  
122 in proposing robust paleogeographic and kinematic reconstructions. The main problem arises  
123 from the kinematics of the Iberia plate (Vissers & Meijer, 2012; Barnett-Moore et al., 2016;  
124 Nirrengarten et al., 2018) because different positions of the Iberia plate imply different widths  
125 of the Alpine Tethys. First, whether the magnetic anomaly M0/J in the southern North Atlantic

126 represents an oceanic magnetic anomaly is debated. If it is not, it cannot be used for restorations.  
127 Second, most of the rifting west and north of Iberia occurred during the Cretaceous magnetic  
128 superchron and is, therefore, not recorded in the magnetic tape recorder. More recent studies  
129 (Nirrengarten et al., 2017; Szameitat et al., 2020) showed that the M0/J anomalies in the  
130 southern North Atlantic did not form on “true” oceanic crust, and can therefore not be used for  
131 kinematic restorations. As a consequence, older restorations for the Alpine Tethys relying on  
132 the M0/J magnetic anomaly as an isochron (e.g., Handy et al., 2010; Vissers et al., 2012; Van  
133 Hinsbergen et al., 2014) must be taken with care. Here we use the interpretation of Decarlis et  
134 al. (2018), who suggested a segmentation of the Alpine Tethys into two sub-basins (Ligurian  
135 and a Piemonte), separated by a Jurassic transform system, like Handy et al. (2010) (Fig. 1a).  
136 Based on this assumption, the Grischun transect discussed in this paper (B''–B''') section in  
137 Figs. 1a and 1c) was part of the Piemonte segment of the Alpine Tethys and was conjugate to  
138 the southeast European margin currently exposed in Sardinia and Calabria (Costamagna, 2016).  
139 Although this hypothesis needs to be further confirmed, we note that it is consistent with  
140 reconstructions by Angrand et al. (2020) and Le Breton et al. (2020).

## 141 **Historical overview**

142 As pointed out by Şengör (2014), the Alps of Western Europe are likely the place where most  
143 new concepts on the formation of rift and oceanic domains and their emplacement in orogens  
144 have been developed (Fig. 2). Major steps in the evolution of the Alpine Tethys and the  
145 characterisation of paleogeographic domains were the pioneering works of Suess (1883; Fig.  
146 2a), Steinmann (1905; Fig. 2b) and Argand (1916; Fig. 2c). Concerning the compressional  
147 reactivation, Bertrand (1884) and Haug (1892) were among the first to demonstrate the evidence  
148 of “nappes” in the Alps. Haug (1892) associated the emplacement of the nappes with the  
149 deposition of the Oligocene Flysch deposits. Termier (1899) then demonstrated that the Alps  
150 were formed by a stack of different “nappes”. Haug (1909) and then Argand (1916) defined a  
151 succession of pre-nappe geosynclines and geanticlines in the Alps, which were the precursors  
152 of what we now refer to as rift basins and rifted margins. Haug (1909) also described the  
153 occurrence of breccias on the flank of the geanticlines, which he interpreted to result from  
154 submarine slumps that were interbedded with deeper marine sediments of the two neighbouring  
155 geosynclines.

156 The start of a more modern view of rifted margins was initiated in the 1960's with the studies  
157 of Elter and Lemoine (Figs. 2d and 2e). Lemoine (1961, 1967) reconsidered the ideas proposed  
158 by Haug (1909), and showed that the structure and related stratigraphy observed in the Alps  
159 resulted from Jurassic and Cretaceous extension. In the same period, Bernoulli (1964) mapped  
160 the first fault-bounded rift basin in the southern Alps (the Generoso Basin). Field observations  
161 from the Alpine Tethys were compared with the first DSDP drilling results in the Central  
162 Atlantic, other present-day rift basins, and seismic images of the Bay of Biscay (Montadert et  
163 al., 1979; Weissert & Bernoulli, 1985). With the advent and success of *Plate Tectonics*, it was  
164 generally accepted in the early 1980's that the Alpine Tethys was floored by “Penrose-type”  
165 oceanic crust and was flanked on both margins by latest Triassic to Late Sinemurian, fault-  
166 bounded rift basins (Lemoine et al., 1987; Eberli, 1988, Bertotti et al., 1993). As a consequence,  
167 stratigraphic contacts between mantle and sediments or between mid-crustal rocks and  
168 sediments, nicely described as “pre-Alpine” prior to Plate Tectonics (e.g., Steinmann, 1905;

169 Decandia & Elter, 1969; Dietrich, 1970; Trümpy, 1975), were reinterpreted as being Alpine  
170 (Hsü & Briegel, 1991). The fact that the ophiolites looked peculiar and did not fit into the classic  
171 three-layered, Penrose-type oceanic crust led the community to assume that they were heavily  
172 overprinted, destroyed or subducted during Alpine convergence.

173

174 It was only in the late 80's and early 90's, when serpentized mantle rocks were drilled  
175 offshore Iberia (Boillot et al., 1987), and detachment systems were described in the  
176 Southwestern United States (Wernicke, 1985), that geologists started to reconsider previously  
177 reported field observations of exhumed mantle (Decandia & Elter, 1969, 1972; Florineth &  
178 Froitzheim, 1994; Lagabrielle, 1994) and detachment faults (Froitzheim & Eberli, 1990;  
179 Froitzheim & Manatschal, 1996). These concepts were reintroduced in the interpretations of  
180 the Alpine Tethys (Lemoine et al., 1987; Froitzheim & Eberli, 1990; Lagabrielle & Cannat,  
181 1990) and applied to present-day "magma-poor" rifted margins (Whitmarsh et al., 2001).

182

183 The significant progress in understanding rifted margins accomplished during the last decades  
184 (e.g., 1990-2020) is intimately linked to the development of new, high-resolution seismic  
185 imaging methods and numerical modelling (Péron-Pinvidic and Manatschal, 2019). However,  
186 access to and comparison with field analogues (Whitmarsh et al., 2001; Lavier & Manatschal,  
187 2006) were fundamental in establishing new concepts for rifted margins. Seismic imaging of  
188 entire rifted margins and attempts to model the way they form provided a first-order  
189 understanding of rift evolution. However, without drillhole data from deep-water rifted  
190 margins, which enabled us to compare the nature of contacts (unconformities or faults) as well  
191 as document ages and depositional settings of sediments, it would not have been possible to  
192 identify the processes occurring during the final stage of rifting, immediately prior to  
193 lithospheric breakup and seafloor spreading. Conversely, it would not have been feasible to  
194 establish new concepts on the tectono-stratigraphic understanding of rifting based on the  
195 fragmented observations derived from the Alps only. Thus, it was historically (and it is still)  
196 the combination of onshore and offshore geological studies that enabled development and  
197 constraints to be placed on new concepts for the processes occurring at higher  $\beta$ -factors when  
198 extending the crust and/or lithosphere.

## 199 **FROM DISPERSED OUTCROPS TO THE ARCHITECTURE OF THE** 200 **ALPINE TETHYS**

201 The Alpine Tethys margins have been partly accreted and subducted during the multi-stage  
202 Cretaceous and Tertiary Alpine compressional events. Thus, when referring to Tethyan margins  
203 in the Alps, we actually mean "remnants" of the former margins, now embedded in the Alpine  
204 nappe stack. Restoring the former margins of the Alpine Tethys relies on the one hand on the  
205 recognition of its remnants, and on the other on our ability to relocate them within the pre-  
206 Alpine rift template. Such a reconstruction is neither easy, nor is there a unique solution.

207 In the following section, we explain how to build a template for the Alpine Tethys margins. In  
208 particular, we describe: (1) the concept of "building blocks" (BB); (2) how to locate the BB in  
209 the spatial Alpine Tethys margin framework; and (3) how reliable the proposed template is and

210 what data are missing to improve and better constrain the understanding of the Alpine Tethys  
211 tectono-stratigraphic evolution. A major aim of this paper is to integrate the concepts and their  
212 derived terminologies used to describe the Alpine Tethys margins into a coherent language  
213 accessible to readers from different geological backgrounds, including those who are not  
214 familiar with Alpine geology. In order not to overwhelm the reader with local details, names of  
215 facies, structures or nappes that sometimes change across national or language borders, we  
216 mention them as sparingly as possible and refer to the original papers where the reader can find  
217 further details and explanations.

## 218 **Datasets and building blocks**

219 To restore the Alpine Tethys rifted margins, we define so-called “building blocks” (BB) (Fig.  
220 3 and Tab. 1). BBs are ideally kilometre-scale outcrops, where primary rift relationships  
221 between basement rocks, fault structures and syn-rift stratigraphic levels are preserved. BBs  
222 provide therefore information about the: (1) nature of the basement (continental crust, exhumed  
223 mantle or fully magmatic crust); (2) depositional environments and stratigraphic thickness, and  
224 hence indications on accommodation, which can be used as a proxy for first-order crustal  
225 thickness by applying isostatic principles (e.g., Tugend et al., 2014); and (3) geometry and age  
226 of rift structures. In the Alps, seven types of BBs can be described. In the following, we refer  
227 to them as BB1, BB2, ... BB7, and exemplify each one with one or several type-localities (Figs.  
228 3a, 3d and 3e and Tab. 1).

229 BB1 (Fig. 3a and Tab. 1) displays classical tilted blocks and half graben bounded by high-angle,  
230 normal faults. Normal faults affect a  $30 \pm 5$  km thick crust, including possible pre-rift  
231 sedimentary layers. The syn-tectonic deposits thicken and diverge towards the bounding faults,  
232 and the corresponding depocenter can be several tens of kilometres in width and several  
233 kilometres in depth (see dark blue layer in BB1; Fig. 3a). Crustal extension is typically low ( $\beta$   
234  $< 1.5$ ), and the total accommodation is minor, as indicated by the facies of the syn- and post-  
235 rift sediments, in line with a bathymetry remaining shallow to moderately deep ( $< 500$  m). The  
236 type locality of BB1 is the Il Motto block in the Upper Austroalpine unit exposed near Livigno  
237 in northern Italy (Eberli, 1988) (Figs. 4d and 4e). Other examples are the Generoso Basin in the  
238 Southern Alps (Bernoulli, 1964) and the Bourg d’Oisans basins in the Western Alps (Lemoine  
239 et al., 1986). BB1 is the most classical building block of rifted margins and failed rifts. It has  
240 been extensively recognized and described from both intracontinental rifts and proximal rifted  
241 margins since the 1960’s. (e.g., Gawthorpe & Leeder, 2000; Gupta et al., 1998).

242 BB2 (Fig. 3a and Tab. 1) shows crustal-scale shear zones typically juxtaposing upper and lower  
243 crustal levels via the attenuation of ductile crustal layers, responsible for major crustal  
244 thinning/necking. An example of BB2 is the Eita shear zone and Grosina detachment fault,  
245 exposed in the Middle Austroalpine Grosina/Campo Nappe at the Swiss-Italian border (Mohn  
246 et al., 2012; 2014) (Figs. 3d et 3e). The link between basement structures and the stratigraphic  
247 record is, however, not well constrained in this area. More recent studies in the Mont Blanc  
248 Massif (Western Alps; Ribes et al., 2020a) and in the Cusio-Biellesse Zone (Southern Alps;  
249 Beltrando et al., 2015; Decarlis et al., 2018) both considered as examples of BB2 ((Figs. 3d et  
250 3e), enabled to investigate the surface expression and its link to the sedimentary evolution  
251 during lithospheric/crustal necking. Both sites display evidence for mid-crustal basement  
252 exhumation linked to detachment faults, erosion and deposition of basement-derived material  
253 in shallow marine silicilastic sequences. The isostatic evolution recorded in the BB2 is complex

254 and suggests uplift, followed by rapid subsidence and sediment aggradation. The total  
255 accommodation and tectono-sedimentary evolution related to BB2 are yet little investigated  
256 and will be discussed in the accompanying paper (Chenin et al. same volume).

257 BB3 (Fig. 3a and Tab. 1) displays a major unconformity, also referred to as “the Briançonnais  
258 unconformity”. This prominent syn-rift unconformity covers the tilted and eroded Triassic and  
259 lower Jurassic platform carbonates and is draped by upper Middle to Upper Jurassic pelagic  
260 carbonates. It is recognized throughout the Western and Central Alps in the so-called  
261 Briançonnais nappes (Lory 1860; 1866; Kilian, 1891; Haug, 1909; Debelmas, 1955; and  
262 Lemoine, 1961) and represents a major characteristic of the former distal European margin.  
263 Crustal thickness of this block is difficult to determine, but by comparison with present  
264 analogues, we reckon it to be  $20 \pm 5$  km. BB3 is typically affected by high-angle faults with  
265 minor offset. The type locality of BB3 is the Fond Froid – Lac de l’Ascension area in the  
266 Briançonnais nappes (Figs. 4d and e), near the city of Briançon (southeastern France; Claudel  
267 and Dumont, 1999).

268 BB4 (Fig. 3a and Tab. 1) comprises successive, in-sequence long-offset extensional detachment  
269 faults responsible for exhumation of mainly upper crustal levels at the seafloor (for details see  
270 Epin and Manatschal, 2018). The exhumed basement is locally overlain by allochthonous  
271 fragments of upper crust and pre-rift and early syn-rift sediments. These structures are covered  
272 by syn-tectonic and post-tectonic deep-water sediments (for details see BB4 in Fig. 3a). The  
273 type-locality of BB4 is the Err-detachment system exposed in the Lower Austroalpine Err  
274 Nappe north of the Julier Pass (Figs. 4d and e) (e.g., Froitzheim & Eberli, 1990; Manatschal &  
275 Nievergelt, 1997; Masini et al., 2012; Epin & Manatschal, 2018). Another nearby example is  
276 the Bernina detachment system (Mohn et al., 2012) (Figs. 4d and e).

277 BB5 (Fig. 3a and Tab. 1) is the least known block and, in contrast to the others, it is not  
278 preserved in one single outcrop. A good description of the tectono-sedimentary relationships is  
279 given in Ribes et al. (2019b). The main characteristic of BB5 is the occurrence of major breccia  
280 bodies of Jurassic and/or Upper Cretaceous age that occur preferentially along the boundary  
281 between the Briançonnais and the Piemonte units, or in units that are interpreted to derive from  
282 the Valais basin (Fig. 4a). These breccias, first described by Haug (1909) and later by Lemoine  
283 (1961; 1967), were interpreted as related to one or more submarine slopes that resulted from  
284 either large faults or flexure. Ribes et al. (2019b) subsequently linked these breccias to  
285 submarine mega-fault scarps that are more favourably formed along an upper plate setting  
286 (Masini et al., 2013; Hauptert et al., 2016; Ribes et al. 2019b). Type localities of BB5 are the  
287 Breccia Nappe in the Préalpes (Chessex, 1959; Badoux, 1962; Hendry, 1972; Weidman, 1972;  
288 Escher et al., 1993; Steffen et al., 1993) (Figs. 4d and e). Other examples are the Monte Galero  
289 breccia system (Decarlis et al., 2015) and the Falknis breccia (Ribes et al, 2020b).

290 BB6 (Fig. 3a and Tab. 1) displays tectonically exhumed serpentized mantle capped by  
291 extensional detachment faults and ophiolites, which are then overlain by deep-water, typically  
292 silica-rich sediments. The nature of the mantle is subcontinental (Mantle-type 1 of Picazo et al.,  
293 2016). The type locality is the Tasna ocean-continent transition (OCT) exposed north of Scuol  
294 in the Lower Engadin valley (Florineth & Froitzheim, 1994; Manatschal et al., 2006; Ribes et  
295 al. 2020b) (Figs. 4d and e). Other examples are the Upper Platta unit (Epin & Manatschal,  
296 2018), Northern Lanzo (Kaczmarek & Müntener, 2008) or the External Ligurides (Marroni et  
297 al., 2002).



