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The Mesozoic Iberia-Eurasia diffuse plate boundary: a wide domain of distributed transtensional deformation progressively focusing along the North Pyrenean Zone

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Abstract

Plate kinematic reconstructions available for the Late Jurassic-Early Cretaceous eastward drift and counterclockwise rotation of the Iberian plate imply a major left-lateral motion of Iberia with respect to Eurasia. According to most authors, this displacement has been accommodated

along the transform North Pyrenean Zone. However, no relevant field evidence exists for the proposed >400 km of horizontal displacement along the North Pyrenean Fault. Several Permian-Mesozoic basins are distributed around the Iberia/Eurasia plate boundary and have been more or less inverted during the Cenozoic Pyrenean Orogeny (i.e. Iberian Chain basins, North and South Pyrenean basins, Basque-Cantabrian Basin, Parentis Basin, Bay of Biscay/Asturian margins). All of these basins experienced a complex kinematic history and shared the same tectono-stratigraphic evolution, with two successive rifting stages: (i) Permian-Triassic rifting following the dismantling of the Variscan belt and recording the early breakup of Pangea and (ii) Late Jurassic-Early Cretaceous rifting developing after a Jurassic post-rift thermal cooling stage. Depending on the different techniques of investigation and on the interpretation of controversial datasets, authors proposed either opening by orthogonal rifting or by transtensional/pull-apart tectonics for these basins.

In this work, we propose a reappraisal of the processes responsible for the Mesozoic Iberia/Eurasia plate boundary compartmentalization by reviewing the tectono-sedimentary history and the kinematic evolution of the sedimentary basins involved in this domain. We shed light on the fact that the Cretaceous left-lateral movement within the plate boundary was not accommodated by localized deformation along the single North Pyrenean Fault wrench structure, but rather by a distributed zone of deformation in which the transtensional regime was recorded by the sedimentary basins therein. We also suggest that other Permian-Mesozoic depocenters located below the Cenozoic foreland basins of the Pyrenean belt (i.e. the Ebro and Aquitaine basins) may have been active segments of this rift system. We then propose that the real extent of the Mesozoic plate boundary is roughly defined by two NW-SE trending lineaments corresponding to the southwestern margin of the Iberian Chain, on the Iberian side, and to the southern Armorican margin and the southwestern border of the French Central Massif, on the Eurasian side. The complex pre-Cretaceous tectono-sedimentary history of this region determined its peculiar pre-rift structure. Such structural inheritance may have favored a distributed rather than a localized mode of

deformation at the Iberia/Eurasia diffuse plate boundary during the Late Jurassic-Early Cretaceous, whilst mechanisms related to the eastward propagation of the Bay of Biscay system might have been responsible for the final localization of the plate boundary along the Basque-Cantabrian/North Pyrenean corridor.

Key words

- Iberia/Eurasia plate boundary

- Diffuse plate boundary

- Plate kinematics

- Structural inheritance

- Polyphased rifting

- Orthogonal vs. oblique rifting

1. INTRODUCTION

Two of the major assumptions in plate tectonics theory are that tectonic plates are rigid undeformable blocks and that deformation localizes in narrow zones at their edges, known as "plate boundaries" (e.g. Wilson, 1965; McKenzie & Parker, 1967). Since the early formulation of this theory in the 1960s, these assumptions have been questioned by geological and geophysical constraints evidencing (i) that intraplate deformation can occur and (ii) that deformation can be accommodated by wide domains along plate edges known as "diffuse plate boundary zones" (e.g. Gordon & Stein, 1992; Gordon, 1998; Kreemer *et al*., 2003 and references therein). In fact, in the case where a large domain surrounding a plate boundary has experienced a coherent tectonic history, it can be reasonably regarded as a diffuse plate boundary zone. Such fossil analogues are of broad interest since diffuse plate boundaries cover a significant portion of Earth's surface (~15%; Gordon, 1998). Indeed, the identification of such broad deforming domains accommodating relative movement between tectonic plates may provide important constraints to plate kinematic models. This can also help to resolve controversies arising from the discrepancy regarding the amount of displacement postulated by plate kinematic models and the amount of deformation estimated from available geological observations.

The Mesozoic Iberia/Eurasia plate boundary is an excellent example to study the partitioning of deformation at plate boundaries and its implications on kinematic models. Indeed, the plate boundary has classically been identified within the narrow Pyrenean realm as the inverted North Pyrenean Zone (e.g. Le Pichon *et al*., 1970; Choukroune *et al*., 1973; Choukroune & Mattauer, 1978). However, it is worth noting that the Pyrenean domain is surrounded by numerous inverted Permian-Mesozoic rift basins, such as the Iberian Chain System (e.g. Alvaro *et al*., 1979) or the Parentis Basin (e.g. Pinet *et al*., 1987; Bois *et al*., 1990) (Fig. 1). On the Iberian side of the Iberia/Eurasia plate boundary, the tectonic history of these domains has been often considered as related to intraplate deformation (e.g. Salas *et al*., 2001, 2010; Guimerà *et al*., 2004; Guimerà,

2018). Nevertheless, the consistency in the tectonic evolution of these inverted basins, on both sides of the Iberia/Eurasia plate boundary, has seldom been taken into account (e.g. Tugend *et al*., 2015a; Lagabrielle *et al*., 2020; Angrand & Mouthereau, 2021).
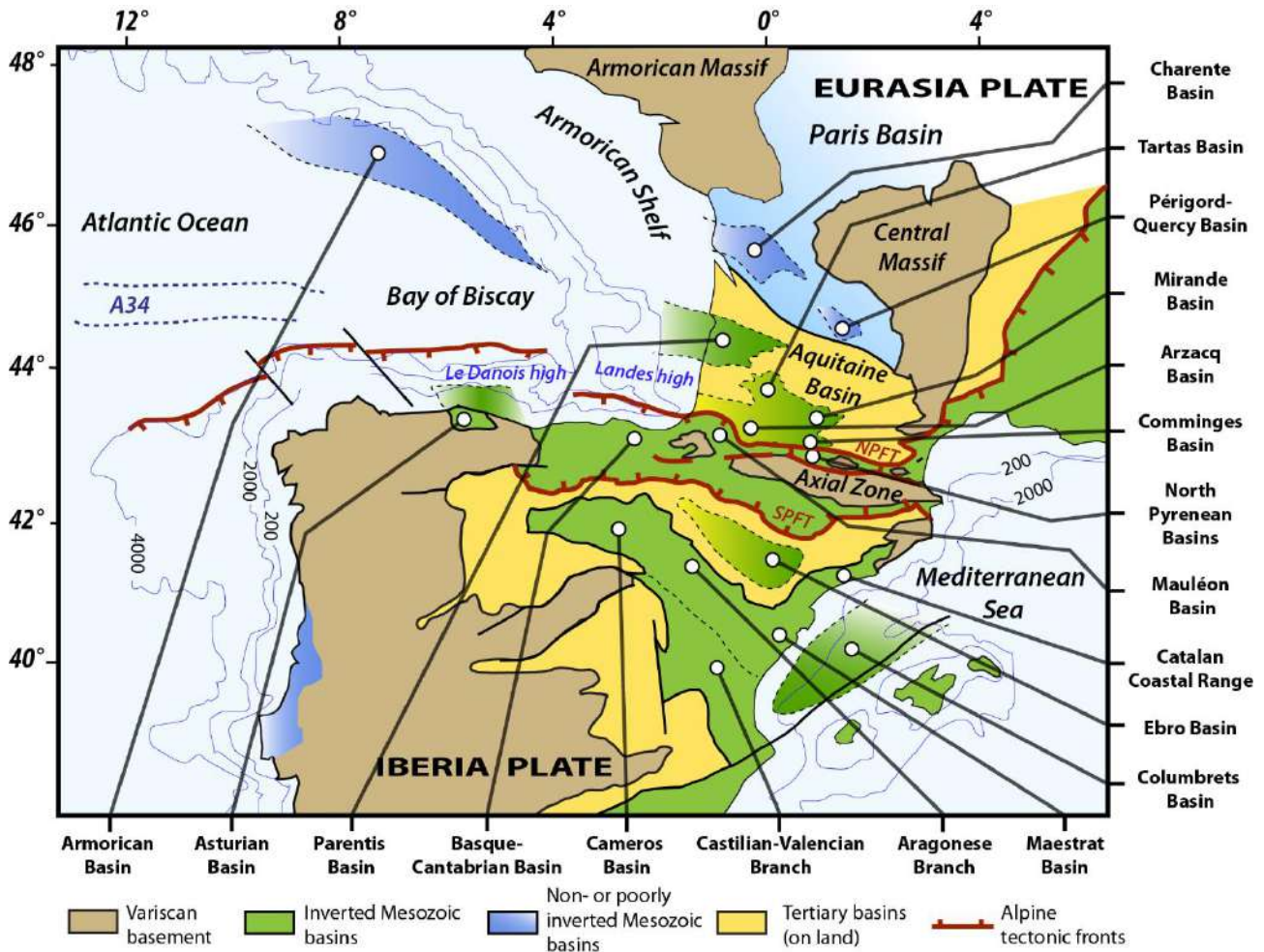
Fig. 1: Simplified structural map of the Iberia/Eurasia diffuse plate boundary zone, with locations of the basins presented in this work (modified after Lagabrielle et al., 2020). Acronyms: NPFT, North Pyrenean Frontal Thrust; SPFT, South Pyrenean Frontal Thrust.

Since early studies in the 1970s (e.g. Le Pichon *et al*., 1970; Le Pichon & Sibuet, 1971; Choukroune *et al*., 1973; Sibuet, 1974; Williams, 1975; Choukroune & Mattauer, 1978), the Late Jurassic-Early Cretaceous kinematic evolution of the Iberia/Eurasia plate boundary, which is strictly linked to the coeval northward propagation of the southern North Atlantic and the opening of the V-shaped Bay of Biscay margins, has been widely debated and remains an unsolved issue in the Western European tectonic framework. The main reasons for this long-lasting debate are the contrasting interpretations given to the M0 magnetic anomaly in the Bay of Biscay/North Atlantic

margins (e.g. Barnett-Moore *et al*., 2016 and references therein) and the fact that considerable amounts of displacement between the Iberian and Eurasia plates occurred during the ~40 Myr-long C34 magnetic chron (i.e. Aptian *pp*-Santonian) (Neres *et al.,* 2013). This context implies lots of uncertainties on the kinematics of the Iberian plate during the Late Jurassic-Early Cretaceous interval and has left space for very contrasting plate kinematic models in the last 50 years.

The main plate kinematic models that have been proposed up to now can be subdivided into three groups characterized by differences in chronology and kinematics, with models proposing (Fig. 2): i) an eastward displacement of Iberia during the Albian-Cenomanian accommodated by the opening of sinistral pull-apart (Choukroune & Mattauer, 1978) or transtensional basins (Olivet, 1996) in the ~E-W trending North Pyrenean Zone; ii) a left-lateral transcurrent displacement between Iberia and Eurasia accommodated along the ~E-W trending North Pyrenean Fault between the Late Jurassic and the Aptian, followed by Albian-Cenomanian orthogonal (N-S) extension (e.g. Jammes *et al*., 2009); iii) a scissor opening of the Bay of Biscay margins during the Albian, implying coeval convergence and subduction of several hundreds of kilometers of oceanic crust in the Pyrenean realm (e.g. Srivastava *et al*., 1990; Sibuet *et al*., 2004; Vissers & Meijer, 2012; Vissers *et al*., 2016). This latter group of models is generally considered to not be supported by geological constraintsdue to the substantial lack of geological evidence for subduction and convergence in the Pyrenean domain prior to the Late Cretaceous and to the fact that these models are based on the restoration of the J series magnetic anomalies. These anomalies have been recently demonstrated to not correspond to oceanic magnetic isochrons and are thus not suitable for plate kinematic reconstructions (Nirrengarten *et al*., 2017; Szameitat *et al.*, 2020). This implies that the most reliable kinematic models correspond to groups i and ii. Although these models agree on the fact that ~400 km of left-lateral displacement must have been accommodated between Iberia and Eurasia, the precise timing of this movement is not a matter of consensus (i.e. Albian-Cenomanian vs. Late Jurassic-Aptian).

Fig. 2: *Main plate kinematic models proposing a different Late Jurassic-Cretaceous evolution of the Iberia/Eurasia plate boundary. a) Model by Olivet (1996) proposing an Albian-Cenomanian eastward displacement of the Iberian plate accommodated along the North Pyrenean Fault (as redrawn by Mouthereau et a., 2014). b) Model by Jammes et al. (2009) proposing left-lateral displacement between Iberia and Eurasia accommodated along the North Pyrenean Fault between the Late Jurassic and the Aptian, followed by Albian-Cenomanian orthogonal (N-S) extension. c) Model by Srivastava et al. (2000) and Sibuet et al. (2004) proposing a scissor opening of the Bay of Biscay, implying subduction of a large oceanic domain during the Albian in the Pyrenean realm (as redrawn by Mouthereau et a., 2014).*

In the past, most of these models (i.e. groups i and ii) have postulated that the large-scale sinistral displacement of Iberia relative to Eurasia was accommodated along the North Pyrenean Fault transform system, which has thus been classically considered to represent the plate boundary between the two plates. However, geological evidence from the Pyrenees shows that the left-lateral displacement accommodated by this structure is one order of magnitude smaller (i.e. few tens of kilometers; e.g. Debroas, 1987, 1990, 1995; Canérot, 2016) than what is required by plate kinematic models. More recently, new models have proposed a distribution of the Late Jurassic-Early Cretaceous deformation between the two plates in two (Angrand *et al*., 2020; Angrand & Mouthereau, 2021; Frasca *et al*., 2021; King *et al*., 2021) or three (Tugend *et al*., 2015a; Nirrengarten *et al*., 2018) transtensional corridors located in the Iberian Chain Rift System, in the Pyrenees, and in the Parentis Basin (Fig. 3). This partitioning of the deformation in space and time along distinct tectonic corridors better conforms with geological observations (e.g., Angrand et al., 2020; Angrand & Mouthereau, 2021). However, the amount of left-lateral displacement remains larger than what can be inferred from geological observations. Due to the lack of reliable oceanic magnetic anomalies to constrain the Late Jurassic-Early Cretaceous kinematic evolution of the Iberia/Eurasia plate boundary, accurate compilation of the available regional geological constraints is key to resolving this long-lasting debate.

The large study region that includes the Bay of Biscay margins and the numerous surrounding basins, formed during Permian-Cretaceous times (see Section 2) between Iberia and Eurasia, recorded a complex polyphase tectonic history. This domain is characterized by large-scale fault systems, inherited from the Variscan Orogeny (e.g. Arthaud & Matte, 1975; Ziegler, 1988; Matte, 2001). These structures controlled the localization of a large number of rift basins that share the same tectonic history, from the Late Carboniferous-Early Permian collapse of the Variscan belt (e.g. Lago *et al*., 2004a, b; Saspiturry *et al*., 2019a), to the Late Permian-Triassic breakup of Pangea (e.g. Sopeña *et al*., 1988; Frizon de Lamotte *et al*., 2015), and finally the progressive opening of the North Atlantic-Bay of Biscay oceanic systems from the Jurassic to the Cretaceous (e.g. Vergés &