298 BB7 (Fig. 3a and Tab. 1) shows magmatic additions that form simultaneous to mantle  
299 exhumation along high- and low-angle faults (for detailed descriptions see Manatschal et al.,  
300 2011; Epin et al., 2019; Coltat et al., 2020). The exhumed, subcontinental mantle mantle is  
301 infiltrated, i.e., reacted with rift-related magma, but did not melt and is not genetically linked  
302 with the overlying basalts (Mantle-type 2 of Picazo et al., 2016). The OCT-basalts are related  
303 to incipient magma production (Amann et al., 2020). The first sediments overlying this domain  
304 are either tectono-sedimentary breccias reworking exhumed mantle, or radiolarian cherts (Ribes  
305 et al., 2019a). The type-locality is the Lower Platta Unit exposed in the Surses valley in  
306 Grischun (Demurs et al., 2001) (Figs. 4d and e). Other examples are the Chenaillet Ophiolite  
307 (Manatschal et al., 2011) (Figs. 4d and e), Southern Lanzo (Müntener & Piccardo, 2003) or the  
308 Internal Ligurides (Tribuzio et al., 2004).

309 Not all observations derived from field studies in the Alps are considered in terms of BBs, and  
310 yet they are important for the restoration of the former Alpine Tethys rifted margins. Examples  
311 include the discovery of a fossil Moho in the Malenco unit (Hermann et al., 1997), and lower  
312 crustal granulites linked to the intrusion of Permian gabbros in the Ivrea zone (Quick et al.,  
313 1992; Barboza et al., 1999) and in the Sondalo region (Petri et al., 2016). These exposures  
314 allowed us to understand the pre-rift crustal structure and inheritance as well as to better  
315 understand the pre-rift lithospheric conditions, to determine the basins exhumation history, and  
316 to assess the Pressure-Temperature-time (PTt) evolution of the basement rocks during rifting.

### 317 **Building a spatial template for Alpine Tethys rifted margins**

318 There are two ways to reconstruct a spatial template of the former Alpine Tethys margins. The  
319 first one is to restore the Alpine nappe stack back to its initial pre-orogenic state. This approach  
320 was first used by Argand (1916) (Fig. 2c), who was one of the first to consider a mobilistic view  
321 and to describe a succession of adjacent paleogeographic domains each one with a characteristic  
322 stratal record. More recently, Beltrando et al. (2014) and Epin et al. (2017) defined diagnostic  
323 criteria to recognize remnants of the former rifted margins in the Alpine orogen and discussed  
324 how they can be restored back to the pre-orogenic stage. However, these authors did not yet  
325 develop the BB approach proposed here. A second way to restore fossil rifted margins is to  
326 define a margin template based on isostatic considerations linked to geological observations  
327 (nature of basement, top-basement, sediments as well as depositional environments and  
328 sedimentary thickness; for a discussion of the approach see Tugend et al., 2014). This approach  
329 resulted in the definition of rift domains. Linking this approach with the average width of  
330 “modern” rift domains observed at present-day rifted margins (Chenin et al., 2017), and  
331 assuming that the processes generating the margins are roughly similar, allows us to place all  
332 the BBs of the Alpine Tethys conceptually along a modern margin template (red line profiles  
333 in Fig. 3a). The BB approach presented in this study enables us to add to the definitions of  
334 domains the key geological information that can be expected in each of the domains. By  
335 combining all these approaches we can use the observations made in the Alps to establish a  
336 more generic description of rifted margins and their evolution, a concept that will be used and  
337 tested in the accompanying paper (part 2) of Chenin et al., in volume).

338 BB1, the “tilted blocks”, have been imaged since the 1970’s in the proximal margin (Montadert  
339 et al., 1979). Until the late 1980’s, the entire margin was assumed to be formed by such  
340 structures. Only with the discovery of exhumed mantle along the Iberia margin (Boillot et al.,

341 1987) made it clear that BB1 cannot explain the whole crustal structure of a magma-poor rifted  
342 margin.

343 Lemoine et al. (1987) was the first to propose the existence of extensional detachment faults in  
344 the Alps, but it was Froitzheim & Eberli (1990) who provided first field evidence for the  
345 occurrence of such structures that characterize BB4. From the 1990's onwards, the structures  
346 related to the former Adriatic margin in the Grischun transect (Figs. 1a and 3b) have been  
347 decrypted. Eberli (1988) and Froitzheim & Eberli (1990) demonstrated that the proximal and  
348 distal margins were controlled by different fault systems, namely pre-Sinemurian high-angle  
349 faults in the proximal margin and post-Sinemurian detachment faults in the distal margin.

350 Lemoine et al. (1987) and Froitzheim & Manatschal (1996) proposed that the Alpine Tethys  
351 margins had to be asymmetric based on the mapping of extensional detachment faults. Although  
352 Decandia and Elter (1969) had already mentioned the existence of exhumed mantle, Florineth  
353 and Froitzheim (1994) and Manatschal and Nievergelt (1997) presented field evidence for  
354 extensional detachment faults related to mantle exhumation in the Tasna and Platta nappes,  
355 which led to the definition of BB6 and BB7. Many studies had investigated the petrology of  
356 magmatic and mantle rocks (Piccardo et al., 1990; Rampone et al., 1995, 1997; Müntener et al.,  
357 2004), however the tectonic link between mantle exhumation along detachment faults and syn-  
358 magmatic activity was only evidenced by Desmurs et al. (2001) and Epin et al. (2019) for the  
359 Lower Platta unit (BB7), and by Manatschal and Müntener (2009) and Manatschal et al. (2011)  
360 for the Chenaillet Ophiolite.

361 While the existence of exhumed mantle and related structures have been described and  
362 generally accepted in the 1990's and early 2000's, the mechanisms of crustal thinning were, at  
363 that stage, still poorly understood. The first models suggested that crustal thinning was  
364 controlled by simple shear (Wernicke, 1985) and detachment systems (e.g., Lemoine et al.,  
365 1987; Froitzheim & Manatschal, 1996). It was only Mohn et al. (2012) who identified evidence  
366 for the necking process (BB2 fingerprints) based on field observations. Ribes et al. (2020) then  
367 described the tectono-sedimentary evolution of this block, highlighting how and when the crust  
368 underwent major crustal thinning.

369 Understanding and linking crustal thinning (BB2), mantle exhumation (BB6), and uplift and  
370 subsidence of the future distal margin (BB3) during the evolution of the Alpine Tethys rifting  
371 became possible thanks to the use of numerical modelling (Chenin et al., 2019, 2020). Finally,  
372 to link massive breccia-bodies with rift models and define corresponding BB5, a detailed  
373 tectono-stratigraphic investigation of the distal European margin was necessary (Ribes et al.,  
374 2019b). Thus, it took more than 50 years to find and define the BBs, the combination of which  
375 led to restore the first-order architecture of the Alpine Tethys margins.

376 Eventually, geologists identified and described BB's all over the Alps; however, it was only in  
377 the Grischun transect (Figs. 1a and 3b) that the relative spatial template of the different blocks  
378 could be directly restored by kinematic restoration of the Alpine nappe stack (Epin et al., 2017).  
379 This exceptional preservation is due to the fact that all of the Adriatic margin remained in the  
380 upper plate of the Alpine subduction and that it was most-likely located against a major pre-  
381 Alpine transform system (the paleo-Subsibic line of the peri-Adriatic system; Fig. 4a)  
382 (Manatschal & Nievergelt, 1997; Mohn et al., 2011; Epin et al. 2017). Indeed, in the Grischun  
383 transect, five out of seven of the BBs have been identified in the nappe stack, each one in a  
384 different nappe (Figs. 4d and e). The restoration of the nappe stack back to the pre-Alpine stage

385 (for details see Epin et al., 2017) enabled us to put the building blocks back to their relative  
386 positions, with BB1 in the Ortler Nappe belonging to the proximal margin, BB2 in the Grosina-  
387 Campo Nappe belonging to the necking zone, BB4 in the Err-Bernina Nappes belonging to the  
388 hyperextended domain, and BB6 in the Upper Platta Unit and BB7 in the Lower Platta Unit  
389 belonging respectively to the inner and outer part of the exhumed mantle domain (Fig. 3b). The  
390 restoration of this nappe stack enabled us to assess the spatial distribution of these BBs across  
391 the margin, but not to establish the detailed continuity and dimensions of the margin along  
392 strike.

393 Concerning the European margin of the Alpine Tethys (Figs. 4d and e), BB1 occurs in the  
394 proximal margin exposed in the Western Alps (Bourg d'Oisans). BB3, BB5 and BB6 in the  
395 distal margin units exposed within the Penninic Nappes, and BB7 in the Chenaillet Ophiolite.  
396 Remnants of BB5, the least understood BB, are preserved all along the Central and Western  
397 Alps at the distal position of the Briançonnais domain and include the Falknis Breccia, the  
398 Breccia Nappe, and the Monte Gallero Breccia (Decarlis et al., 2015; Ribes et al., 2019b). A  
399 comprehensive restoration like that proposed for the Grischun transect is, however, not possible  
400 for the European margin. The reason is that this margin has been more heavily affected by  
401 multi-stage Alpine deformation. Indeed, most of the former distal parts of the European margin  
402 were either subducted or destroyed during the Alpine collision. Moreover, the change in  
403 kinematics from N-S to E-W-directed convergence renders the restoration of the nappe stack in  
404 the Western Alps difficult (Rosenbaum et al., 2002).

#### 405 **From 2D sections to a map view of the Alpine Tethys margins**

406 In order to reconstruct the three-dimensional architecture of the Alpine Tethys rift system, the  
407 lateral continuity of the margins must be integrated. This requires us to first determine the extent  
408 of distinct rift domains along dip (proximal, necking, hyperextended and exhumed  
409 mantle/proto-oceanic domains (Fig. 3a)), and then to correlate and map them along strike. If  
410 we assume that the width/scales of the Alpine margins are comparable to those of present-day  
411 rifted margin domains (Chenin et al., 2017), we can also hypothesize paleogeographical maps  
412 and sections of the former margins, and locate the BBs in such sections/maps in future studies.

413 A next step is to determine whether or not the rift system was segmented, and if so, to locate  
414 the position of the main transfer zones. BB3 and BB5, which are characteristic of an upper plate  
415 margin (hanging wall of the main extensional detachment system during hyperextension; e.g.,  
416 Hauptert et al., 2016), are missing in the Grischun transect. This, together with the polarity of  
417 exhumation faults (BB4; i.e., location of fault breakaways and transport direction), leave little  
418 doubt that the distal part of the Grischun transect belongs to a lower plate margin (footwall of  
419 the main extensional detachment system during hyperextension; Epin & Manatschal, 2018).  
420 Based on an evaluation of the distribution of BB3, which is diagnostic for an upper plate margin,  
421 Decarlis et al., (2018) proposed a change in the polarity from an upper- to a lower-plate margin  
422 across a transfer fault that correspond to the Peri-Adriatic Insubric system. This observation  
423 enabled us to define a pinpoint along the Adriatic margin, which can be linked to the  
424 corresponding pinpoint on the conjugate European margin. Unfortunately, this pinpoint has not  
425 yet been located on the European margin, which makes a restoration of the Alpine Tethys non-  
426 unique for the time being. Our preferred view, which needs however to be confirmed by future  
427 studies, is that the conjugate pinpoint lies at the southern limit of Sardinia and corresponds to  
428 the paleo Iberia-European plate boundary. In such an assumption, the conjugate of the Grischun

429 lower plate transect would correspond to the Sardinia-Calabria section (Fig. 4). If correct, the  
430 Sardinia-Calabria section would represent an upper-plate margin. Furthermore, following this  
431 model, the Southern Alps and the Apennine (former Southern Adria margin), would correspond  
432 to an upper plate margin and be conjugate to the Ebro/Balearian (lower plate) margin during  
433 Jurassic rifting (Fig. 4a).

## 434 **THE SYN-RIFT TECTONO-STRATIGRAPHIC RECORD OF THE** 435 **ALPINE TETHYS**

436 With the BB approach explained in the previous chapter, it is possible to locate outcrops back  
437 into the spatial template of the former Alpine Tethys margins. Some outcrops give access to  
438 rocks derived from different pre-rift crustal and mantle levels, and thus provide information on  
439 the thermal, spatial and isostatic evolution of the rift system. Such information is recorded by  
440 the petrological, thermochronological and stratigraphic data., which enable us to: 1) define the  
441 conditions and timing of crustal thinning, mantle exhumation and onset of magma production,  
442 2) characterise the type of mantle and magma involved, and 3) describe the different spatial  
443 isostatic/bathymetric evolution of rifting. In this section we describe how the tectono-  
444 stratigraphic record of some of the key outcrops can help unravelling the chronostratigraphic,  
445 structural and isostatic evolution of rifting.

### 446 **The stretching phase recorded in the proximal domains (BB1)**

447 The former proximal domains of the Alpine Tethys margins are preserved in the Dauphiné and  
448 Helvetic nappes on the European margin, and in the Upper Austroalpine nappes and Southern  
449 Alps on the Adriatic margin (Fig. 1). These domains are characterized by numerous half-graben  
450 basins (Fig. 4c), some of the most emblematic are described below.

451 *The Bourg d'Oisans Basins (Figs. 4d, 4e and 5a)* located at the European margin include three  
452 tilted blocks, namely from west to east the La Mure, the Taillefer and the Grandes Rousses.  
453 These basement blocks are bounded by master faults, the best exposed and described being the  
454 Ornon normal fault east of the Taillefer Block (Fig. 5a). The Ornon Fault had an approximate  
455 throw of 1500 m (Barf ty, 1985) and largely controlled the formation of a c. 15 km-wide half-  
456 graben. Chevalier et al. (2003) showed that diffuse extension initiated along numerous small  
457 faults with limited throw and low strain-rate between the Hettangian and the Sinemurian, and  
458 that these structures were widely distributed across the entire future proximal domain.  
459 Extension then localized in Early Sinemurian, and most of the 1500 m throw was rapidly  
460 accumulated along the single Ornon fault linked to a fast tectonic subsidence (Fig. 5a). Similar  
461 to the other neighbouring half-graben basins, the syn-rift sedimentary record of the Bourg  
462 d'Oisans Basin grades from thin and shallow marine platform deposits on fault block crests, to  
463 thick and deeper (locally detrital) basinal deposits made of interbedded calcarenites and  
464 hemipelagic to pelagic marls towards the master fault. Based on lithofacies and faunas, the  
465 inferred maximum bathymetry of these European half graben depocenters was about 100-250  
466 m. Their tectono-stratigraphic architecture revealed that they remained underfilled, an  
467 interpretation supported by the local occurrences of olistoliths made of pre-rift and basement  
468 rocks next to normal fault scarps. During the Sinemurian–Pliensbachian, rifting ceased  
469 throughout the future proximal European margin, and the major rift faults were sealed. At that

470 time, depositional depths increased to 250–400 m in basins that record the subsequent  
471 progradation of carbonate platforms from late to post-rift times (Chevalier, 2002).

472 ***The Il Motto Basin (Figs. 4d, 4e and 5b)*** was located at the former northern proximal Adriatic  
473 margin. This basin is one out of several fault-bounded rift basins exposed in the Upper  
474 Austroalpine Ortler Nappe (Eberli, 1988). It exposes a high-angle master fault bounded by a  
475 half-graben extensional basin preserving the relationships between the pre-, syn- and post-  
476 tectonic sediments (Eberli, 1987; Froitzheim, 1988; Conti et al., 1994; Ribes et al., 2019a). The  
477 syn-tectonic sediments are made of Early Jurassic megabreccias that grade upsection and away  
478 from the master fault into finer breccias and then calciturbidites. The megabreccias include  
479 reworked carbonates sourced from the emerged and eroded pre-rift cover of the adjacent fault  
480 scarp. It is important to note that the absence of clastic sedimentation implies that basement  
481 rocks were never exposed and reworked at this stage in the syn-rift sediments. The syn-tectonic  
482 sediments are interleaved with hemipelagic carbonates and shales dated as Early Hettangian to  
483 Late Sinemurian (Furrer, 1985; Eberli, 1985; 1988). The master fault is sealed by ammonite-  
484 bearing carbonates dated as Late Sinemurian to Early Pliensbachian. A younger marker  
485 horizon, corresponding to the Toarcian Oceanic Anoxic Event (T-OAE; 183 Ma) occurs within  
486 the post-tectonic sequence (Ribes et al. 2019a). Lithofacies, faunas and ichnofacies recorded in  
487 background sediments indicate water depths in the range of 50–200 m during rifting, and deeper  
488 conditions (several hundreds of meters) during the post-tectonic phase (Eberli, 1987). Thus, the  
489 environments remained relatively shallow even after rifting, compared to the future distal  
490 margins. The overall thinning- and fining-upward sequence is interpreted to reflect both an  
491 aggradation-dominated setting (instead of a prograding fan), a progressive decrease of  
492 redeposited sediments with respect to background sedimentation and a starvation of the  
493 sedimentary system, most likely related to a deepening of the setting, and because the Adriatic  
494 micro-continent became disconnected from continent-derived sedimentary sources (Eberli,  
495 1987).

496  
497 ***The Monte Generoso Basin (Figs. 4d, 4e and 5c)*** is one of several rift depocenters that have  
498 been identified on the former southern Adriatic proximal margin. These depocenters formed  
499 from Norian time onward; however, major normal faults began to separate half-graben and 30–  
500 50 km wide horsts only during the Rhetian (Bernoulli, 1964; Bertotti et al., 1993). For the best-  
501 investigated Monte Generoso Basin, Bertotti (1991) showed that its master fault (the Lugano-  
502 Val Grande fault zone) penetrated basement and soled out at mid-crustal levels. Since the whole  
503 domain has not been significantly thickened during Alpine convergence, the present crustal  
504 thickness may not be much thicker than at the end of rifting. Thus, the Generoso Basin is likely  
505 to have formed over a crust that was > 25 km thick. Rifting was linked to hemipelagic and  
506 pelagic sequences, interleaved with calciturbidites, which were deposited with typical *half-  
507 graben related, wedge-shaped* architecture. These sequences include spongolitic, calcarenite  
508 and mud-rich turbidites with deepening upward facies, as well as megabreccias and olistoliths  
509 in the vicinity of major fault scarps. Tectonic activity decreased drastically during the  
510 Sinemurian and ceased during the Pliensbachian.

511 In summary, the syn-tectonic sedimentary record in the Bourg d'Oisans, Il Motto and Monte  
512 Generoso Basins displays wedge-shaped deposits from latest Triassic to Sinemurian–  
513 Pliensbachian time. In the temporal framework of the Alpine Tethys rifting, this corresponds to  
514 the early syn-rift history referred to as *the stretching phase*. From Pliensbachian–Toarcian time

515 onward, the three basins and their surrounding proximal domains recorded passive infill, i.e.,  
516 the upper part of the syn-rift has a post-tectonic depositional geometry. The proximal domains,  
517 however, continued to subside but remained at relatively shallow water conditions, suggesting  
518 that the underlying crust remained relatively thick, i.e., likely > 25 km.

### 519 **The necking phase recorded in the necking/distal domains (BB2 & BB3)**

520 The necking phase has been described in the Mont Blanc region (Ribes et al., 2020a) on the  
521 European margin, in the Cusio–Arona region in the Southern Alps (Beltrando et al., 2015) and  
522 in the Austroalpine Campo–Grosina nappe (Mohn et al., 2010). However, since there is no  
523 sedimentary record in the latter, it will not be further described. All these domains are located  
524 between the proximal and distal domains of the Alpine Tethys margins. The syn-rift  
525 stratigraphy of the necking phase is also preserved in the Briançonnais domain near Briançon,  
526 and at the Sostegno-Fenera region in the Biellese in Northern Italy. Below we discuss the syn-  
527 necking evolution of the different sites.

528 *The Mont Blanc region (Figs. 4d, 4e and 6a)* has been interpreted to preserve the necking  
529 domain of the former European margin (Masini et al., 2013; Ribes et al., 2020a). In contrast to  
530 more proximal regions that display half-graben deposits, the syn-rift record of the necking  
531 domain is more complex. The pre-rift (Triassic) and early syn-rift (Lower Jurassic) sediments  
532 are either lacking, or are discontinuous and occur as allochthons. In the past, the lack of the pre-  
533 and early syn-rift sediments was interpreted to result either from non-deposition, or from post-  
534 depositional uplift and erosion. Where present, pre- and early syn-rift sediments were assumed  
535 to represent the primary autochthonous sedimentary cover of the Mont Blanc Massif (Landry,  
536 1978). More recently, Ribes et al. (2020a) emphasized that the basement of the Mont Blanc is  
537 locally topped by pre-Alpine black gouges and cataclasites. They suggested that these rocks  
538 indicate the presence of a major extensional detachment fault, which would be responsible for  
539 the exhumation of the Mont Blanc granite and polymetamorphic basement to the seafloor. In  
540 this interpretation, the Mont Blanc Massif corresponds to a core complex domal structure linked  
541 to the tectonic creation of a new depositional surface, referred hereafter as “new real estate”.  
542 The extensional detachment system has also been interpreted to be responsible for crustal  
543 thinning (for examples see Mohn et al., 2012 and Ribes et al., 2020a). Exhumation is best  
544 documented by early to mid-Jurassic siliciclastic sediments, the so-called *Grès Singuliers*.  
545 These sediments rework the above-mentioned gouges/cataclasites and underlying basement,  
546 and unconformably overlie the basement. Laterally, they overly conformably Lower Jurassic  
547 carbonates (Ribes et al., 2020a). Note that the siliciclastic *Grès Singuliers* are surrounded by a  
548 depositional setting that is dominated by carbonates (Trümpy, 1971). Nevertheless, the detrital  
549 zircon provenance analysis reported in Ribes et al. (2020a) indicates that the *Grès Singuliers*  
550 are locally sourced from the Mont Blanc and Aiguilles Rouges basement. Evidence for several  
551 cycles of erosion and re-sedimentation within these deposits indicate a complex depositional  
552 environment and explain the relative maturity of sandstones, despite the local sourcing. This is  
553 in line with a significant tectonic exhumation and sub-aerial exposure of the local basement  
554 during the necking phase. Additionally, sedimentary structures found in the *Grès Singuliers*  
555 testify to a tidal influence, which further supports a shallow marine depositional environment  
556 (Fig. 6a). The local siliciclastic source for the *Grès Singuliers* became inactive from the  
557 Toarcian–Aalenian onward, either due to the cessation of the Mont Blanc extensional  
558 detachment, and/or to the drowning of the European necking zone (Ribes et al., 2020a). The

559 overlying Middle- and Upper Jurassic syn-rift deposits are post-tectonic (i.e., deposited in a  
560 non-tectonic setting; see Part 2 by Chenin et al., same volume, for definition and discussion of  
561 syn-rift vs. pre/syn/post-tectonic). These post-tectonic deposits correspond to progradational  
562 and aggradational carbonate platforms indicating that the area subsided, although the overall  
563 crust remained thick enough to support shallow marine conditions on the proximal part of the  
564 necking domain. In contrast, on the distal part, a major deepening is documented by the  
565 occurrence of either thicker Upper Jurassic to Early Cretaceous post-rift sediments, or by the  
566 occurrence of deep-water turbidites. A comparable situation can be found all along the external  
567 massifs in the Western Alps.

568 **The Cusio–Biellese region (Fig s. 4d, 4e and 6c)** lies in Northern Italy at the transition between  
569 the proximal and distal southern Adriatic margin. This region displays a complex syn-rift  
570 stratigraphic record that was interpreted by Beltrando et al. (2015) and Decarlis et al. (2018) to  
571 be linked to crustal necking. In this region, Sinemurian shallow-marine carbonates and  
572 sandstones (the so-called San Quirico Sandstone) unconformably overlie a karstified Triassic  
573 platform in the west (Sostegno and Fenera) and basement in the east (Gozano and LAVORIO;  
574 Berra et al., 2009; Beltrando et al., 2015). The analysis of the San Quirico Sandstone indicates  
575 that it has been sourced from a local basement (Permian volcanic rocks and Variscan  
576 metamorphic basement), suggesting basement emersion in the Cusio–Biellese region. The  
577 occurrence of zircon grains with fission track ages of  $177 \pm 6$  Ma in the San Quirico Sandstone,  
578 with a depositional age of Late Pliensbachian to Toarcian ( $\sim 182$  Ma), suggests that the  
579 deposition of this sandstone was contemporaneous with basement exhumation (Beltrando et al.,  
580 2015). This observation is similar to observations made in the *Grès Singuliers* at Mont Blanc  
581 (see previous section; Fig. 6a) and suggests that the Cusio-Biellese region necking may also  
582 have been linked with the development of extensional detachment faults and core complex  
583 structures. Although such structures have not yet been identified in the Cusio–Biellese region  
584 due to poor exposure, the fission track ages suggest tectonic exhumation associated with erosion  
585 (Decarlis et al., 2018). The surrounding depositional environment remained shallow-marine,  
586 and no coeval sub-aerial uplift of the more proximal parts are reported during the necking stage,  
587 ruling out the possibility of long-wavelength dynamic topography (Flament et al., 2013). It is  
588 also important to note that the whole domain is sealed by post-tectonic carbonates of  
589 Pliensbachian age (Berra et al., 2009; Decarlis et al., 2017). This shows a complex stratigraphic  
590 record of the syn-necking depositional sequence. This sequence is characterized by an  
591 unconformable base and a conformable boundary with the overlying post-tectonic sediments.

592 **The Briançonnais domain (Fig s. 4d, 4e and 6b)** belongs to the distal part of the European  
593 margin. During the necking stage, this domain was part of the keystone (H-block) before its  
594 delamination and fragmentation between the two margins (Lavie & Manatschal, 2006; Masini  
595 et al., 2013; Hauptert et al., 2016). Restorations of this domain suggest that it was several tens  
596 of kilometres wide and extended over several hundred kilometres along strike. Early Jurassic  
597 emersion is documented by a regional unconformity and karstification of pre-existing shelf  
598 carbonates, resulting in a major, regional, syn-rift stratigraphic gap (De Graciansky et al., 2011).  
599 This stratigraphic unconformity contrasts with the continuous sedimentary record of the  
600 adjacent areas, which were subsiding at this time (Claudel & Dumont, 1999; Decarlis & Lualdi,  
601 2008). The onset of emersion occurred at the earliest during the Sinemurian, which is the age  
602 of the youngest marine carbonates affected by the karst (Tricart et al., 1988). Drowning of the  
603 Briançonnais domain started at the latest during the late Bathonian ( $\sim 165$  Ma), since shallow

604 marine carbonates of this age locally overlie the karst (Claudel & Dumont, 1999). Due to the  
605 existence of widespread pelagic limestones of Callovian age (~164–161 Ma) draping the entire  
606 Briançonnais domain, rapid drowning and deepening was commonly assumed (e.g., Claudel &  
607 Dumont, 1999; De Graciansky et al., 2011). More recently, however, Hauptert et al. (2016)  
608 suggested that drowning may have started earlier but was not recorded because passive infill  
609 was restricted to the lows on either side of the Briançonnais High. The pelagic nature of the  
610 sediments draping the Briançonnais domain does not necessarily indicate a fast subsidence  
611 and/or a deep water, since pelagic sediments can be deposited in relatively shallow water depth  
612 (below wave base) on isolated highs disconnected from sedimentary sources. The regionality  
613 of the Jurassic karst, and the consistent nature and age of the overlying sediments, suggest that  
614 both emersion and drowning were, from a geological timescale perspective, largely  
615 synchronous over the entire Briançonnais domain and occurred during the necking phase.

616 In summary, the observations from the Mont Blanc Massif and the Cusio–Biellese region and  
617 the Briançonnais domain show that: 1) the base of the syn-tectonic sequence is conformable  
618 over the proximal domain, whereas it is unconformable in the necking domain where core  
619 complexes formed (e.g., new real estate over the necking domain) and over the syn-rift karst  
620 capping the Briançonnais High (BB3); 2) the necking domains and the keystone (H-block)  
621 between them were local regions of either siliciclastic deposits (sandstones) due to local sub-  
622 arial exhumation of basement rocks going along with erosion/deposition, or karstification of  
623 the pre-rift carbonate platform; 3) water-depth remained relatively shallow during the necking  
624 phase and the thickness of related syn-tectonic sediments is therefore limited; and 4) the  
625 transition into the overlying post-tectonic sediments is conformable where sedimentation rates  
626 remained high, and unconformable where starvation occurred due to the fast subsidence post-  
627 dating necking.

## 628 **The hyperextension phase exposed in the distal domains (BB4, BB5 and BB6)**

629 Exposures of the former hyperextended and exhumed mantle domains of the European margin  
630 occur in the Sub-, Lower- and Middle Penninic nappes, and the Pre-Piemonte units, which are,  
631 except the Tasna OCT (Ribes et al. 2020b), strongly overprinted by Alpine deformation. In  
632 contrast, the distal domains of the former Adriatic margin are much better preserved and are  
633 exposed in the Lower Austroalpine Err and Bernina Nappes, and the South Penninic Upper  
634 Platta Unit (Epin et al., 2017). In the western Southern Alps, there are some exposures in the  
635 Canavese Zone that may represent remnants of the former hyperextended margin (Beltrando et  
636 al., 2014; Decarlis et al., 2017). Due to the complex, segmented margin architecture, the along-  
637 strike distribution of the hyperextended domain is complex, in particular along the European  
638 distal margin. For instance, former studies suggested that the Valais Basin formed during  
639 Cretaceous rifting, simultaneous with the Bay of Biscay (Frisch, 1979; Stampfli, 1993; Handy  
640 et al., 2010), despite a Jurassic age of the basin was already proposed by Steinmann (1994) (for  
641 details see Pfiffner, 2014). A Jurassic age for the Valais Basin is supported by the syn-tectonic  
642 sediments in this basin that are Middle Jurassic and mantle exhumation dated late Middle  
643 Jurassic (Ribes et al. 2020b). In this study we propose that the Valais Basin formed as a  
644 segmented, en-echelon, hyperextended basin simultaneous with the Ligurian and Piemonte  
645 basins. In a subsequent stage (after the late Middle Jurassic) these basins connected, giving  
646 birth to the Alpine Tethys proto-oceanic domain (for paleogeographic evolution see Fig. 4).



647 **The Tasna OCT (Figs. 4d, 4e and 7a)** exposes a well-preserved relict of the most distal part of  
648 the European margin (western edge of the Briançonnais continental ribbon) and its transition  
649 into the exhumed mantle domain of the former Valais Basin (Trümpy, 1980). The Tasna Nappe  
650 was intensively studied by Florineth & Froitzheim (1994), Manatschal et al. (2006), and more  
651 recently revisited by Ribes et al. (2020b). The Tasna OCT preserves upper- and lower, pre-rift  
652 continental crust that is juxtaposed along an extensional detachment, the Upper Tasna  
653 Detachment. It is noteworthy that middle crustal material is missing in this section (Ribes et al.  
654 2020b). Continent-wards (i.e., towards the Briançonnais in this area), a stratigraphic interface  
655 can be defined between the upper crust, made of a polymetamorphic Variscan basement, and  
656 pre-rift Triassic dolostones, and Lower Jurassic syn-rift limestones. This succession is part of a  
657 block that is interpreted to represent an extensional allochthon (e.g., Fig. 4 in Ribes et al.  
658 2020b). Oceanwards (i.e., towards the north), the top basement is strongly tectonized and  
659 covered with tectono-sedimentary breccias including clasts of Triassic dolomites and of the  
660 underlying, exhumed basement. Top basement is therefore interpreted as a detachment fault  
661 surface (Ribes et al., 2020b). Two main detachment faults have been identified: the Upper  
662 Tasna Detachment between the lower crust and upper crust, or lower crust and syn-tectonic  
663 sediments where the upper crust is lacking; and the Lower Tasna Detachment between mantle  
664 and the base of the lower crust, or the mantle and syn- or post-tectonic sediments where the  
665 mantle was exhumed to the seafloor (Florineth & Froitzheim, 1994). The first post-tectonic  
666 sediments deposited onto the syn-rift tectono-sedimentary breccias throughout the Tasna OCT  
667 are dark shales. They are locally interbedded with sandstone beds that also seal the allochthon  
668 block located further south in the section. The syn-tectonic sediments are interpreted to be late  
669 Lower Jurassic to Upper Jurassic (Ribes et al., 2020b). This is in line with Ar/Ar cooling ages  
670 yielding an age of  $170.5 \pm 04$  Ma obtained from phlogopites. These phlogopites are derived from  
671 a spinel websterite from within the mantle lying in the footwall of the Lower Tasna Detachment  
672 (Manatschal et al., 2006, see also the discussion in Ribes et al. 2020b).