García-Senz, 2001). Considering the space/time tectonic evolution of this large system of basins as a whole may reveal important constraints on the polyphase evolution of the Iberia/Eurasia plate boundary.
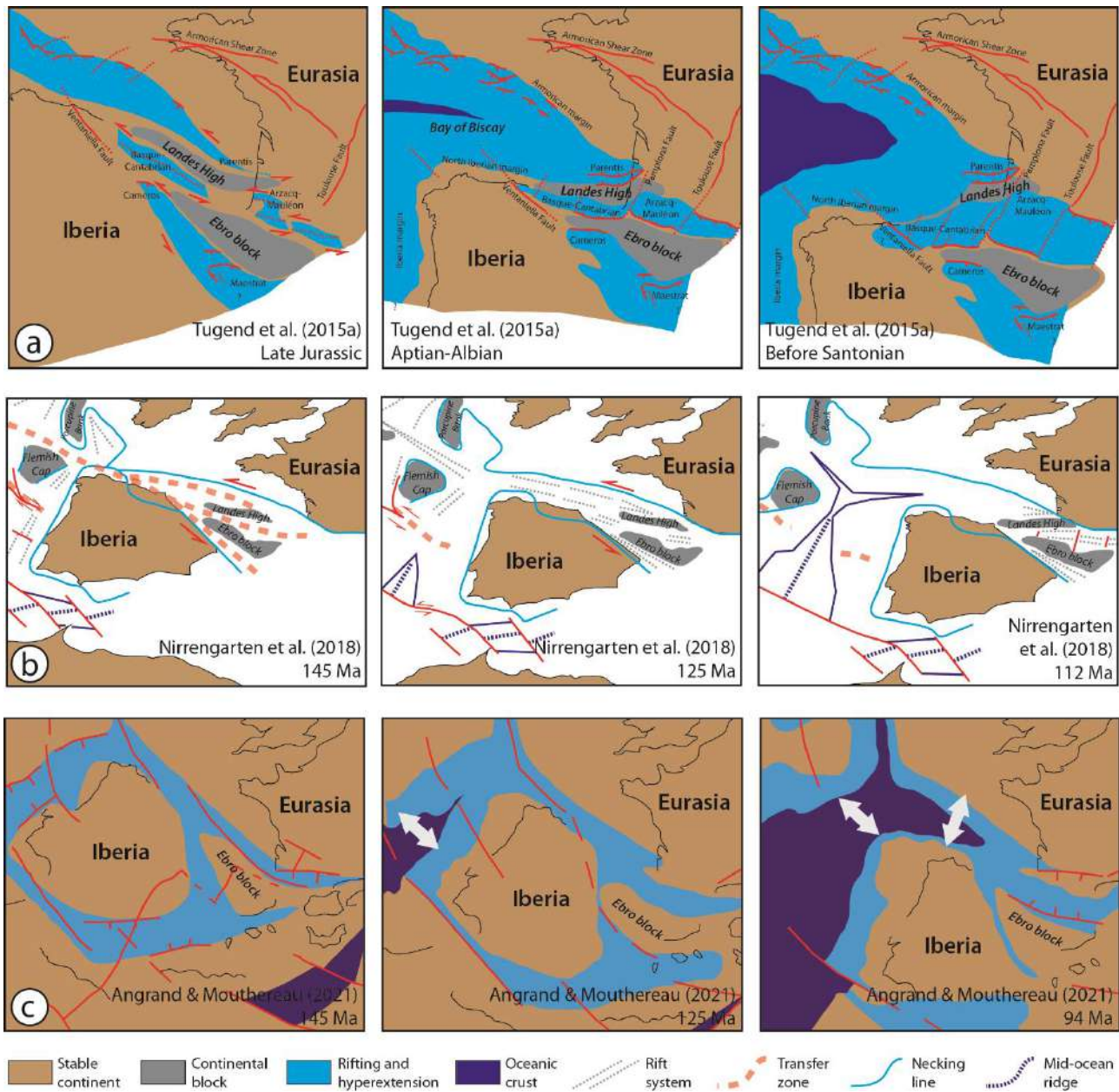
Fig. 3: *Recent plate kinematic models by (a) Tugend et al. (2015a), (b) Nirrengarten et al. (2018), and (c) Angrand & Mouthereau (2021) proposing a partitioning of the Late Jurassic-Early Cretaceous deformation between Iberia and Eurasia in two or three transtensional corridors located in the Iberian Chain Rift System, in the Pyrenees, and in the Bay of Biscay/Parentis system, separated by one or two intervening micro continental blocks (i.e. the Landes High and the Ebro Block).*

In this study, we propose a reappraisal of the Mesozoic tectonic and kinematic evolution of the sedimentary basins across the Iberia/Eurasia plate boundary (Fig. 1). These basins are: (1) the

Bay of Biscay margins, (2) the Asturian Basin, (3) the Iberian Chain basins, (4) the Pre-Ebro Basin, (5) the Basque-Cantabrian Basin, (6) the North Pyrenean Zone basins, (7) the Organyà Basin, (8) the Parentis Basin and (9) the Southern and Northern Aquitaine basins. This review allows for discussion of several key aspects of the Late Jurassic-Early Cretaceous history of this domain, such as the accommodation of the deformation over a large domain at the plate boundary and the parameters that favored distributed rather than localized deformation. This study also allows for inferences to be made about the timing of rotation of the Iberian plate during the Mesozoic and to highlight the contribution to the Mesozoic evolution of the plate boundary of some domains within the Cenozoic foreland basins of the Pyrenees that have been, up to now, disregarded.

2. REVIEW OF THE TECTONO-SEDIMENTARY EVOLUTION OF THE IBERIA/EURASIA PLATE BOUNDARY BASINS

The aim of this section is to review the tectono-stratigraphic evolution of the rift basins included in a ~400 km wide region that spans from the Iberian Chain basins (in the southwest) to the Northern Aquitaine Basin (in the northeast) (Fig. 1). This wide domain recorded: i) late-Variscan evolution (Late Carboniferous-Early Permian) as well as Pangea breakup (Late Permian to Triassic-Jurassic boundary); ii) the complex kinematic evolution of these basins during the Late Jurassic-Early Cretaceous rifting in relation to the opening of the V-shaped Bay of Biscay margins, the eastward drift of Iberia, and its coeval counterclockwise rotation. During this latter phase, the Keuper evaporites played an important role as an efficient decoupling layer at the base of the pre-rift sedimentary cover.

2.1. Bay of Biscay margins

The Bay of Biscay is a V-shaped E-W trending oceanic basin that opened during the Early Cretaceous between the non-volcanic Western Approaches/Armorican Margin (to the north) and the North Iberian Margin (to the south) (Figs. 1, 2 and 3). Its opening occurred in relation to the northward propagation of the North Atlantic Rift System (e.g. Montadert *et al*., 1974; 1979; Malod & Mauffret, 1990; Barnett-Moore *et al*., 2016 and references therein). This region is characterized by a strong Late Paleozoic tectonic inheritance, mainly consisting of lithospheric scale strike-slip shear zones (e.g. Ziegler, 1988; Burg *et al*., 1994; Matte, 2001; Bourrouilh, 2012 and references therein). The beginning of rifting is constrained to the Early Cretaceous (Berriasian-Late Aptian) (Williams, 1975; Montadert *et al*., 1979). Oceanic spreading started during the Late Aptian, as indicated by a Middle/Late Aptian breakup unconformity on the margins (Thinon *et al*., 2002; Roca *et al*., 2011; Cadenas & Fernández-Viejo, 2017), and ceased before Chron 33 (i.e. Santonian-Campanian boundary) (Williams, 1975). During this period, the northern and southern margins of the Bay of Biscay underwent post-rift thermal subsidence (Thinon *et al*., 2002; Cadenas & Fernández-Viejo, 2017). Despite general consensus that oceanic spreading occurred during magnetic Chron C34 (i.e. Late Aptian-Santonian) (e.g. Williams, 1975; Sibuet *et al*., 2004), the timing of lithospheric breakup and of the formation of the first oceanic crust in the Bay of Biscay domain have been widely debated over the last decades. The debate is mainly due to contrasting interpretations of the M-series magnetic anomalies (see Nirrengarten *et al*., 2017 for an exhaustive review), which in the Bay of Biscay system developed in the distal part of the ocean-continent transition, rather than in the true oceanic domain (Thinon *et al*., 2003; Tugend *et al*., 2014). Even though there is general agreement on the fact that the Late Jurassic-Early Cretaceous Bay of Biscay margins opened in the general context of left-lateral transtension, the relative chronology of the transtensional vs. orthogonal rifting phases and the spatial distribution of the deformation are still widely debated (e.g. Choukroune & Mattauer, 1978; Olivet, 1996; Jammes *et al*., 2009; Tugend *et al*., 2015a; Barnett-Moore *et al*., 2016; Nirrengarten *et al*., 2018; Peace *et al*., 2019; Angrand *et al*., 2020; Frasca *et al*., 2021; King *et al*., 2021).

2.1.1. The Northern Margin

The Armorican Basin, located at the foot of the Armorican continental slope (Fig. 1), developed during the opening of the Bay of Biscay margins and is filled with up to 7 km of sediments (Bacon *et al*., 1969; Avedik & Howard, 1979; Thinon *et al*., 2002, 2003). It lies on the extremely thinned (<4 km) continental crust (Avedik & Howard, 1979; Montadert *et al*., 1979; Le Pichon & Sibuet, 1981) of an ~80 km wide ocean-continent transition (Thinon *et al*., 2003), which also includes a domain of possibly exhumed sub-continental mantle (Tugend *et al*., 2014). In general, the Armorican Margin (to the east) seems to be characterized by a wider ocean-continent transition and by fewer tilted crustal blocks compared to the Western Approaches Margin (to the west) (Thinon *et al*., 2003). The pre-rift series consist of Late Jurassic (Kimmeridgian-Tithonian) platform carbonates (Thinon *et al*., 2002, 2003). The extensional history of the basin is marked by a rapid pre-Berriasian compartmentalization into horsts and grabens, followed by a significant Berriasian-Aptian subsidence phase which culminated in an abrupt Late Aptian collapse of the transitional domain (Thinon *et al*., 2002). Tectonic inversion related to the Pyrenean convergence phase is very limited on the Armorican side of the Bay of Biscay margins (Thinon *et al*., 2001).