673 In the nearby Falknis nappe, which has been interpreted to sample the continent-ward  
674 continuation of the Tasna OCT (i.e., the Briançonnais platform), massive breccias interfinger  
675 with syn-rift sediments in the Tasna OCT, and are therefore interpreted as contemporaneous  
676 (Ribes et al., 2020b). These breccias were interpreted to reflect the long-lasting existence of a  
677 mega-fault scarp that is characteristic for an upper-plate margin (e.g., BB5; Ribes et al. 2019b).  
678 The breccias are also interlayered with a starved sediment horizon, comprised of radiolarian  
679 cherts, which marks the termination of rifting. It also suggests deep depositional environments  
680 (below the Carbonate Compensation Depth) at the end of rifting in the Tasna OCT.

681 **The Bernina and Err Nappes (Figs. 4d, 4e and 7b)** preserve the most distal part of the North-  
682 Adria margin. The Err Nappe was the first place in the Alps where rift-related detachment faults  
683 were discovered (Froitzheim & Eberli, 1990). The Err detachment system was subsequently  
684 studied by Manatschal & Froitzheim (1996), Manatschal & Nievergelt (1997), Masini et al.  
685 (2011, 2012), Epin et al. (2017) and Epin & Manatschal (2018). A second major detachment  
686 system has been described in the Bernina nappe and was referred to as the Bernina detachment  
687 system (Mohn et al., 2011, 2012). The Err detachment system is made of at least four, in-  
688 sequence detachment faults (namely the Err, Jenatsch, Agnel and Upper Platta detachment  
689 faults), whose length and size of associated breakaway blocks decrease oceanward (Epin &  
690 Manatschal, 2018). The last detachment is responsible for mantle exhumation exposed in the  
691 Upper Platta Unit. The detachment faults are characterized by diagnostic black gouges and  
692 green cataclasites (Manatschal, 1999), which separate the massive, little deformed but hydrated

693 basement (continental crust or subcontinental mantle) in the footwall either from allochthonous  
694 blocks of continental crust and/or pre-rift sediments, or from late syn- to post-rift sediments  
695 (Masini et al., 2012). These detachment faults reached the seafloor at low angles ( $< 15^\circ$ ), as  
696 shown by the low angle between the exhumation surface and the overlying sediments (Epin &  
697 Manatschal, 2018; Fig. 7b). The lack of quartz mylonites along the detachment fault suggests  
698 that deformation occurred in the brittle regime, at relatively low temperature ( $< 300^\circ\text{C}$ ;  
699 Manatschal, 1999). Crustal basement exhumed in the Err Nappe corresponds to pre-rift upper  
700 crust, pre-rift lower crust has not been reported so far.

701 The Err detachment system is characterized by a discontinuous record of pre-rift sediments,  
702 complex, discontinuous syn-tectonic sedimentary deposits, and continuous post-rift sediments  
703 (Masini et al., 2012; Epin et al., 2017; Ribes et al. 2019a). The tectono-stratigraphy of the syn-  
704 tectonic sediments is well preserved in the so-called Samedan Basin (Masini et al., 2011, 2012;  
705 Epin & Manatschal, 2018). In this basin, pre-rift sediments are only found as allochthonous  
706 blocks overriding a detachment fault (e.g., Bardella allochthon; Fig. 7b). Based on structural  
707 and stratigraphic relationships, and structural restorations of Alpine sections, Masini et al.  
708 (2011) suggested that the Samedan Basin initiated as a classical half graben basin. Continued  
709 extension on the high-angle fault led to its flattening and to the exhumation of deeper basement  
710 levels. At this stage, the initial high-angle fault had become an extensional detachment fault  
711 whose inactive portion lied at a low-angle to the seafloor. In such an example, the syn-tectonic  
712 sediments were deposited directly onto exhumed basement (new real estate). Those sediments  
713 are composed of siliciclastic, basement-derived, immature and locally derived breccias that  
714 directly overlie the detachment system (e.g., member A of the Saluver Fm. of Finger et al.,  
715 1978). In contrast, the sediments near the hanging wall block are exclusively made of carbonate  
716 breccias sourced from the adjacent allochthonous block (e.g., Bardella Fm.; Fig. 7b). The  
717 different composition of the coeval early syn-tectonic sediments have been interpreted to reflect  
718 a change from hanging wall-derived breccias composed of Triassic pre-rift carbonate clasts to  
719 footwall-derived breccias made of basement clasts as the result of the exhumation of basement  
720 rocks along the detachment fault flooring the basin (Masini et al., 2011). These sediments are  
721 overlain by a succession of fining- and thinning-upward turbiditic deposits that grade upwards  
722 into hemipelagic sediments and radiolarian cherts, recording sediment starvation and deepening  
723 of the basin once detachment faulting locally stopped (for details see Masini et al., 2011).

724  
725 ***The Canavese zone*** in northern Italy preserves remnants of the distal part of the South-Adria  
726 margin. Although the sedimentary record of the Canavese zone is poorly exposed and was  
727 largely dismembered during the Alpine orogeny, its similarity with that of the Err nappe  
728 prompted Ferrando et al. (2004) to interpret it as an equivalent distal part (hyperextended  
729 domain). It was subsequently studied by Decarlis et al. (2013) and Beltrando et al. (2015).  
730 Recently, Decarlis et al. (2017) suggested, based on tectono-stratigraphic and paleo-  
731 geomorphological studies, that the Canavese zone belongs to an upper plate margin, and thus  
732 that it would be more comparable with the Tasna OCT. The Canavese displays two types of  
733 exhumed basement: an amphibolite-grade basement with shallow intrusions of Permian  
734 granitoid, interpreted as pre-rift upper crust; and a migmatitic basement with mafic granulite  
735 intrusions, recognized as a pre-rift lower crust (Ferrando et al., 2004). The Middle Jurassic syn-  
736 tectonic sedimentary deposits include both matrix-supported breccias dominated by basement  
737 clasts and matrix-supported breccias rich in Triassic dolostone clasts. Both facies overly  
738 tectonized, exhumed basement covered with tectono-sedimentary breccias (Ferrando et al.,

739 2004). Middle Jurassic breccias are overlain by Middle- to Late Jurassic radiolarian cherts (Bill  
740 et al., 2001). Thus, the overall observations show similarities to the Tasna OCT.

741 In summary, Falknis-Tasna and Err-Bernina display a comparable tectono-sedimentary record  
742 controlled by extensional detachment faults. While Err-Bernina belong to a lower plate margin,  
743 Falknis-Tasna are part of an upper plate margin (Fig. 7c). However, it is important to note that  
744 these sites were not conjugate although they may have belonged to the same margin segment  
745 (Fig. 4). All sites presented here display: 1) dismembered pre-rift and early syn-rift sediments  
746 only preserved in allochthonous and breakaway blocks; 2) exhumed basement along  
747 extensional detachment faults overlain by syn-tectonic tectono-sedimentary breccias containing  
748 clasts of the underlying exhumed basement; 3) evidence for deepening and starvation, which  
749 we here link to the hyperextension phase. Apart from these similarities, there are also some  
750 differences. In the Bernina-Err example (Figs. 7), which belongs to a lower plate margin (Epin  
751 & Manatschal, 2018), detachment faults cut from the upper crust directly into mantle and the  
752 lower crust is missing. In contrast, at the upper-plate margins (e.g., Tasna OCT and Canavese),  
753 lower crustal rocks are exhumed (Fig. 7c).

## 754 **The proto-oceanic domains of the Alpine Tethys (BB7)**

755 Remnants from the proto-oceanic domain are found all along the Alpine belt in the South  
756 Penninic units that constitute the Alpine suture (Fig. 1c). The best-preserved exposures of the  
757 proto-oceanic domain, that escaped subduction and a related high-pressure metamorphic  
758 overprint, are exposed in the Lower Platta Unit in SE-Switzerland and in the Chenaillet  
759 Ophiolite in the French-Italian Alps (Fig. 8).

760

761 ***The Lower Platta unit (Figs. 4d, 4e and 8a)*** belongs to the south Penninic units exposed in SE-  
762 Switzerland. It samples an exhumed mantle domain that includes magmatic additions. This  
763 domain can be restored back to the northern Adriatic distal margin. It was extensively studied  
764 by Dietrich (1969), Manatschal & Nievergelt (1997), Desmurs et al. (2001, 2002), Schaltegger  
765 et al. (2002), Müntener et al. (2004), and more recently by Epin et al. (2017, 2019), Coltat et al.  
766 (2019a, 2019b, 2020) and Amann et al. (2020). The Platta Nappe can be subdivided into two  
767 units based on their lithology, namely the Upper- and Lower Platta Units (Desmurs et al., 2002)  
768 (Fig. 8c). The Upper Platta Unit corresponds to the most proximal part of the exhumed mantle  
769 domain. It is comprised of inherited subcontinental lithospheric mantle with minor magmatic  
770 additions, and a few overlying continental allochthons. In this study, we consider it as part of  
771 the hyperextended domain (Epin & Manatschal, 2018). In contrast, the more distal Lower Platta  
772 Unit is made of (re-)fertilized mantle, i.e., a mantle that reacted with percolating magma (for  
773 details see Müntener et al., 2004), with an increasing volume of intrusive and extrusive  
774 magmatic bodies oceanward (Fig. 8c). It also preserves remnants of a proto-Oceanic Core  
775 Complex (OCC; Epin et al., 2019). The entire exhumed mantle domain is characterized by  
776 intense hydrothermal fluid activity and precipitation of metalliferous mineralization (Coltat et  
777 al., 2019a, 2019b). Extensional structures found in the lower Platta Unit are characterized by  
778 polyphase extensional detachment faults, truncated by late, high-angle normal faults (Fig. 8c;  
779 Epin et al., 2019; Coltat et al., 2020). They include serpentinite cataclasites and serpentinite  
780 gouges, topped by ophicalcites and overlain by tectono-sedimentary breccias, which are typical  
781 fingerprints of Jurassic detachment faults (Epin et al., 2017). An idealized section through the  
782 footwall of a Platta detachment fault includes, from base to top (see Fig. 8a section 1): (1)

783 foliated massive serpentized peridotite or gabbro; (2) serpentized peridotite or gabbro  
784 cataclasites marked by fractures and veins filled with syn-kinematic chlorite and serpentinite;  
785 the intensity of cataclastic deformation increases upward, passing into (3) serpentinite gouges  
786 (fault core).

787 The first sediments overlying the detachment surface are tectono-sedimentary breccias and  
788 ophicalcites (see section 3 in Fig. 8a; Desmurs et al., 2001; Epin et al., 2019). The latter form  
789 in a post-exhumation stage at low temperature (< 100–150°C), except where they are overlain  
790 by basalts resulting in higher formation temperatures (see section 2 in Fig. 8a; Coltat et al.,  
791 2020). The next sedimentary sequence (see Fig. 8b sections 2 and 3) is made of red clays and  
792 radiolarian cherts that include also hydrothermal cherts and reworked, continent-derived  
793 material. These sediments are dated Bajocian to Kimmeridgian in the Alpine Tethys (Bill et al.,  
794 2001; Baumgartner, 2013). The upper part of the radiolarian cherts overlies, and thus post-dates  
795 the basalts and tectono-sedimentary breccia in the exhumed mantle domain. The basalts that  
796 show a T-MOR composition (Desmurs et al., 2002), also referred to as OCT basalts by Amann  
797 et al. (2020), have been interpreted to be syn-tectonic based on the observation of basalts  
798 thickening into high-angle faults and basalts that are deformed during mantle exhumation (Epin  
799 et al., 2019; Coltat et al., 2020) (see also sections 2 and 3 in Fig. 8b).

800

801 ***The Chenaillet Ophiolite (Figs. 4d, 4e and 8b)*** displays pillow lavas with a MORB signature  
802 overlying mafic plutonic rocks and serpentized subcontinental mantle (Mantle type 2, i.e.,  
803 refertilized mantle of Picazo et al., 2016). The first tens to hundreds of meters of the exhumed  
804 mantle are affected by brittle deformation and pervasive hydrothermal circulation. Pillow lavas  
805 are undeformed, implying they were emplaced after mantle exhumation, during a subsequent  
806 phase of high-angle normal faulting (Manatschal et al., 2011) (see observation at Cima le Vert;  
807 Fig. 8b). Manatschal et al. (2011) suggested that these faults served as feeder channels for the  
808 extrusives. The first observed sediments overly the exhumed mantle and/or intrusive mafic  
809 rocks (see observation at La Loubatière; Fig. 8b). They include breccias, graded sandstones and  
810 siltstones that are exclusively made of mantle-derived materials, and to a lesser extent gabbro-  
811 and continent-derived material. Rare basaltic clasts are found in these sediments, which were  
812 emplaced by gravitational processes over the exhumed mantle. This implies formation of  
813 topography during mantle exhumation similar to present-day OCC at mid ocean ridges  
814 (MacLeod et al., 2009). At the Chenaillet Ophiolite, syn-magmatic sediments are rare, apart  
815 from hyaloclastites and subordinate siliciclastic sediments that are interbedded within the  
816 volcanic sequence (Masini, pers. com). No younger, Upper Jurassic to Upper Cretaceous post-  
817 rift sediments are described.

818 In summary, the tectono-sedimentary record of the Lower Platta Unit and the Chenaillet  
819 Ophiolite, both preserving remnants of proto-oceanic crust, shows that: (1) mantle exhumation  
820 was achieved via slip on successive extensional detachment faults later affected by high-angle  
821 normal faults; (2) topography was created during mantle exhumation and resulted in  
822 gravitational sedimentation; (3) the late normal faults may have served as feeder channels for  
823 the overlying volcanic sequences; (4) no pre-rift or early syn-rift deposits exist in the exhumed  
824 mantle domain; and (5) the first syn- to post-magmatic sedimentary deposits (radiolarian cherts)  
825 show the importance of fluids (hydrothermal cherts), reworking (distal turbidites and/or  
826 contourites), and pelagic sedimentation (radiolarites).

827 **READING THE TECTONO-STRATIGRAPHIC TAPE RECORDER OF**  
828 **THE ALPINE TETHYS**

829 **A Wheeler diagram for the Alpine Tethys margins**

830 In this section we describe the syn-rift mega-sequence within a margin-scale Wheeler diagram,  
831 which is an excellent tool to represent the tectono-sedimentary evolution during the entire rift  
832 event (Fig. 9). Ribes et al. (2019a) subdivided the syn-rift mega-sequence into four tracts, the  
833 stretching-, necking-, hyperextension-, and proto-oceanic tracts. In order to compare our results  
834 to seismic sequences from present-day rifted margins, we use here tectonic sequences (TS) that  
835 replace the “tracts” terminology previously proposed (for definition and discussion of TS, see  
836 accompanying paper by Chenin et al., same volume). Each TS encompasses all the sediments  
837 deposited during the period of activity of the corresponding rift phase. Thus, it represents a  
838 chronostratigraphically significant package of sediments related to a distinct phase of the basin  
839 evolution. Each TS includes syn- and post-tectonic packages that are diachronous along the  
840 margin, the former linked to faults, the later sealing faults or showing passive infill of rift-  
841 related relief. Using this approach allows us to describe the sedimentary development during  
842 rifting and to link the kinematic evolution to sediment deposition. This approach relies on the  
843 stratigraphic tape recorder to describe the evolution of rifting from incipient rifting to final  
844 break-up. The first prerequisite is that sediments can be attributed to a specific rift domain, a  
845 second is that a time frame for the different rift events, faults, TS, and magmatic additions can  
846 be defined. However, in contrast to present-day margins where timelines (reflective horizons in  
847 a reflection seismic section) can be followed from the proximal towards the distal parts, in the  
848 case of the Alps, this time frame is more difficult to establish. For the Grischun transect (Fig.  
849 3b), Ribes et al. (2019a) proposed global or regional events that can be used as timelines to  
850 correlate across the margin. These lines are the Rhaetian (base of the syn-rift sequence,  
851 ~200Ma), the Toarcian Oceanic Anoxic Event (~185Ma), the Bajocian-Bathonian bio-Siliceous  
852 Event (~165Ma) and the Tithonian Carbonate Event (~155Ma). It is, however, important to  
853 note that the tectonic events are not necessarily synchronized with these time markers.

854 In this paper, we use the following ages for the successive rift phases in the Alpine Tethys  
855 domain that are based on the studies of Masini et al. (2013) and Ribes et al. (2019a):  $200 \pm 5$  Ma  
856 for onset of rifting,  $185 \pm 5$  Ma for necking;  $175 \pm 5$  Ma for onset of hyperextension along the  
857 distal margin,  $165 \pm 5$  Ma for onset of tectono-magmatic accretion in a proto-oceanic domain.  
858 Ages may in fact differ between the various rift segments of the Alpine Tethys. Here we mainly  
859 refer to the central, i.e., the Piemonte segment. It is important to note that the age of onset of  
860 hyperextension is ill-defined. Note also that we define only the onset of mantle exhumation, but  
861 not the onset of seafloor spreading, since in the Alpine Tethys there is no direct evidence for  
862 steady-state seafloor spreading.

863 Apart from the time frame, we also need to insist on the definition of syn-rift vs. syn-tectonic  
864 that is exemplified within our Wheeler diagram. Syn-rift corresponds to the time between onset  
865 and end of rifting, in our example between  $200 \pm 5$  Ma and  $165 \pm 5$  Ma (Fig. 9). Syn-tectonic  
866 corresponds more specifically to a sedimentary sequence that is deposited during extension and  
867 involving syn-fault deposition. In contrast, pre-tectonic and post-tectonic refer to sediments that  
868 are deposited during the syn-rift time interval, but that precede or post-date active faulting in a  
869 given place. Another important notion introduced already before, but fundamental to the  
870 understanding of the Wheeler diagram shown in Fig. 9, is the concept of “*new real estate*”. It

871 refers to newly created surfaces offered to potential deposition, which are either due to tectonic  
872 exhumation along extensional detachment faults or related to accretion of new oceanic crust.  
873 Introducing “new real estate” is a major modification to the classical Wheeler diagrams used in  
874 basin analysis, and has major implications for the description of the time-space relation between  
875 the stretching, necking, hyperextended and proto-oceanic tracts (for more details see Chenin et  
876 al. this volume). In the Wheeler diagram shown in Fig. 9, we also represent the seven BBs. For  
877 the proto-oceanic tract, we also include the extrusive magmatic additions that were the main  
878 syn-tectonic depositional sequence.

## 879 **Linking the tape recorder with the rift evolution**

880 Rift onset in the western Tethys realm is difficult to define, since rifting nucleated in different  
881 areas at different times: in the east and south (Meliata/Vardar) during the Permo-Triassic, in the  
882 Alpine Tethys at the end of Triassic, and in the southern North Atlantic/Pyrenean domain in the  
883 Late Jurassic-Early Cretaceous (Figs. 1b and 4). In this study we focus on the Alpine Tethys  
884 and more precisely, the Piemonte segment.

885 During the stretching phase (latest Triassic to late Sinemurian/early Pliensbachian:  $200 \pm 5$  to  
886  $185 \pm 5$  Ma), extension was distributed over a wide area, i.e., from the Cevennes in France to  
887 the Trento-plateau in Northern Italy, with the development of many half-graben/graben (e.g.,  
888 Bourg d’Oisans, Generoso and Il Motto Basins; Fig. 10d). All these depocenters were bounded  
889 by listric normal faults that were at a high angle in the upper crust and are interpreted to sole  
890 out in the middle ductile crust (see Bertotti, 1991). The total amount of extension  
891 accommodated during this event may have been minor. Considering the Alpine Tethys, Bertotti  
892 et al. (1993) calculated  $\beta$  values of 1.22 for this stage. Even if some basins may have  
893 accommodated several kilometres of syn-tectonic sedimentary deposits, the overall crustal  
894 thinning remained moderate and widely distributed.

895 The necking phase initiated during the late Sinemurian to early Pliensbachian ( $185 \pm 5$  Ma). It  
896 was associated with the cessation of distributed extension, local uplift of the future distal  
897 domain, and subsequent localization of extension and fast subsidence (Fig. 10c). This evolution  
898 resulted in a complex stratigraphic record of the necking stage recorded by the necking TS.  
899 Local uplift, some few hundreds of meters at maximum, and first occurrence of exhumation  
900 faults on both sides of a keystone (H-Block) resulted in major unconformities that are recorded  
901 at the transition between the proximal and future distal margins (Mont Blanc at the European  
902 margin and Cusio–Biellese and Campo-Grosina at the Adriatic margin). Uplift of the domain  
903 in-between, the so-called Briançonnais domain, can also be observed, resulting in a major  
904 unconformity that can be mapped along the margin and that is of necking age. Numerical  
905 modelling suggests that necking-related uplift occurs when the upper lithospheric mantle,  
906 considered as the strongest layer within the lithosphere, yields beneath a little-thinned  
907 continental crust (Chenin et al., 2019, 2020). This event is followed by a flexural rebound that  
908 can potentially lead to the emersion of formerly marine regions. In the Alps, this event occurred  
909 after the Late Sinemurian and before the Toarcian, i.e., most likely during the Pliensbachian  
910 ( $185 \pm 5$  Ma). As a consequence, the base of the necking TS is characterized by a conformable  
911 contact over the proximal domain that seals the stretching phase faults, and an unconformable  
912 contact over the future distal parts of the margin (for example see the Bardella allochthon, Fig.  
913 7b). The top is either conformable or, in sediment-starved margins like the Alpine Tethys,  
914 paraconformable due to a delayed passive infill (onlaps on rift-related residual topography).

915 The hyperextension phase (Fig. 10b) is well documented in the Lower Austroalpine Err and  
916 Bernina Nappes, which belong to the distal northern Adriatic margin, and in the Tasna OCT,  
917 which belong to the distal European margin. These domains are formed by a crustal wedge that  
918 is interpreted to taper to zero due to the action of in-sequence extensional detachment faults  
919 that propagate towards the future site of breakup. Hyperextension was the only rift phase during  
920 which deformation was clearly asymmetrical, resulting in an upper and a lower plate margin.  
921 At this stage, the entire rifted area was subsiding. While the proximal parts, which were  
922 tectonically inactive, subsided thermally, the distal parts are interpreted to have subsided due  
923 to a combination of crustal thinning (tectonic subsidence) and thermal equilibration (thermal  
924 subsidence). The successive in-sequence detachment faults that developed during the  
925 hyperextension phase rooted in the mantle and eventually led to the exhumation of mantle to  
926 the seafloor (Fig. 10b). Hyperextension and mantle exhumation occurred during and after the  
927 Toarcian anoxic event, but before the deposition of the radiolarian cherts, namely in the 180 to  
928 165 Ma time interval. Sedimentation occurred simultaneous to fast subsidence, resulting in a  
929 general deepening of the depositional environments. While in the lower plate margin breccias  
930 are mainly linked to exhumation faults flooring the hyperextended basins, the upper plate  
931 margin shows massive breccia-bodies linked to long-lived mega-fault scarps lying between the  
932 residual keystone block (BB3) and the hyperextended domain (BB5) (Ribes et al. 2019b; Figs.  
933 3a and 7c).

934 The proto-oceanic phase (Fig. 10a) initiated with the first extrusion of magma that post-dates  
935 initial mantle exhumation. Both are intimately related with extensional faulting (detachment  
936 and high-angle normal faults) and hydrothermal activity. Recent studies show that the first  
937 exhumed mantle at the tip of a magma-poor margin is the subcontinental lithospheric mantle  
938 that was underlying the pre-rift continental crust (mantle-type 1 of Picazo et al., 2016). In more  
939 advanced stages of exhumation, deeper parts of the subcontinental mantle can be exhumed  
940 (mantle-type 2 of Picazo et al., 2016). This type of mantle shows two important features: i) it  
941 was depleted; i.e., it melted during the Permian (e.g., Rampone et al., 1998; Müntener et al.,  
942 2004; McCarthy & Müntener, 2015) and/or showed inherited depleted heterogeneities  
943 (Tribuzio et al., 2004; Sanfilippo et al., 2019); and ii) it was infiltrated by magma during  
944 subsequent rifting (e.g., Rampone et al., 1997, 2020; Piccardo *et al.*, 2004, 2007; Müntener et  
945 al., 2010). In the Alpine Tethys, mantle exhumation was associated with minor magmatic  
946 activity, sometimes interpreted as embryonic seafloor spreading (Manatschal & Müntener,  
947 2009; Manatschal et al., 2011; Epin et al., 2019; Jagoutz et al., 2007; Lagabrielle et al., 2015).  
948 The magma-poor nature of the Alpine Tethys OCT may have originated due to Permian  
949 depletion (Chenin et al., 2017), which was linked to a massive magmatic event throughout  
950 Western Europe following the Variscan orogenic collapse (Bonin et al., 1998; Petri et al., 2017  
951 cum references). The Jurassic rift-related infiltration may have been protracted, with a  
952 magmatic plumbing system that built from the base of the lithosphere to the top. It is assumed  
953 to have initiated with asthenospheric melt infiltration at ca. 185 Ma and to have reached the  
954 surface with first magma extrusions (basalts) and shallow intrusions (gabbros) at about 165 Ma  
955 (Desmurs et al., 2001, 2002; Schaltegger et al., 2002). Recent studies on the first basalts (Renna  
956 et al., 2018; Amann et al., 2020) showed a complex magmatic evolution during the mantle  
957 exhumation phase, partially controlled by the inherited mantle lithosphere composition. At  
958 present, there is no evidence that the Alpine Tethys developed a mature, steady-state spreading  
959 system (Picazo et al., 2016). The radiolarian cherts that are dated from Bajocian to  
960 Kimmeridgian (Bill et al., 2001) were deposited during the accretion of the proto-oceanic

961 domain. These sediments formed at deep marine conditions below the calcite compensation  
962 depth and are time equivalent to pelagic, but not necessarily very deep sediments on the more  
963 proximal parts of the sediment-starved Adriatic microcontinent.

## 964 **Rift phases and their stratigraphic record**

965 Observations from the Alpine Tethys presented in this paper clearly show that the syn-rift  
966 stratigraphic record is complex and cannot be explained by the simple, monophasic rift model  
967 proposed by McKenzie (1978). The Alpine Tethys stratigraphic tape recorder shows evidence  
968 for an abrupt switch from distributed (stretching) to localized (necking) deformation during the  
969 Sinemurian/Pliensbachian ( $185 \pm 5$  Ma). In contrast, precisely identifying the transition from  
970 necking to hyperextension in time and space is difficult in the Alpine system, while the  
971 transition is spatially well defined at present rifted margins where the crust is thinned to less  
972 than 10 km and the whole crust is in the brittle domain (Sutra et al., 2013; Nirrengarten et al.,  
973 2016). Pinto et al. (2015) showed that a characteristic of the hyperextension phase is the  
974 occurrence of mantle-derived fluids percolating faults and sediments and leaving a geochemical  
975 signature. Another major switch occurs when magma becomes the dominant phase, which in  
976 the Alpine Tethys occurs in the Bathonian to Callovian ( $165 \pm 5$  Ma) and coincides with the  
977 proto-oceanic phase.

978

### 979 *Link between rift phase and tectonic sequences*

980 A major conclusion of this paper is that each of the four rift phases results in a distinct syn-  
981 tectonic sedimentary record and can be described by the four successive stretching-, necking-,  
982 hyperextension- and proto-oceanic TS (see also accompanying paper by Chenin et al. this  
983 volume). The reason is that during each phase accommodation is created in a different way  
984 (Fig. 10). The stretching TS is linked to the accumulation of sediments showing characteristic  
985 syn-tectonic, wedge-shaped architectures in distributed fault-bounded basins (Fig. 10d). The  
986 subsequent necking, hyperextension and proto-oceanic TS (Figs. 10c to 10a) have been little  
987 investigated so far, but show diagnostic features reflecting their distinct structural and isostatic  
988 evolution. The necking phase is characterized by a localization of extension simultaneous to  
989 local uplift of a keystone and along hinge zones, and formation of exhumation faults at both  
990 extremities, shortly followed by fast subsidence. As a consequence, the necking tract has an  
991 unconformable base, accommodation is controlled by the interplay of detachment faults and  
992 uplift/subsidence in disconnected and transient depocenters, and its top is conformable if  
993 sedimentation rates are high. The hyperextension tract forms in a rapidly subsiding basin during  
994 in-sequence stepping of extensional detachment faults towards the site of future breakup. As a  
995 consequence, this tract is post-tectonic over the future proximal margins, the syn-tectonic  
996 sequence is stepping ocean-wards on the lower plate but remains stationary over the location of  
997 exhumation in the upper plate. Horizontal accommodation results from the formation of new  
998 real estate and vertical accommodation from rapid subsidence. This combination of the  
999 simultaneous creation of vertical and horizontal accommodation results in banana-shaped  
1000 sequences often referred to as “sag” basins in the present-day hyperextended magma-poor rifted  
1001 margins. The proto-oceanic TS is more complex, often sediment-starved, and includes syn-  
1002 tectonic magma emplacement. The most characteristic observation is that of flip-flop faults  
1003 create highs and lows, which lows becoming highs during subsequent stages. This explains two



1004 main observations: 1) detachment faults are truncated by younger high-angle faults, and 2) the  
1005 occurrence of breccias and debris flows at the break away of younger faults, i.e., over highs.

1006 *Link between deformation and the syn-rift stratigraphic record*

1007 A key in controlling the way accommodation is created (i.e., horizontal vs. vertical) is the  
1008 geometry of faults. This geometry depends on the bulk crustal/lithospheric rheology, which in  
1009 turn controls the strain localization, and hence shape and width of crustal necking zones (Chenin  
1010 et al., 2018b). However, stratigraphic architecture depends also on the availability of sediments  
1011 and the relation between sedimentation and extension rates. In the Alpine system, sedimentation  
1012 during early rifting was controlled by carbonates and thus not controlled by a significant  
1013 siliciclastic source. Locally, basins may have been disconnected and a local-derived siliciclastic  
1014 source evolved (e.g., *Grès Singuliers* in the example of the Mont Blanc system; Fig. 6a). During  
1015 hyperextension, gravitational processes resulted in the deposition of breccias, debris flows and  
1016 turbidites. In the case of the Adriatic margin, distal parts were not fed from the continent, and  
1017 pelagic and hemipelagic sedimentation, together with reworking of exhumed mantle and  
1018 magma emplacement, became the dominant depositional processes. It is important to mention  
1019 that the Alpine Tethys example is sediment starved and underfilled. The fill of a humid clastic  
1020 run-off dominated basin could look very different as discussed in Part 2 by Chenin et al. (this  
1021 volume).

1022 From a structural point of view, two main ways of accommodation creation can be envisaged.  
1023 Either by listric, concave-upward normal faults or by concave-downward  
1024 extraction/detachment faults (Wilson et al., 2001; Mohn et al., 2012; Petri et al., 2019). While  
1025 the first fault type creates fault-bounded basins, the latter fault type results in the creation of  
1026 new real estate. In this case, basins become both deeper and wider with increasing extension,  
1027 with syn-tectonic sediments directly overlying new real estate (Wilson et al., 2001; Gillard et  
1028 al., 2015). It is also important to determine how faults are linked to local flexure and regional  
1029 isostasy, which define not only the morphology of top basement, but also the relative timing  
1030 between extension and the creation of accommodation. During the stretching phase  
1031 accommodation was created mainly by faults, whereas post-rift thermal subsidence was of  
1032 minor importance. This is what can be seen in the stretching domain in the proximal margins  
1033 of the Alpine Tethys. During the subsequent necking phase, the morphotectonic evolution was  
1034 complex and controlled by the formation of a keystone (for a description see Chenin et al.,  
1035 2019, 2020), but, on the scale of the rift system, it was symmetric. Examples of this stage are  
1036 documented in the Mont Blanc and Briançonnais areas along the European margin, and in the  
1037 Cusio–Biellese region on the Adriatic margin (Fig. 6). During the following hyperextension  
1038 stage, however, the morphotectonic evolution of the two conjugate systems was very different.  
1039 On the lower plate margin (northern Adria), accommodation was controlled proximally  
1040 (proximal/inner) by thermal subsidence while distally (distal/outer) it was controlled by the  
1041 formation of new real estate between breakaway blocks and extensional allochthons (e.g., Epin  
1042 & Manatschal, 2018) (Fig. 7). In contrast, the upper plate showed formation of mega-fault  
1043 scarps, which were linked to the formation of prominent and long-lasting mega-breccias and  
1044 gravitational systems that faced oceanward a domain of new real estate (see the 2 settings of  
1045 type-2 basins in Fig. 5 of Masini et al., 2013). As a consequence, these breccias interfinger with  
1046 deep marine sediments at their tips (see Fig. 9 in Ribes et al., 2019b). Some of the breccias may  
1047 also have been deposited over exhumation faults and were moved away and were disconnected  
1048 from their source on the conjugate margin (see Fig. 8 in Epin et al., 2019). Examples are the

1049 Bernina-Err systems or the lower plate and the Falknis-Tasna for the upper plate setting (Ribes  
1050 et al., 2019b) (Fig. 7c). During the formation of the proto-oceanic domain, the system returned  
1051 to a more symmetric state and detachment faults showed an interplay between exhumation and  
1052 high-angle faulting (Manatschal et al., 2011). This deformation phase, which is observed in the  
1053 Platta nappe (Epin et al., 2019), is reminiscent of that observed at slow spreading mid-ocean  
1054 ridges (Sauter et al., 2013; Cannat et al., 2019).