2.1.2. The Southern Margin

The North Iberian Margin displays much evidence for significant tectonic inversion related to the Late Cretaceous-Cenozoic Pyrenean compression, with incipient southward subduction of the oceanic domain below the continental margin (Boillot *et al*., 1979; Malod *et al*., 1982; Pulgar *et al*., 1996; Alvarez-Marrón *et al*., 1997; Fernández-Viejo *et al*., 1998, 2011; Roca *et al*., 2011). In the eastern (Cantabrian) part of the North Iberian Margin continental shelf, the E-W trending structural high of the Landes Plateau separates the Late Jurassic-Early Cretaceous Parentis Basin (to the north) from the Basque-Cantabrian Basin (to the south) (e.g. Pinet *et al*., 1987; Roca *et al*., 2011) (Fig. 1). In the central (Asturian) part of the continental shelf, the E-W trending structural high of the Le Danois Bank separates the southern ocean-continent transition of the Bay of Biscay margins

(to the north) from the Late Jurassic-Early Cretaceous Le Danois Basin (to the south), which represents the offshore part of the Asturian Basin (see Section 2.2. Asturian Basin) (e.g. Pulgar *et al.*, 1996; Fernández-Viejo *et al.*, 1998; Roca *et al.*, 2011; Cadenas & Fernández-Viejo, 2017; Cadenas et al., 2018, 2020) (Figs. 1 and 4a). Apatite fission track ages support exhumation of the lower granulitic crust on the continental slope to the north of the Le Danois Bank during the Berriasian-Early Aptian interval (Fügenschuh *et al.*, 2003).

A notable difference between the northern and southern Bay of Biscay margins is the record of the Permian-Triassic rifting phase and its subsequent impact. Indeed, on the Armorican side of the Bay of Biscay system, the sedimentary record of this rifting event seems to be absent (Thinon *et al.*, 2002), which implies lack of deposition of the Late Triassic Keuper evaporites in this sector. By contrast, along the North Iberian Margin, the Permian-Triassic rifting is well developed as indicated by the deposition of thick Late Triassic evaporitic deposits at the base of the Jurassic post-rift series (e.g. Lepvrier & Martínez-García, 1990; Espina *et al.*, 2004; Cadenas & Fernández-Viejo, 2017; Zamora *et al.*, 2017 and references therein). The Keuper evaporites played a fundamental decoupling role during the Late Jurassic-Early Cretaceous evolution of the North Iberian Margin (e.g. Boillot *et al.*, 1979; Ferrer *et al.*, 2012, 2014; Cadenas & Fernández-Viejo, 2017; Zamora *et al.*, 2017; Lagabrielle *et al.*, 2020). This configuration is similar to the situation for the easterly Aquitaine Basin, where the Celtaquitaine Flexure, which aligns to the NW with the Armorican continental slope, marks the northern limit of the distribution of the Keuper evaporites in the region (e.g. BRGM *et al.*, 1974; see Section 2.8. Aquitaine Basin).

2.2. Asturian Basin

The Asturian Basin is an onshore/offshore partially inverted Permian-Mesozoic basin localized on the northern Iberian margin, at the transition between the continental Iberian plate and the oceanic Bay of Biscay system (Lepvrier and Martínez-García, 1990; Cadenas & Fernández-

Viejo, 2017) (Figs. 1 and 4a). In the literature, the offshore part of this basin is also referred to as the Le Danois Basin (e.g. Roca *et al*., 2011) (Fig. 4a). This basin experienced a polyphase rifting history, with an initial Permian-Triassic rifting phase related to Pangea breakup, and a second Middle(?)/Late Jurassic-Early Cretaceous rifting phase resulting from the opening of the Bay of Biscay margins (López-Gómez *et al*., 2019a and references therein). The maximum thickness of its sedimentary fill may reach 10 km (Cadenas & Fernández-Viejo, 2017; López-Gómez *et al*., 2019a). The Upper Triassic sediments consist of Keuper facies evaporitic deposits (Suárez-Rodríguez, 1988; Pieren *et al*., 1995; Cadenas & Fernández-Viejo, 2017). Early/Middle Jurassic carbonate sediments are transgressive over the Permian-Triassic deposits and are overlain by Late Jurassic (Kimmeridgian) syn-rift siliciclastic sequences (Valenzuela *et al*., 1986; Cadenas & Fernández-Viejo, 2017; García-Senz *et al*., 2019a). After a phase of erosion/non deposition between the latest Tithonian and the earliest Barremian (at least in the onshore part of the basin), these deposits are covered by a Barremian to Santonian Cretaceous succession (e.g. González-Fernández *et al*., 2004, 2014). Aptian post-rift limestones unconformably overlie both the Paleozoic and the Jurassic successions on blocks separated by NE-SW trending faults that previously controlled uplift and erosion (Lepvrier and Martínez-García, 1990; González-Fernández *et al*., 2004, 2014; Cadenas & Fernández-Viejo, 2017). However, García-Senz *et al*. (2019a) proposed that the Aptian-Albian series should be considered as syn-rift deposits due to significant thickness variations in these deposits and by analogy with the successions of the Basque-Cantabrian Basin. In the offshore part of the basin, the Neocomian phase of erosion/non deposition is not reported by Cadenas *et al*. (2018, 2020), who rather propose a unique Late Jurassic to Barremian rifting phase. The sag-shaped geometry of the syn-rift infill evidenced by seismic reflection profiles (e.g. Cadenas & Fernández-Viejo, 2017; Cadenas *et al*. 2018; Martín-Chivelet *et al*. 2019) might be related to the presence of a salt decoupling layer at the base of the sedimentary cover, as in other salt-controlled smooth-slope basins of the region (see review in Lagabrielle *et al*., 2020). In addition, Keuper-related syn-rift

halokinetic features have been recognized elsewhere in the basin (Boillot *et al*., 1979; Cadenas & Fernández-Viejo, 2017).

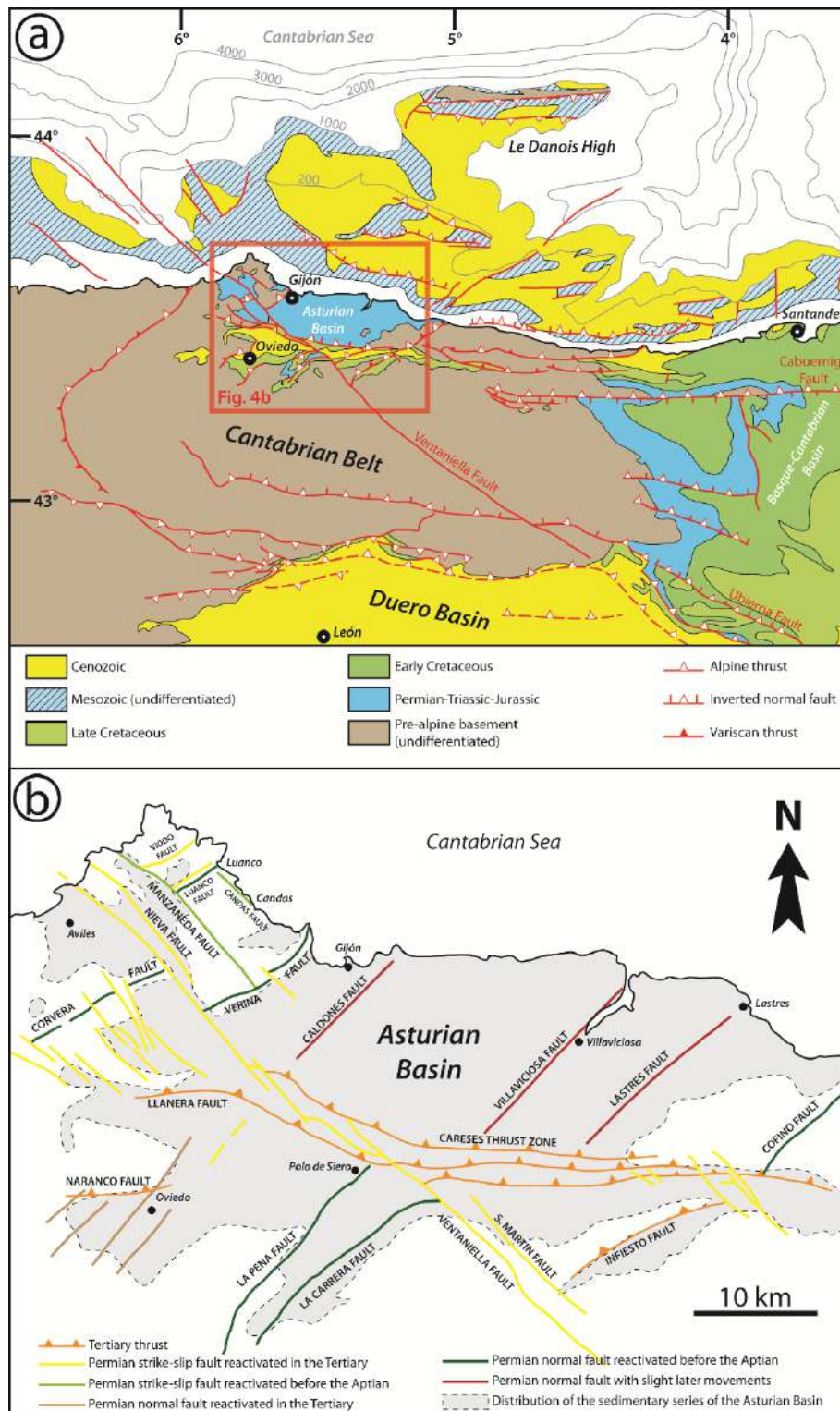**Fig. 4:** *Geological map and structural pattern of the Asturian Basin. a) Geological map of the Asturian Basin (including the offshore Le Danois sub-basin) and of the surrounding Cantabrian region (redrawn and modified after Pulgar et al., 1999). b) Simplified structural map of the onshore Asturian Basin with age of (re)activation of the main structures (redrawn and modified after Lepvrier & Martínez-García, 1990).*