#### 1055 *Syn-tectonic fluids and their link to rift evolution*

1056 Little attention has been paid so far to fluids and their link to the rift evolution and related syn-  
1057 rift sediments. Pinto *et al.* (2015) and Incerpi *et al.* (2017, 2020) showed that the nature of fluids  
1058 was intimately linked to hydrothermal activity during Alpine Tethys rifting. Incerpi *et al.* (2020)  
1059 showed that the fluid signature changed from the necking to the hyperextension phase. During  
1060 the necking stage, marine-derived fluids interacted with crustal rocks resulting in Si-rich fluids.  
1061 Pinto et al. (2015) showed that during the hyperextension phase, marine-derived fluids  
1062 circulated along faults down into the exhuming mantle and were channelled back to the seafloor  
1063 along detachment faults, forming kilometre-scale hydrothermal cells. These fluids interacted  
1064 with the serpentinizing mantle and transported mantle-derived V, Cr, Ni along faults back to  
1065 the seafloor. This is well-documented by enrichment of fault rocks and the overlying syn-  
1066 tectonic sediments in these mantle-derived elements (Pinto et al., 2015). Coupling the fluid  
1067 history with the diagenetic evolution of the syn-rift sediments, as proposed by Incerpi et al.  
1068 (2020), may not only enable us to define the depositional setting and the rift domain in which  
1069 a sediment has been deposited, but also to use new techniques, such as U-Pb on cements to date  
1070 the age of syn-tectonic sequences. Moreover, potential links between the widespread  
1071 development of siliceous sediments and organisms, i.e., radiolarite, and tectonic exhumation  
1072 and related Si-rich fluids may exist (Pinto et al., 2017).

### 1073 **ALONG STRIKE VARIABILITY OF THE ALPINE TETHYS RIFT** 1074 **SYSTEM**

1075 Our description of the Alpine Tethys rift system and of its evolution is mainly based on a few  
1076 building blocks (BBs), most of which are located in the central Piemonte segment of the Alpine  
1077 Tethys (Figs. 1a and 4). However, this 2D view underscores the non-cylindricity and 3D  
1078 complexity of the Alpine Tethys, which can be regarded from different scale perspectives.

1079 From a large-scale perspective, the Alpine Tethys is only one out of at least three rift systems,  
1080 referred to as *Meliata/Vardar/E-Mediterranean*, *Alpine Tethys*, and *Iberia/Pyrenees/Bay of*  
1081 *Biscay* (Fig. 1b). The spatial overlap of these multistage rift systems makes it difficult not only  
1082 to determine the timing of rift onset in individual areas, in particular where such systems  
1083 overlap, but also to unravel the subsidence history, since the post-rift of one stage can  
1084 correspond to the pre-rift of the next stage. An example is, for instance, the thick Triassic  
1085 dolostones in the Adria and Briançonnais domains, which are syn- to post-rift with respect to  
1086 the Meliata/Vardar, but pre-rift with respect to the Alpine Tethys (violet sediments in Fig. 10e).  
1087 Other examples are the Cretaceous extensional systems that are linked to the Pyrenean phase  
1088 of rifting, which overlap Jurassic rift domains in the European margin (Tavani et al., 2018). At  
1089 present, the partitioning of extension between these rift systems remains little constrained since  
1090 no undisputed data exist on the width, rates, or the detailed kinematic movements of any of

1091 these interacting systems. Our preferred view is that the Alpine Tethys rifting initiated during  
1092 the latest Triassic ( $200 \pm 5$  Ma) and stopped in the Late Jurassic time as a failed, approximately  
1093 300 to 400 km-wide, proto-oceanic domain (Li et al. 2013).

1094 At the scale of the Alpine Tethys, three segments can be distinguished, the *Ligurian*, *Piemonte*  
1095 and *Valais* basins (Figs. 1a and 4). Thermochronological and stratigraphic data suggest that  
1096 these basins may have started to become well-defined structural entities during the  
1097 Sinemurian/Pliensbachian (Fig. 4b), and may have merged into a unique, proto-oceanic or  
1098 oceanic domain, which is referred to as the Alpine Tethys, during the Callovian/Bathonian  
1099 (Baumgartner, 2013) (Fig. 4a). However, many questions remain open, especially about the  
1100 southern Ligurian domain. To date, it remains unclear whether this basin had a Triassic  
1101 precursor basin and if it was linked to the eastern Mediterranean, as suggested by its thick and  
1102 locally deep Triassic sequences (Speranza et al., 2012). The Alpine Tethys rift system may have  
1103 had overstepping accommodation zones, similar to what has been proposed for the Cretaceous  
1104 basins in the Pyrenean domain (Lescoutre & Manatschal, 2020). The formation of continental  
1105 ribbons such as the Briançonnais, separating two V-shaped hyperextended domains that were  
1106 set up as overstepping accommodation zones and merged during breakup, may be a reasonable  
1107 explanation. This hypothesis remains yet to be confirmed by field data. However, it is  
1108 interesting to note that in the Biscay and Pyrenean domains, rift segment centres of  
1109 hyperextended basins are often completely reactivated (subducted). In contrast at rift segment  
1110 boundaries rift structures remain preserved in the mountain belt. This has been explained by  
1111 thrusts forming shortcuts to link one segment to another (Lescoutre & Manatschal, 2020;  
1112 Manatschal et al., 2021). This may explain why the best-preserved rift structures occur on both  
1113 sides of the Peri-Adriatic system (e.g., Insubric line that is the boundary between an upper and  
1114 a lower plate; Decarlis et al., 2017) or are linked to the Briançonnais block. In any case,  
1115 whatever the actual architecture of the Alpine Tethys looked like, it was clearly not cylindrical.  
1116 The relationships between the Ligurian and Piemonte segments of the rift system appear, in  
1117 contrast, simpler and seem to coincide with the paleo-Insubric transfer system.

1118 The complex paleogeography of the former Alpine Tethys rift system is not very different from  
1119 that observed at many Atlantic margins. The paleogeography of the Alpine Tethys rift system  
1120 has important implications on how and where sediments were transported into the deep margins  
1121 (e.g., Ribes et al., 2019b). The segmented paleogeography of the Alpine Tethys can partly  
1122 explain why the Alpine terminology is so complex and the paleogeographic domains cannot be  
1123 correlated along the Alps in a simple way. Thus, most of the controversies based on correlations  
1124 between Alpine units and domains may be solved, or at least better defined, when the complex  
1125 pre-Alpine paleogeography and segmentation are better understood.

1126 **THE SYN-RIFT STRATIGRAPHIC ARCHITECTURE OF THE ALPINE**  
1127 **TETHYS: AN OUTLOOK**

1128 The Alps have been the origin of many ideas and concepts presently used not only in orogenic  
1129 belts, but also at rifted margins (De Graciansky et al., 2011). Long before rifted margins were  
1130 seismically imaged, Alpine geologists proposed that the sediments were deposited in large  
1131 basins, first called geosynclines, and later rift basins. Research over the past decades show that  
1132 the Alps are not so much influenced by the subduction of an oceanic domain, but more by the  
1133 reactivation of former rifted margins. Research on rifted margins has mainly been triggered in  
1134 the last years by the energy industry and the related progress in reflection seismic imaging and  
1135 numerical modelling. However, the Alps have played a critical role in the latest development  
1136 of understanding distal rifted margins with the discovery and description of the hereby  
1137 described “building blocks” / BBs, their integration in a margin template and the access of a  
1138 data set and observations that are not accessible at present-day rifted margins. This enabled us  
1139 to develop a model to describe the tectono-stratigraphic evolution of a magma-poor rifted  
1140 margin. A prerequisite for such a model is to integrate time into a spatial reconstruction. This  
1141 can be done by relying on two types of data. On one hand, petrological and thermochronological  
1142 data provide insight into the PTt path of basement rocks. On the other hand, sedimentary  
1143 deposits act as a tape recorder of the evolving deformation and depositional environments.  
1144 Whenever timelines can be identified throughout the margin, they give insight into the  
1145 migration of deformation through time along the margin (e.g., Ribes et al., 2019a). In addition,  
1146 with the development of the BB approach it was possible to build generic sections across  
1147 margins. As each BB represents a “frozen” structural template with specific structural,  
1148 stratigraphic, thermochronological and petrological characteristics, the comparison between the  
1149 different BB characteristics provides insights into the temporal evolution of the entire rift  
1150 system. This methodology enabled Masini et al. (2013) and Ribes et al. (2019a) to demonstrate  
1151 the migration of rift activity from the future proximal margins towards the future distal margins,  
1152 and to point out the onset of rifting in the Alpine Tethys domain, the crucial moment of necking,  
1153 and the timing of first magma-production in the exhumed domain. Moreover, it allowed them  
1154 to identify different rift phases, and to define rift domains and the specific time periods during  
1155 which they formed.

1156 However, the study of rifted margins worldwide, in particular of their distal parts, is notoriously  
1157 limited by the lack data. While many high-resolution 3D seismic blocks of distal margins  
1158 became recently available to researchers, and although numerical modelling is able to reproduce  
1159 some of the key rift structures in 2D and/or 3D, the access to samples from these domains  
1160 remains the biggest issue. Deep-sea drilling is expensive and remains exceptional; where  
1161 available, it is generally restricted to structural highs, rarely transecting deep depocenters and/or  
1162 penetrating the underlying basement. Therefore, a major asset for research on rifted margins in  
1163 the Alps is the possibility to put specific outcrops into the framework of a present-day rifted  
1164 margin, to further calibrate the new models and to make predictions of how the new concepts  
1165 can be used to make projections from well calibrated proximal to undrilled distal domains.  
1166 However, there are also limitations to the extent the Alpine Tethys can be used to compare with  
1167 other margins. Future research will need to better define the 3D spatial framework and temporal  
1168 development of rifting processes. This includes the semi-quantitative analysis of vertical and  
1169 horizontal movements during rifting, the characterization of the successive deformation events,  
1170 and the recognition of related depositional systems. Part II of this review (Chenin et al., this  
1171 volume) will build on observations and concept derived from the Alps to propose a more  
1172 generic model that will then be tested/discussed using global examples of present-day rifted  
1173 margins.

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1181

1182 Author contributions:

1183 Gianreto Manatschal: Writing – original draft, Conceptualization, Investigation,  
1184 Funding acquisition

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1187 Charlotte Ribes: Writing – review & editing, Conceptualization, Investigation

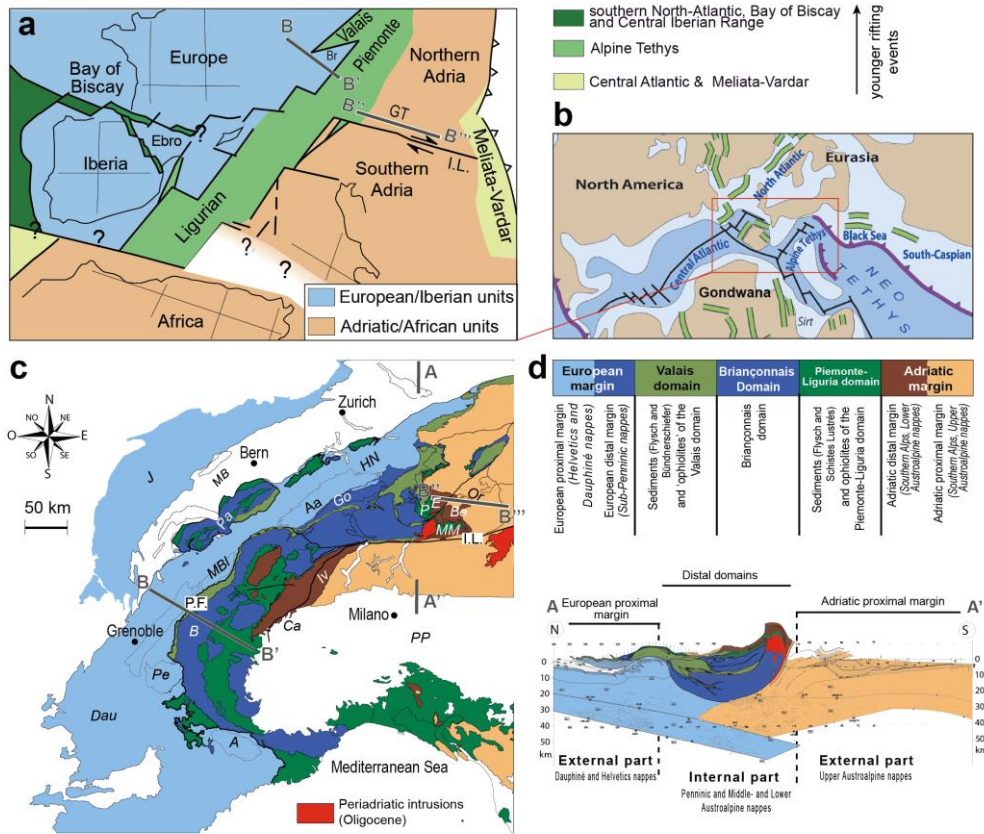
1188 Emmanuel Masini: Writing – review & editing, Conceptualization, Investigation

1189

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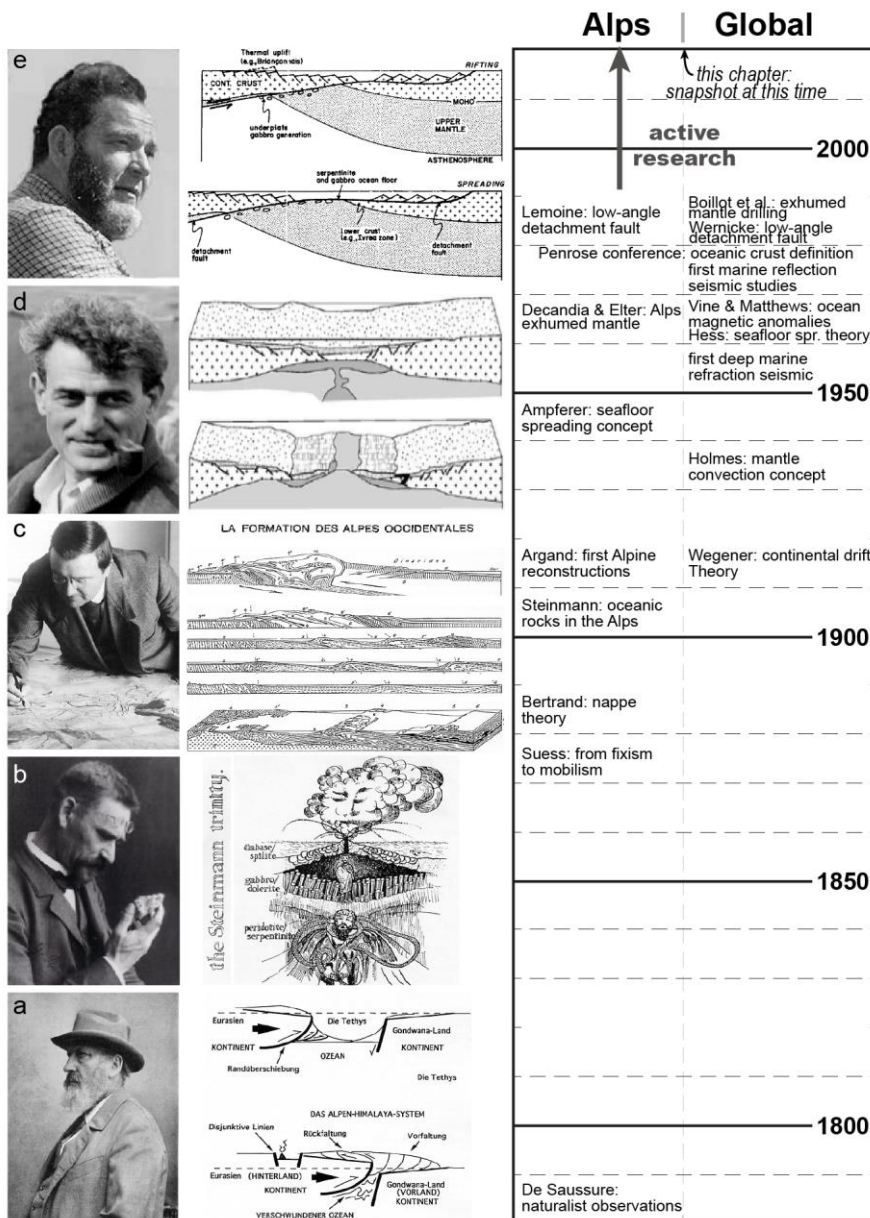
1191 Data Availability Statement:

1192 Data sharing is not applicable to this article as no new data were created or analyzed  
1193 in this study.



1195

1196 **Fig. 1.** (a) Schematic paleogeographic map of the Tethys – Atlantic domain during Early  
 1197 Cretaceous showing three main rift-systems (Meliata-Vardar; Alpine Tethys (Liguria,  
 1198 Piemonte, Valais); and southern North Atlantic with the Bay of Biscay-Pyrenean and  
 1199 Central Iberian Range) and surrounding major plates (Africa and Europe) and  
 1200 microplates (Iberia, Ebro, Southern Adria and Northern Adria) (modified after  
 1201 Manatschal and Müntener, 2009). (b) Paleogeographic map of the western Tethys and  
 1202 Central and North Atlantic system during Early Cretaceous (modified after Frizon de  
 1203 Lamotte et al., 2015). The red box shows the outline of the area shown in Fig. 1a. (c)  
 1204 Tectonic map of the Alps in Western Europe showing the distribution of the main  
 1205 paleogeographic rift domains of the Alpine Tethys in the Alpine orogen (modified  
 1206 after Schmid et al., 2004). (d) Interpretation of the NFP 20 seismic section across the  
 1207 Central Alps showing the position of the main paleogeographic rift domains in the  
 1208 Alpine section (modified after Schmid et al., 1996). For location of the section A-A'  
 1209 see (c). The B–B' and the B''–B''' transects correspond to sections across the former  
 1210 European margin and the Grischun transect (GT) across the northern Adria margin that  
 1211 will be presented in Figs. 3, 9 and 10. Note that these two transects are not conjugate  
 1212 margins but are interpreted to belong both to the Piemonte rift segment. Abbreviations  
 1213 used in Fig. 1: A: Argentera Massif, Aa: Aar Massif, B: Briançonnais Nappes, Be:  
 1214 Bernina Ca: Canavese Nappe, Dau: Dauphiné Nappes, E: Err Nappe, Go: Gotthard  
 1215 Massif, HN: Helevetic Nappes, Iv: Ivrea, J: Jura, MB: Molasse Basin, MBI: Mont  
 1216 Blanc Massif, MM: Malenco-Margna Nappes, Or: Ortler Nappe, P: Platta Nappe, Pa:  
 1217 Prealps, Pe: Pelvoux Massif, PP: Po Plain.

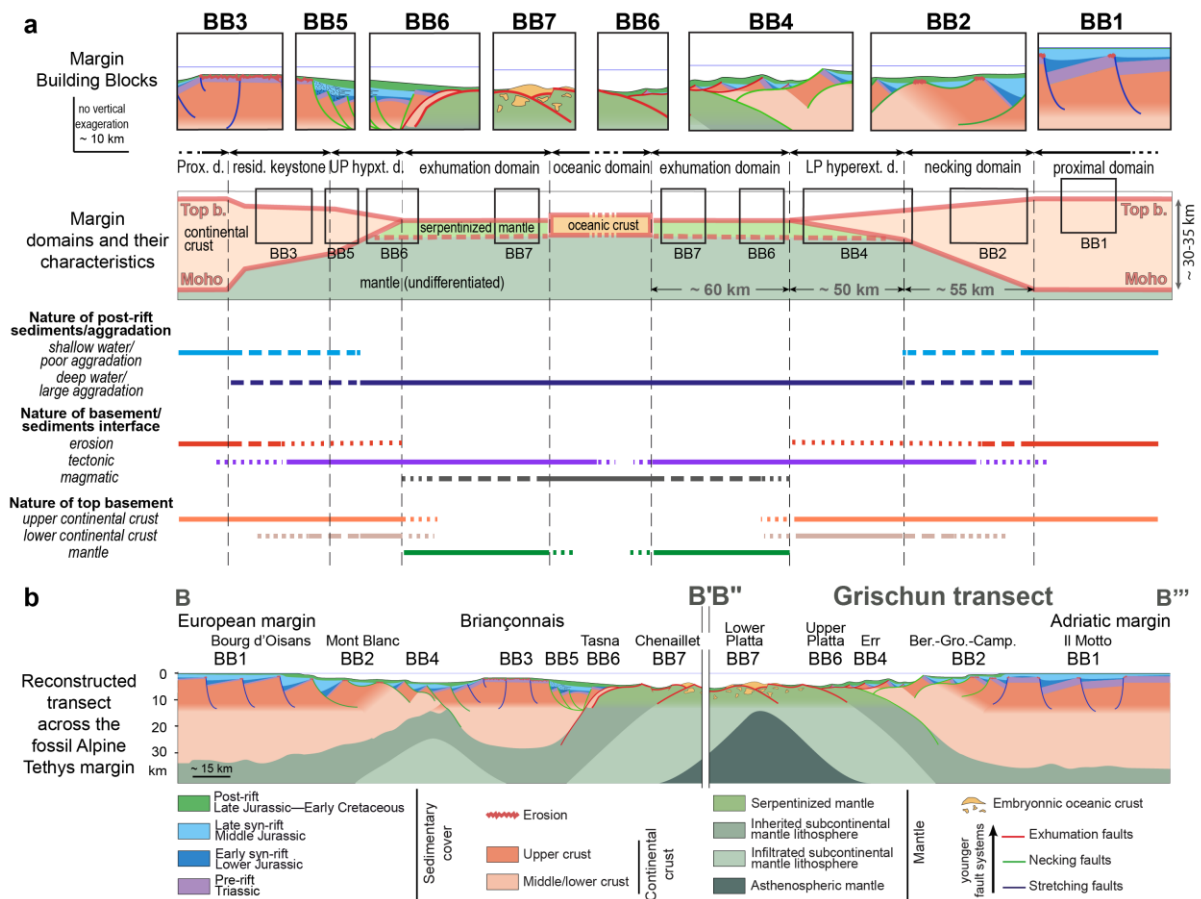


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**Fig. 2:** Evolution of concepts and important players of the “Alpine Tethys”: (a) Eduard Suess was among the first to recognize the existence of former basins in the Alpine domain, which he named Tethys; figure from Suess (1875). (b) Gustave Steinmann was among the first to recognize and map the serpentinized peridotite–“diabase”–radiolarite association (the Steinmann trinity) throughout the Alps and to interpret it as a remnant of a fossil deep ocean floor; figure from Coleman (1977). (c) Emile Argand was among the first to propose detailed tectonic and paleogeographic restorations by applying a mobilistic view to the Alpine system; figure from Argand (1916). (d) Piero Elter was the first who suggested mantle exhumation and therefore the magma-poor nature of the Alpine Tethys; figure from Decandia and Elter (1972). (e) Marcel Lemoine was at the forefront of the present-day interpretation of the Alpine Tethys as a magma-poor, polyphase rifted margin; figure from Lemoine et al. (1987).

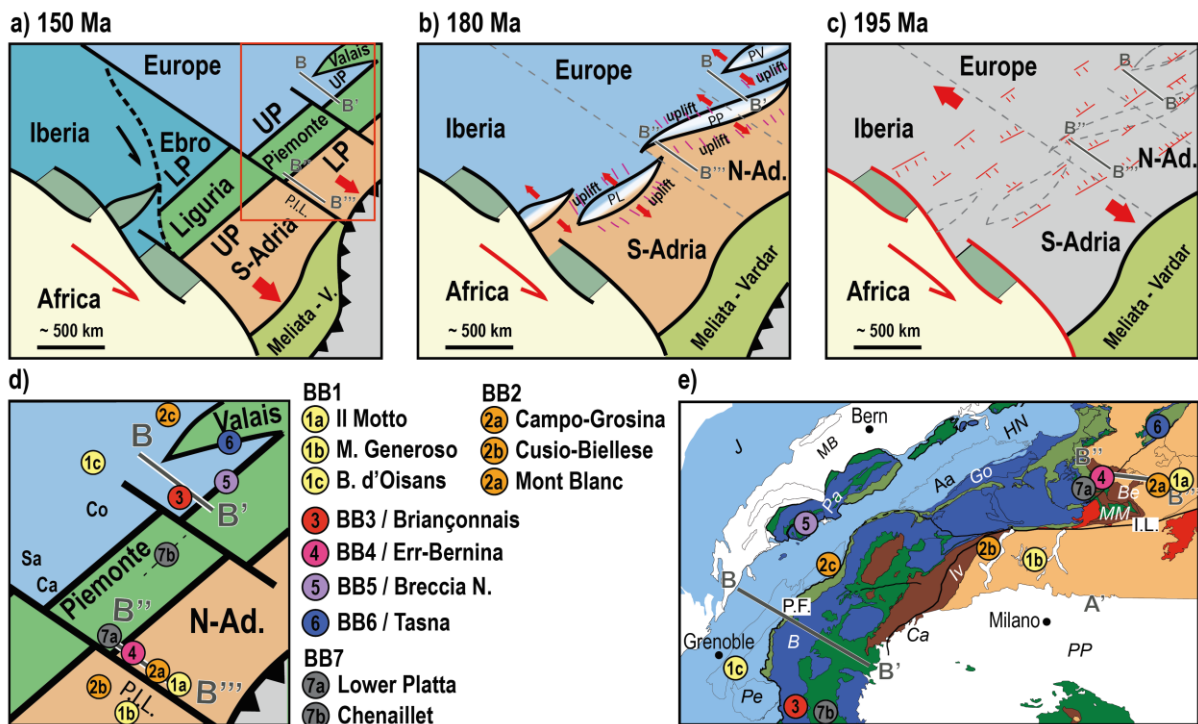
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1231

1232 **Fig. 3:** The Building Block (BB) concept applied to the Alpine Tethys margins: (a) Schematic  
 1233 section across a conjugate magma-poor rifted margin pair and location of the building  
 1234 blocks (BB) (see text for details). The nature of top basement, post-rift sediments, and  
 1235 of their interface vary across the margin and is diagnostic to define rift domains (see  
 1236 Tugend et al., 2014). b) Synthetic reconstruction of a composite transect across the  
 1237 Alpine Tethys rift system (see Fig. 1 for location of B–B' and B''–B''' transects) and  
 1238 location of BBs and projected position of type localities.

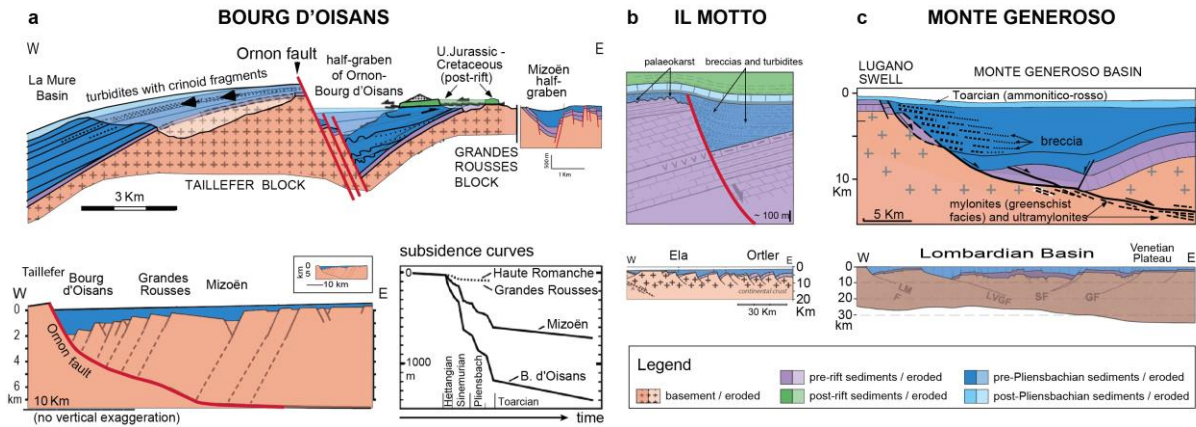




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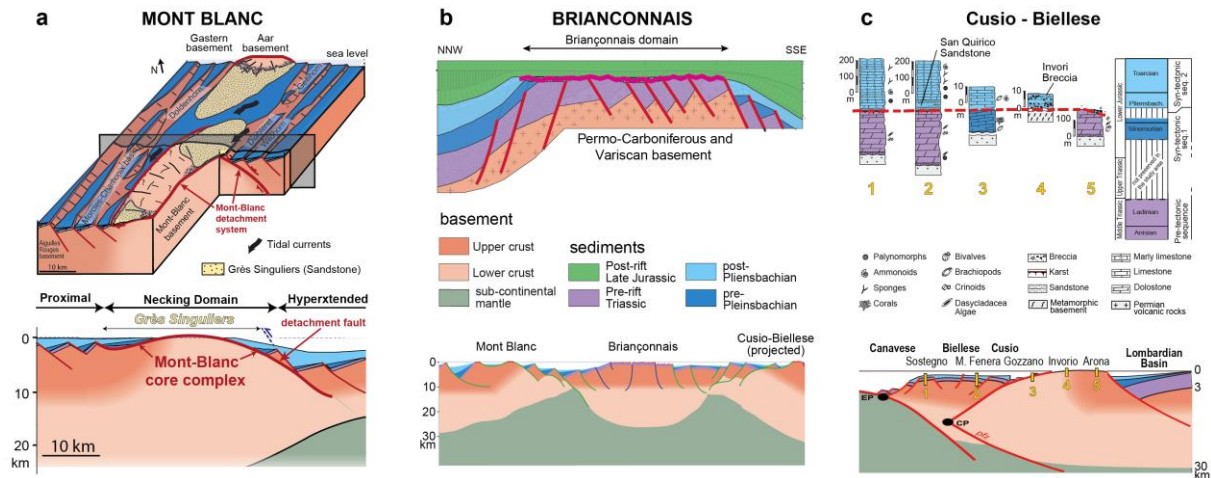
**Fig. 4:** Schematic maps showing the paleogeographic evolution of the Alpine Tethys at: (a) Late Jurassic (150 Ma; assumed end of rifting/spreading in the Alpine Tethys); (b) Late Pliensbachian (180 Ma; end of necking phase); (c) Early Jurassic (195 Ma; stretching phase). Interpretation based on Decarlis et al., 2018). The interpretation assumes that individual basins formed as segmented en-echelon rift segments separated by accommodation/transfer zones. (d) Zoom on (a) showing the locations of the BBs described in this study at Late Jurassic time, and (e) location of the BBs on the present-day structural map shown in Fig. 1c. Abbreviations: P.I.L.: Paleo Insubric Line; PL: Proto Liguria basin; PP: Proto Piemonte basin; PV: Proto Valais basin; LP: Lower Plate; UP: Upper Plate. Transects B-B' and B''-B''' (for position see also Fig. 1).



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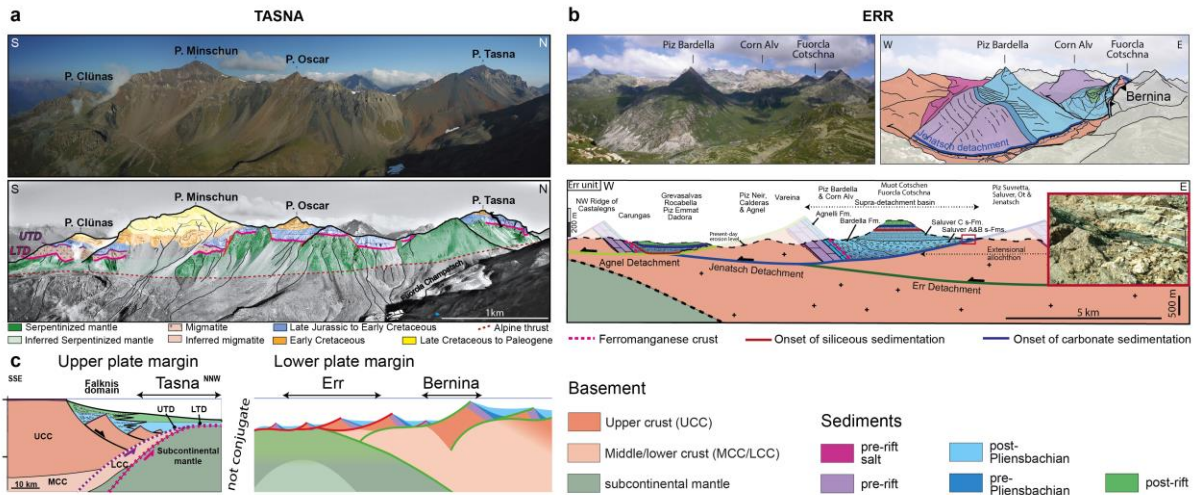
1253 **Fig. 5:** Examples of BB1 exposed in the Alps documenting the stretching stage observed in  
 1254 the proximal Alpine Tethys margins: (a) the Bourg d'Oisans Basin in the proximal  
 1255 European margin (modified after Chevalier, 2002 and De Graciansky et al., 2012); (b)  
 1256 Il Motto Basin in the Upper Austroalpine Ortler Nappe near Livigno in northern Italy  
 1257 belonging to the former northern Adria proximal margin (modified after Eberli, 1988  
 1258 and De Graciansky et al., 2012); and (c) the Monte Generoso Basin in the Southern  
 1259 Alps belonging to the former southern Adria margin (modified after Bertotti et al.,  
 1260 1993 and De Graciansky et al., 2012).

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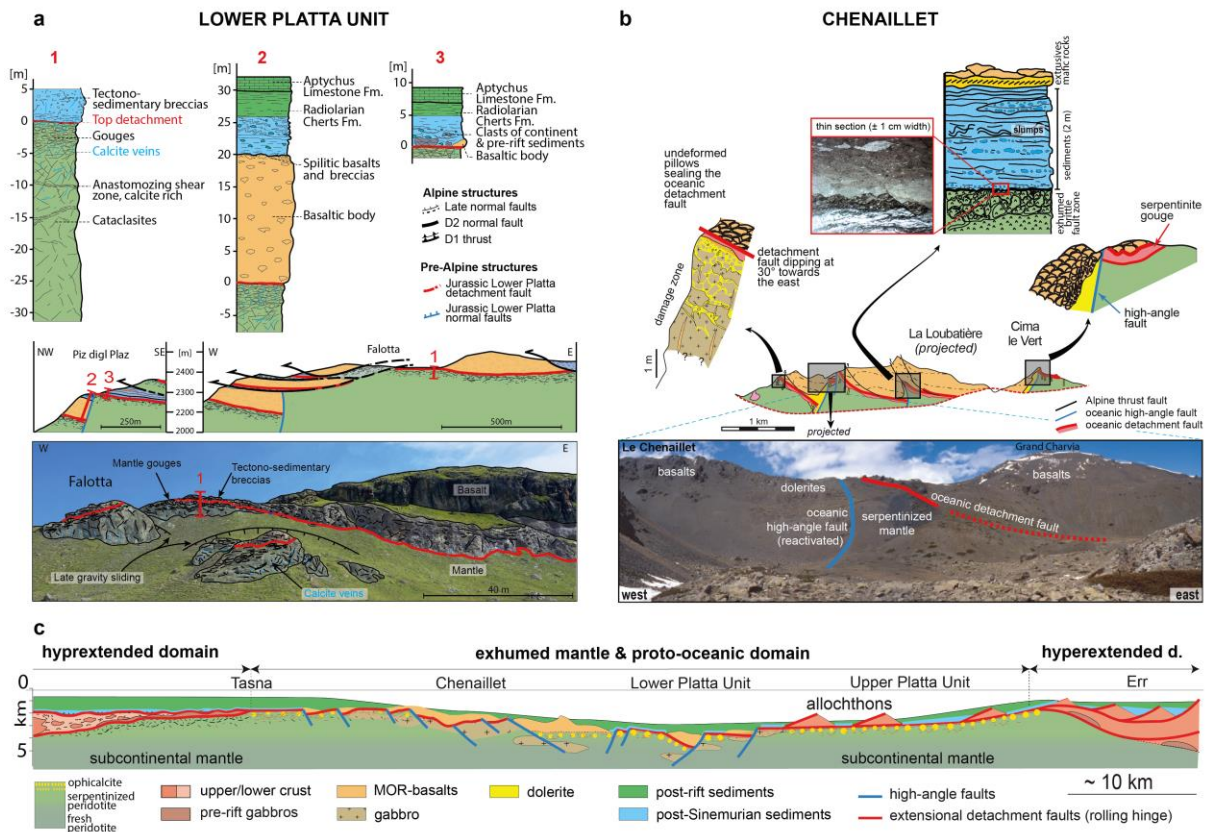
1263 **Fig. 6:** Examples of BB2 and BB3 exposed in the Alps documenting the necking stage  
 1264 observed at the transition between the proximal and distal Alpine Tethys margins and  
 1265 the future distal part before onset of hyperextension: (a) Mont Blanc core complex  
 1266 representing a remnant of the European necking domain (modified after Ribes et al.,  
 1267 2020a); (b) the Briançonnais domain representing the future distal European margin  
 1268 characterized by a major necking unconformity (modified after De Graciansky et al.,  
 1269 2012); (c) The Cusio – Biellese domain showing the stratigraphic record of the  
 1270 southern Adria necking domain (modified after Beltrando et al., 2015 and Decarlis et  
 1271 al., 2018).



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1273 **Fig. 7:** Examples of BB4, BB5 and BB6 exposed in the Alps documenting the hyperextension  
 1274 and early mantle exhumation stage observed in the distal Alpine Tethys margins: (a)  
 1275 Tasna exposed in the Tasna Nappe in the Engadin Window north of Scuol (SE  
 1276 Switzerland) and representing the most distal European/Briançonnais margin  
 1277 (modified after Ribes et al., 2020b) (UTD: Upper Tasna Detachment; LTD: Lower  
 1278 Tasna Detachment); (b) Err exposed in the Lower Austroalpine Err Nappe in SE  
 1279 Switzerland showing the Bardella extensional allochthon at the Julier Pass; photograph  
 1280 on the lower right corner shows the onlap of syn-tectonic sandstones onlapping onto  
 1281 the detachment surface indicated by the black colored fault gouges; modified after  
 1282 Epin & Manatschal (2018) and Ribes et al. (2019a).

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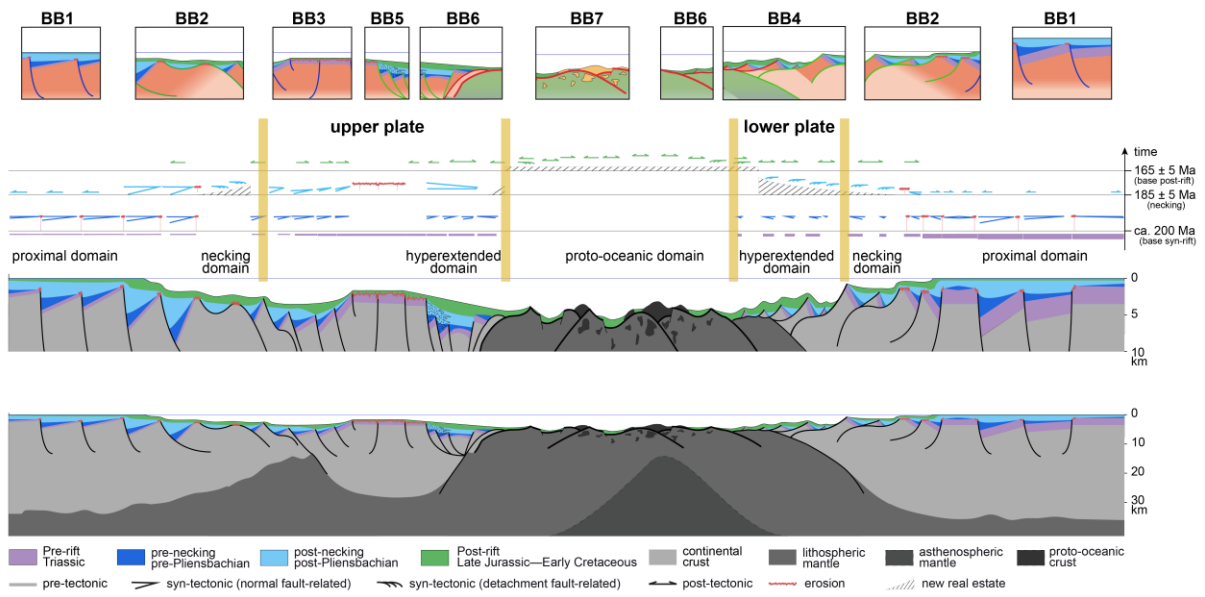
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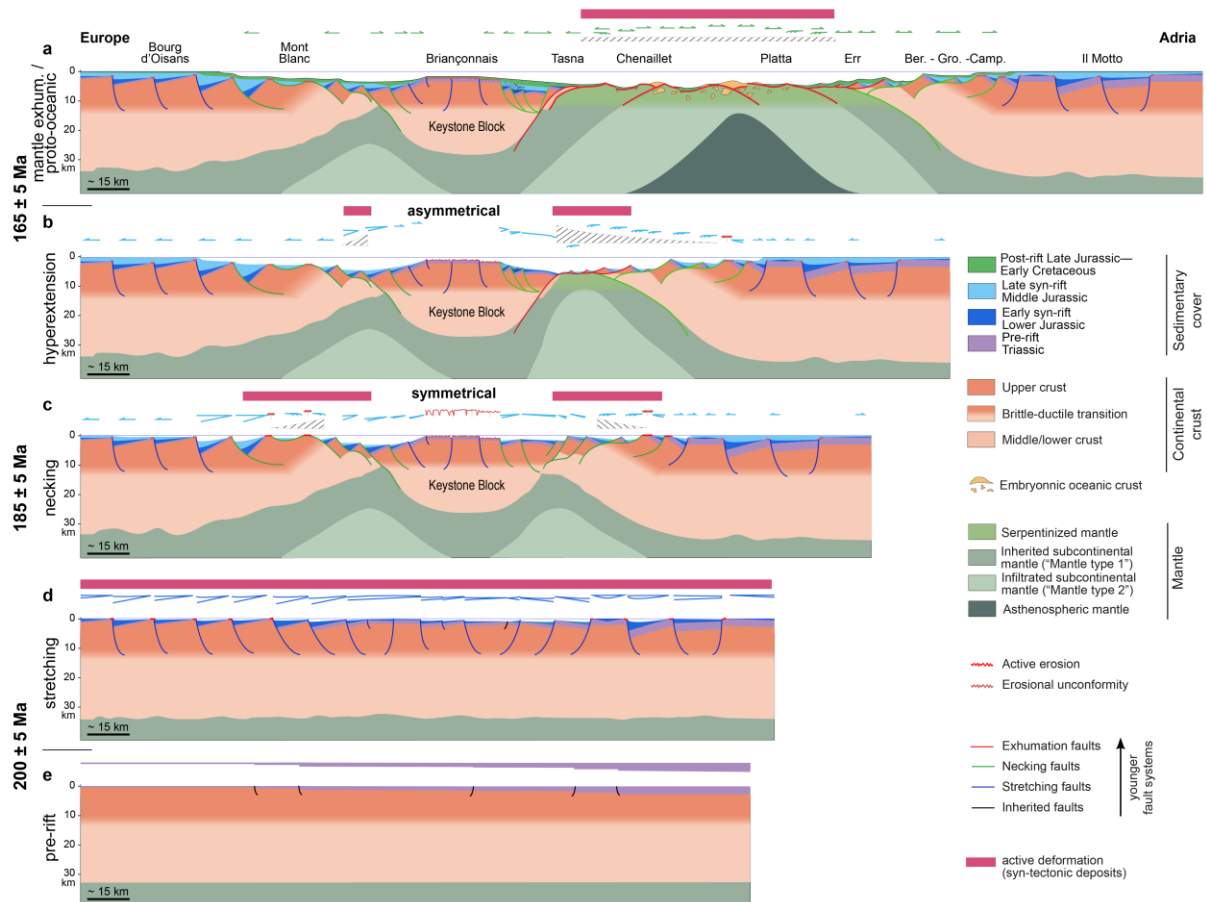
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**Fig. 8:** Examples of BB7 exposed in the Alps documenting early seafloor spreading (proto-oceanic crust) of the Alpine Tethys: (a) the Lower Platta Unit exposed in the Val Surses in SE Switzerland (modified after Epin et al., 2019); (b) Chanaillet Ophiolite exposed near Briançon along the French/Italian border in the Western Alps (modified after Manatschal et al., 2011); (c) section across the Ocean Continent Transition of the Alpine Tethys showing the transition from hyperextended crust to exhumed mantle and then to proto-oceanic crust (modified after Manatschal & Müntener, 2009). Section (c) shows localities in (a) and (b).



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**Fig. 9:** Section across the Alpine Tethys corresponding to transects introduced in Fig. 3 (for location see Figs. 1 and 4). The section highlights the stratigraphic architecture of the Alpine Tethys margins across the central Piemonte segment and document the syn-tectonic evolution in the overlying Wheeler diagram (for details see text).



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1300 **Fig. 10:** Main phases of the Alpine Tethys rift evolution: (a) mantle exhumation phase and  
 1301 embryonic seafloor spreading (formation of proto-oceanic crust); (b) hyperextension  
 1302 phase; (c) necking phase; (d) stretching phase; (e) initial stage. See text for a detailed  
 1303 description of the evolution.

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|  | BB1                                   | BB2                                      | BB3  | BB4  | BB5                                       | BB6                   | BB7   |
|--|---------------------------------------|--|--|--|---|-----------------------|---|
| <b>rift domain</b>                         | proximal (stretching)                 | necking                                  | necking (residual keystone)                    | hyperextended (lower plate)                    | hyperextended (upper plate)               | exhumed               | proto-oceanic                               |
| <b>type basement</b>                       | upper continental crust               | upper/middle continental crust           | upper continental crust                        | upper/lower continental crust                  | upper/lower continental crust             | exhumed mantle        | exhumed mantle/magma                        |
| <b>fault-types</b>                         | high-angle faults                     | high/low-angle faults                    | high-angle faults                              | high/low-angle faults                          | high/low-angle faults                     | high/low-angle faults | high/low-angle faults                       |
| <b>bathymetry (during formation)</b>       | shallow/subareal                      | shallow/subareal                         | shallow/subareal                               | deep/subsiding                                 | deep/subsiding                            | deep/subsiding        | deep/subsiding                              |
| <b>type locality (see Fig. 4d &amp; e)</b> | Il Motto / Generoso / Bourg d'Oisans) | Campo-Grosina/ Mont Blanc/Cusio-Biellese | Fond Froid – Lac de l'Ascension (Briançonnais) | Err & Bernina                                  | Breccia Nappe                             | Tasna OCT             | Lower Platta / Chenaillet                   |
| <b>key reference</b>                       | Eberli 1988<br>Ribes et al. 2019a     | Mohn et al. 2012<br>Ribes et al. 2020a   | Claudel & Dumont, 1999                         | Masini et al., 2012<br>Epin & Manatschal, 2018 | Escher et al. 1993<br>Steffen et al. 1993 | Ribes et al. 2020b    | Epin et al. 2019,<br>Manatschal et al. 2011 |

1306

1307 Table 1: Overview of Building Blocks defined for the Alpine Tethys

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