The Meso-Cenozoic fault pattern of the basin is inherited from the late Variscan (Permian) period (Lepvrier and Martínez-García, 1990) and includes three sets of fault groups oriented NW-SE, NE-SW and E-W (Juliver *et al.*, 1971; Lepvrier and Martínez-García, 1990; Uzkeda *et al.*, 2016; Granado *et al.*, 2018) (Fig. 4b). Even though the Mesozoic directions of extension are difficult to estimate, the Late Jurassic-Aptian reactivation of NW-SE trending Permian strike-slip faults and NE-SW trending Permian normal faults (Fig. 4b; Lepvrier and Martínez-García, 1990) may suggest a NW-SE direction of stretching for the Late Jurassic-Early Cretaceous rifting phase. This interpretation is consistent with structural studies showing the influence of NE-SW trending normal faults on the Late Jurassic syn-rift sedimentay profile (Uzkeda *et al.*, 2013, 2016). However, these same studies document the coeval Late Jurassic activity of E-W and NW-SE trending sets of normal faults (Uzkeda *et al.*, 2016), thus revealing complex syn-rift kinematics. In this context, Granado *et al.* (2018) proposed a NNE-SSW direction of extension based on interpretation of joint set directions. This is consistent with the preponderance of E-W trending normal faults controlling E-W oriented depocentres during rifting in the offshore part of the Asturian Basin, as evidenced by Cadenas & Fernández-Viejo (2017). Moreover, recent studies on the offshore part of the basin proposed a tectonic regime of left-lateral transtension for the Late Jurassic to Barremian rifting phase (Cadenas *et al.*, 2020).

2.3. The Iberian Chain basins system

The Iberian Chain is a NW-SE trending intraplate belt, which results from the Late Cretaceous-Cenozoic inversion of former Permian-Mesozoic sedimentary basins (e.g. Alvaro *et al.*, 1979; Guimerà & Alvaro, 1990; Casas & Faccenna, 2001; Salas *et al.*, 2001; Guimerà, 2018 and references therein) (Figs. 1 and 5a). It is composed of two sub-parallel sedimentary troughs separated by the Cenozoic Almazan Basin and that merge in the SE in the Teruel region (Fig. 5a). This latter area corresponds to a significant gravimetric minimum indicating a higher crustal

thickness (Salas *et al*., 2001). The southwesterly Castillan-Valencian Branch terminates with the southeasterly South Iberian Basin (and possibly the offshore Columbrets Basin). The northeasterly Aragonese Branch includes the northwesterly Cameros Basin, the Central Iberian Basin and the southeasterly Maestrat (or Maestrazgo) Basin (e.g. Capote *et al*., 2002; Martín-Chivelet *et al*., 2019). To first order, all these basins experienced the same tectono-stratigraphic evolution, with an initial Permian-Triassic rifting and a subsequent Late Jurassic-Early Cretaceous rifting (e.g. Alvaro *et al*., 1979; Salas & Casas, 1993; Van Wees *et al*., 1998; Salas *et al*., 2001; Martín-Chivelet *et al*., 2019). In the following section, we will describe separately the Cameros Basin and the rest of the Iberian Chain basins.

Fig. 5: a) Simplified geological map of the Iberian Chain System with locations of the main Permian-Mesozoic rift basins (modified after Guimerà, 2018); isolines in the Ebro Basin refer to base of the Cenozoic series. b) Evolution of extension directions (gray arrows) between the Late Jurassic and the Albian in the Cameros and in the Basque-Cantabrian basins as retrieved by anisotropy of magnetic susceptibility (AMS) analysis (modified after Soto et al., 2008). c) Backstripped tectonic subsidence curves for the southeastern part of the Iberian Chain System (Maestrat and South Iberian basins); see (a) for location of the wells (redrawn and modified after Salas et al., 2001). Note that during the Late Jurassic-Early

Cretaceous rifting, tectonic subsidence curves are steeper for the Early Cretaceous interval than for the Late Jurassic one.

2.3.1. Cameros Basin

The Cameros Basin is an inverted sedimentary basin settled in the northwestern part of the Iberian Chain, which experienced polyphased rifting between the Permian and the Early Cretaceous (e.g. Salas & Casas, 1993; Salas *et al*., 2001). It consists of a ~120 km long and ~80 km wide WNW-ESE trending sedimentary basin, which was compartmentalized during the Late Jurassic-Early Cretaceous rifting stage that affected the Iberian Basin Rift System (Platt, 1990; Martín-Chivelet *et al*., 2019 and references therein) (Fig. 5a). The basin experienced an initial Permian-Triassic rifting episode (Alvaro *et al*., 1979; Salas *et al*., 2001), which culminated with the deposition of a ~500 m thick series of Keuper evaporitic deposits that decouple the overlying Mesozoic cover from its basement (Guiraud and Séguret, 1985; Casas-Sainz, 1993; Casas-Sainz and Gil-Imaz, 1998; Casas *et al*., 2009; Rat *et al*., 2019). This stage was followed by a Jurassic phase of thermal subsidence, which resulted in the deposition of unconformable carbonate platform sequences reaching up to ~800 m in thickness (Platt, 1990; Aurell and Meléndez, 1993; Aurell *et al*., 2003). A second rifting phase took place in the Late Jurassic (Tithonian)-Early Cretaceous (Early Albian) with the deposition of a thick syn-rift sedimentary pile reaching a maximum thickness of ~6500 m to 9000 m (Platt, 1990; Mas *et al*., 1993, 2011; Casas-Sainz and Gil-Imaz, 1998; Salas *et al*., 2001; Omodeo Salè *et al*., 2014; García-Lasanta *et al*., 2017; Martín-Chivelet *et al*., 2019; Rat *et al*., 2019). This phase includes a Late Berriasian–Early Barremian interval of reduced subsidence (Omodeo-Salé *et al*., 2017). The Late Jurassic-Early Cretaceous Cameros Basin has been interpreted by several authors as a ramp-syncline basin developed over a buried S-dipping detachment fault resulting from regional N-S to NNE-SSW extension (Mas *et al*., 1993, 2011; Guimerà *et al*., 1995; Salas *et al*., 2001; Omodeo-Salé *et al*., 2014). Recent interpretations have proposed that the syncline shape of the basin was accentuated by homogeneous ductile thinning of the basement below the Keuper evaporites décollement layer (Lagabrielle *et al*., 2020; Saspiturry *et*

al., 2021). The lower portion of the sedimentary fill of the basin experienced HT/LP metamorphism due to severe lithosphere thinning and significant sedimentary burial during the Early Cretaceous rifting (e.g. Guiraud & Séguret, 1985; Golberg *et al*., 1988; Mata *et al*., 2001; Rat *et al*., 2019; Saspiturry *et al*., 2021).

Field studies have shown that relevant thickness variations in the Late Jurassic-Early Cretaceous sequences are controlled by both NW-SE and NE-SW trending faults (Platt, 1990). Despite important thickness variations observed across NE-SW trending faults, Platt (1990) interpreted this set of faults as transverse faults (*sensu* Gibbs, 1984) and the NW-SE trending faults as normal faults. Even though most authors agree that the main direction of extension during the Late Jurassic-Early Cretaceous rifting is ~NE-SW, anisotropy of magnetic susceptibility (AMS) studies highlighted a phase of NW-SE stretching recorded in the basin between the Barremian and the Aptian (Soto *et al*., 2008; García-Lasanta *et al*., 2014) (Fig. 5b). This phase seems to coincide also with an increase in tectonic subsidence (e.g. Mas *et al*., 1993; Omodeo-Salé *et al*., 2017; Angrand *et al*., 2020). These observations are consistent with K-Ar dating, documenting an Early Cretaceous phase of left-lateral transtensional displacement along the Río Grío Fault, which bounds the northeastern margin of the transition between the Cameros Basin and the Central Iberian Chain Basin and is part of the larger North Iberian Fault system (Aldega *et al*., 2019).

2.3.2. The Central and Southern part of the Iberian Chain basins

This part of the Iberian Chain is composed of several inverted Permian-Mesozoic rift basins, which span from the Cameros Basin (in the NW) to the Mediterranean coastline of the Valencia trough (Figs. 1 and 5a). During the Permian-Triassic rifting, subsidence was controlled by late- to post-Variscan normal and sinistral strike-slip faults (e.g. Alvaro *et al*., 1979; Arche & López-Gómez, 1996; Antolín-Tomás *et al*., 2007). This phase culminated with the deposition of thick Keuper evaporitic series (e.g. Canérot, 1991; Arche & López-Gómez, 1996; Salas *et al*., 2001; Ortí *et al*., 2017; Vergés *et al*., 2020) and was followed by an Early/Middle Jurassic phase of post-rift

thermal subsidence resulting in the formation of stable carbonate platforms (e.g. Canérot, 1991; Aurell *et al*., 2003, 2019 and references therein).

Rifting reactivation started during the latest Oxfordian and lasted until the Early Albian (Salas and Casas, 1993; Van Wees *et al*., 1998; Salas *et al*., 2001; Martín-Chivelet *et al*., 2019). This Late Jurassic-Early Cretaceous rifting phase includes two main subsidence phases (latest Oxfordian-Berriasian and Late Hauterivian-Early Albian) separated by a Valanginian-Early Hauterivian phase of reduced subsidence (e.g. Salas and Casas, 1993; Salas *et al*., 2001), whose base is marked by an angular unconformity (Martín-Chivelet *et al*., 2019 and references therein). However, these two phases are not exactly synchronous at the scale of the whole Iberian Chain system. For example, rifting started in the latest Oxfordian in the Maestrat Basin and in the Early Tithonian in the southern part of the Castillan-Valencian Branch (Salas & Casas, 1993; Salas *et al*., 2001). Moreover, the Aragonese Branch recorded a more continuous sedimentary history than the Castillan-Valencian Branch, which is characterized by non-deposition during the Valanginian-Early Hauterivian phase (Martín-Chivelet *et al*., 2019). It is worth noting that in each of the two branches the thickness of the Late Jurassic-Early Cretaceous deposits increases towards the southeast (Martín-Chivelet *et al*., 2019 and references therein).

The structural evolution of the Late Jurassic-Early Cretaceous rifting phase of the Iberian Chain basins is rather complex. It is mainly controlled by NW-SE and NE-SW trending faults (e.g. the High Tagus and Rio Grio faults; Roca *et al*., 1994; Antolín-Tomás *et al*., 2007; Liesa, 2011; Aldega *et al*., 2019; Aurell *et al*., 2019; Martín-Chivelet *et al*., 2019), which were also strongly conditioned by pre-existing structures (e.g. Alvaro *et al*., 1979; Roca *et al*., 1994; Salas *et al*., 2001). Faults and fractures with a NW-SE orientation dominate in the western part of the Maestrat Basin while E-W and NE-SW directions, respectively, prevail in the central and eastern portions of the basin. This pattern has been related to radial extension (i.e. vertical $\sigma_1 \gg \sigma_2 \sim \sigma_3$) during the Early Cretaceous (Aranda & Simón, 1993; Antolín-Tomás *et al*., 2007; Martín-Chivelet *et al*., 2019

and references therein). Accordingly, magnetic fabric distribution reflecting the extension directions during the Early Cretaceous reveals two main perpendicular stretching directions, with mainly NE-SW extension directions characteristic of the northwestern part of the basin, while NW-SE directions of stretching prevail in the southeastern domain (García-Lasanta *et al.*, 2015). More generally, the southeastern termination of the Iberian Chain basins system shows an overall NW-SE direction of extension during the Late Jurassic-Early Cretaceous rifting, as indicated by extensional fault patterns and associated salt tectonics (mainly trending NE-SW) in the southeastern Maestrat Basin (Roca *et al.*, 1994; Vergés *et al.*, 2020) and by the opening direction of the Columbrets Basin in the Valencia trough (Etheve *et al.*, 2018). The Columbrets Basin consists of an 8-9 km thick Mesozoic sequence that was initiatied in the Triassic (Roca, 1994) and experienced hyper-extension during the Albian (Etheve *et al.*, 2018). This same Early Cretaceous extensional pattern is also reported for the Garraf Basin of the Catalan Coastal Range (to the east of the Ebro Basin), which also experienced the same Permian-Mesozoic tectonic evolution (Anadón *et al.*, 1979; Salas *et al.*, 2001; López-Gómez *et al.*, 2019b; Martín-Chivelet *et al.*, 2019). From a kinematic perspective, Martín-Chivelet *et al.* (2019) suggest that the distribution of accelerated subsidence in small depocenters during the Late Hauterivian-Early Albian phase could be explained by strike-slip transtensional tectonics related to left-lateral displacement between Iberia and Eurasia.

2.4. Pre-Ebro Basin

The Ebro Basin represents the pro-foreland basin of the Pyrenean Belt that developed during the Late Cretaceous-Cenozoic inversion of the Pyrenean rift system (e.g. Puigdefàbregas *et al.*, 1986; Gaspar-Escribano *et al.*, 2001; Vergés *et al.*, 2002). It is delimited to the north by the South Pyrenean Frontal Thrust, to the east by the Catalan Coastal Range and to the south by the Iberian Chain (Figs. 1 and 5a). As with most of the domains included between the Iberian Chain and the Aquitaine Basin, the Pre-Ebro Basin recorded a phase of Permian-Early Lias rifting which

culminated in the deposition of thick Keuper evaporites (Deselgaux & Moretti, 1988; Arche & López-Gómez, 1996; Arche *et al*., 2004, 2007; Vargas *et al*., 2009; De la Horra *et al*., 2013; Ortí *et al*., 2017). During this phase, a structural high, characterized by no sedimentation, was probably located in the northeastern sector of the present-day Ebro Basin. This high has been previously referred to in the literature as the Ebro High (e.g. Vergés *et al*., 1998; Arche *et al*., 2004; Vargas *et al*., 2009; Ortí *et al*., 2017). In general, the Permian-Early Lias rifting was strongly influenced by structures inherited from the Variscan period (e.g. Arche & López-Gómez, 1996; Arche *et al*., 2004). Since the Middle Lias, the Pre-Ebro Basin was characterized by post-rift thermal subsidence that allowed the development of stable Jurassic carbonate platforms (Deselgaux & Moretti, 1988; Filleaudeau, 2011). Traces of the Late Jurassic-Early Cretaceous rifting phase are hardly perceptible in the present-day Ebro Basin. Indeed, these have been overprinted by the Pyrenean compression and the post-compressive sedimentary fill (Fig. 5a). This led most authors to suggest that the Late Jurassic-Early Cretaceous rifting did not affect this domain (e.g. Deselgaux & Moretti, 1988; Gaspar-Escribano *et al*., 2001). Such an interpretation implies that the Ebro realm behaved as an undeforming continental block between the Iberian Chain and the Pyrenean rift corridors during the opening of the Bay of Biscay margins (e.g. Gaspar-Escribano *et al*., 2001; Tugend *et al*., 2015a; Nirrengarten *et al*., 2018; Angrand *et al*., 2020; Lagabrielle *et al*., 2020; Angrand & Mouthereau, 2021; Frasca *et al*., 2021; King *et al*., 2021) (Fig. 3). However, there are observations consistent with the occurence of a Late Jurassic-Early Cretaceous rifting phase in this domain. Indeed, tectonic subsidence analyses performed on the Bujaraloz-1 well (located ~50 km to the ESE of the town of Zaragoza; Fig. 6a), show two intervals of increasing subsidence rates (Vargas *et al*., 2009; Fig. 6b). The first subsidence phase is recorded during the Kimmeridgian-Tithonian(?), followed by a second and more pronounced subsidence phase during the Barremian-Aptian(?) (Vargas *et al*., 2009; Fig. 6b). A map of the Mesozoic substratum of the Cenozoic deposits of the Ebro Basin, constructed by interpolating well log data, shows that Early Cretaceous sediments are locally preserved below the foreland basin cover (Fig. 6a), particularly to the east of Zaragoza and in the northwestern corner of

the basin between the Basque-Cantabrian and Cameros basins (Filleaudeau, 2011). Finally, modeling of gravimetric data suggests the presence of up to ~4000 m of Mesozoic sediments preserved in the northwestern part of the Ebro Basin below the Cenozoic cover (Wehr *et al*., 2018; see their Fig. 14a), even though this interpretation contrasts with interpretation of seismic data proposing the presence of only few hundred meters of Mesozoic sediments (Casas-Sainz *et al*., 2000). These reconstructions suggest significant E-W variations in the distribution (preservation?) of the Mesozoic sediments. During the Late Cretaceous (Campanian?), in response to the onset of Iberia/Eurasia N-S convergence, the Pre-Ebro Basin realm underwent emersion and erosion before the development of the Cenozoic foreland basin sedimentation (e.g. Puigdefàbregas & Souquet, 1986; Gaspar-Escribano *et al*., 2001). This evolution is consistent with a tectonic context of syn-convergence flexural bulging of the subducting plate. This is probably one of the main reasons why evidence of the Late Jurassic-Early Cretaceous rifting phase is so poorly preserved below the Cenozoic cover.
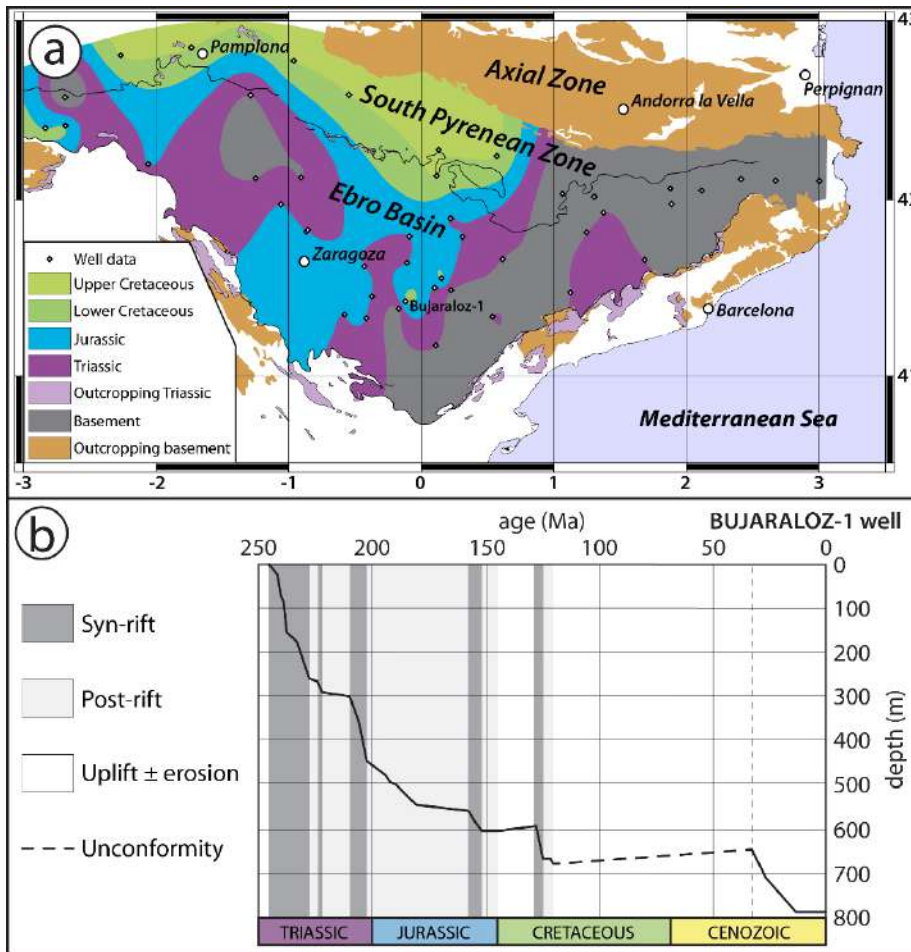
Fig. 6: *a) Age map of the sedimentary series underlain below the Cenozoic cover of the Ebro Basin as derived from interpolation of well log data (modified after Filleaudeau, 2011). b) Tectonic subsidence curve for the Bujaraloz-1 well (see location in (a); modified after Vargas et al., 2009).*

2.5. Basque-Cantabrian Basin

The Basque-Cantabrian Basin is a mildly inverted Permian-Mesozoic basin localized at the transition between the Basque Pyrenees of France (Mauléon Basin) and the Cantabrian Range (Rat, 1988; García-Mondéjar *et al*., 1996; Carola *et al*., 2013; Pedrera *et al*., 2017; Teixell *et al*., 2018) (Figs. 1 and 7). After an initial Permian-Triassic rifting phase, which included the deposition of a thick pile of evaporitic Keuper sequences (Rat, 1988; Espina *et al*., 2004; Serrano and Martínez del Olmo, 2004), a subsequent Late Jurassic (Tithonian)-Early Cretaceous rifting phase took place (Rat, 1988; García-Mondéjar *et al*., 1996; 2004; Cámara, 2017; Pedrera *et al*., 2017 and references therein). The thickness of the sedimentary fill in the basin may reach 12,5 km (Rat, 1988; García-

Mondéjar *et al*., 1996; 2004). The latter extensional phase also included a Valanginian-Barremian generalized regression (e.g. Cámara, 2020 and references therein), and eventually resulted in severe crustal thinning leading to subcontinental mantle exhumation below the Cretaceous basin during the Late Albian-Early Cenomanian interval (Mendia and Gil-Ibarguchi, 1991; Roca *et al*., 2011; DeFelipe *et al*., 2017; Pedrera *et al*., 2017). During this phase, pre- and syn-rift sedimentary series exposed in the eastern part of the basin (Nappe des Marbres Unit) recorded HT-LP metamorphism in relationship with mantle exhumation (e.g. Ducoux *et al*., 2019 and references therein) (Fig. 7). The crucial role of the decoupling level represented by the Keuper evaporites during the Late Jurassic-Early Cretaceous rifting has been evidenced by several studies in the region (e.g. Serrano and Martínez del Olmo, 2004; Tavani *et al*., 2013; Cámara, 2017; DeFelipe *et al*., 2017; Pedrera *et al*., 2017; Bodego *et al*., 2018; Lescoutre, 2019; Cámara, 2020; Lagabrielle *et al*., 2020; Lescoutre & Manatschal, 2020).
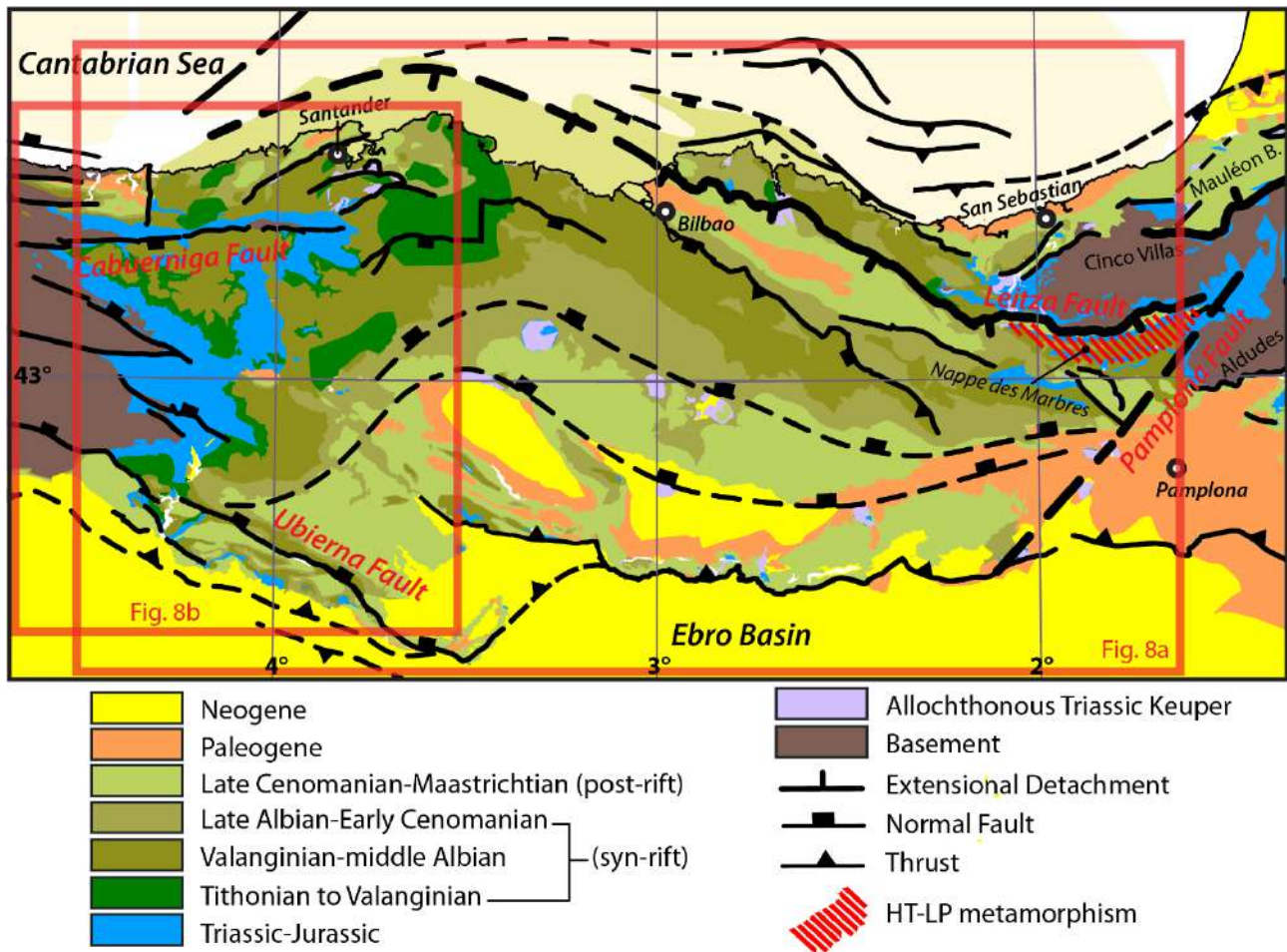
Fig. 7: *Geological map of the Basque-Cantabrian Basin with locations of the structural maps presented in Fig. 8 outlined in red boxes (modified after Pedrera et al., 2017).*

From a kinematic perspective, a lateral component of displacement has been proposed during this last extensional event (Rat, 1988; García-Mondéjar, 1989; García-Mondéjar *et al*., 1996 and references therein). Even though a complex fault pattern was recorded during the Aptian-Albian (Urgonian) rifting phase, which was potentially related to a strong Permian-Triassic inheritance (García-Mondéjar, 1989; García-Mondéjar *et al*., 1996 and references therein), evidence for sinistral strike-slip activity on NW-SE trending faults suggests a transtensional regime during this phase rather than a purely orthogonal rifting mode (García-Mondéjar, 1989; García-Mondéjar *et al*., 1996; Agirrezabala & Dinares-Turell, 2013; Cámara, 2017) (Fig. 8). Accordingly, meso-structural data (Soto *et al*., 2007; Tavani & Muñoz, 2012) and AMS studies (Soto *et al*., 2007, 2008; Oliva-Urcia *et al*., 2013) highlight a complex pattern of extension directions during the Late Jurassic-Early Cretaceous, with dominant NE-SW and subordinate NW-SE directions of extension. This

complex pattern of extension directions may have resulted from variations in the orientation of the stress regime during rifting, with a pre-Aptian NE-SW direction of extension (related to orthogonal rifting) and a NW-SE direction of stretching during Aptian/Albian times (Soto *et al*., 2007 and references therein). This scenario is consistent with minor mid-Cretaceous (Aptian/Albian) inversion on N-S trending structures in relation to sinistral strike-slip movement along en-echelon NW-SE oriented faults in the western Basque-Cantabrian Basin (Soto *et al*., 2011) (Fig. 8b). In contrast, other authors have argued for a strictly NE-SW orthogonal direction of extension associated with folding of the supra-salt sedimentary cover during the entire Late Jurassic-Albian rifting stage, based on structural analyses (e.g. Tavani & Muñoz, 2012; Tavani *et al*., 2013). Accordingly, the same authors attribute the sinistral strike-slip movement recorded along NW-SE structures to the Late Cretaceous-Cenozoic convergence (e.g. Tavani, 2012; Tavani *et al*., 2013). However, this interpretation disagrees with paleomagnetic studies showing a Late Albian anticlockwise rotation of syn-rift deposits in the northern part of the Basque-Cantabrian Basin, with that rotation interpreted as being related to ESE-trending sinistral wrench faulting (Agirrezabala & Dinares-Turell, 2013).
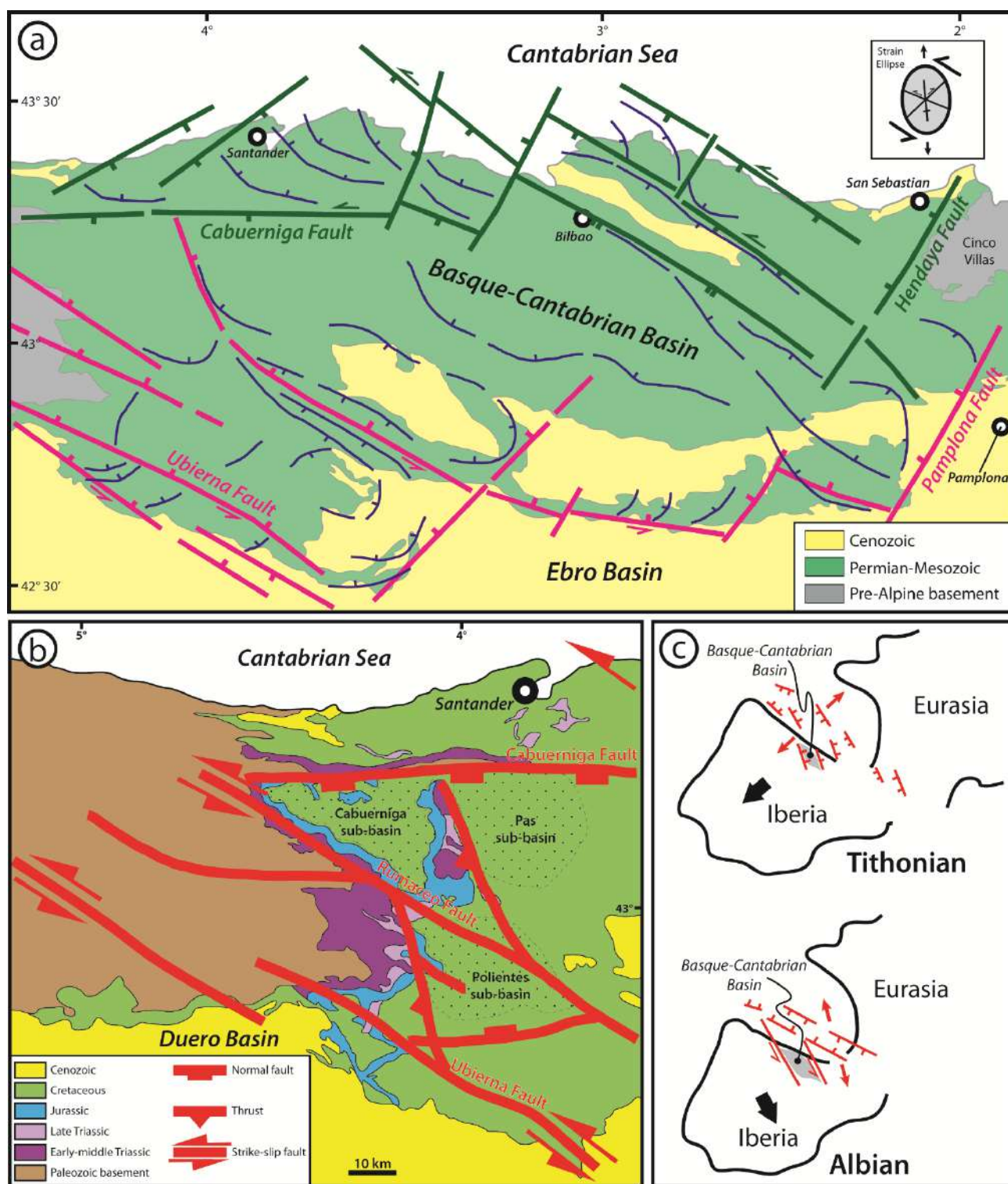
Fig. 8: *a) Simplified geological map of the Basque-Cantabrian Basin with structural patterns related to the Late Jurassic-Early Cretaceous rifting phase and with the strain ellipse (in the inset) related to the inferred tectonic regime (redrawn and modified after Cámara, 2017). In green: Northern rift fault system; in magenta: southern rift fault system; in blue: listric faults and salt-related structures. b) Geological map of the western Basque-Cantabrian Basin showing the structural pattern of the main faults controlling the mid-Cretaceous transtensional phase (modified after Soto et al., 2011). c) Schematic reconstruction of the relative position of Iberia and Eurasia between the Late Jurassic and the Early Cretaceous with the location of the Basque-Cantabrian Basin in the frame of the plate kinematic context; this model proposes an initial Tithonian phase of orthogonal rifting followed by a subsequent phase of left-lateral transtensional extension during the Albian (redrawn and modified after Soto et al., 2007).*

2.6. The Pyrenean belt and its Mesozoic basins

The Pyrenean belt is an E-W trending alpine orogen marking the border between France and Spain. It resulted from the Late Cretaceous-Cenozoic convergence and collision between Iberia and Eurasia, which induced the tectonic inversion of an Early Cretaceous hyperextended rift system (e.g. Choukroune & ECORS Team, 1989; Roure *et al.*, 1989; Muñoz, 1992; Teixell, 1998; Vergés *et al.*, 2002; Teixell *et al.*, 2018; Espurt *et al.*, 2019; Saspiturry *et al.*, 2020a; Angrand *et al.*, 2021). Between the northerly Aquitaine retro-foreland basin and the southerly Ebro pro-foreland basin, the Pyrenees are classically subdivided into three different E-W trending structural domains (from north to south) (Fig. 9): i) the North Pyrenean Zone, ii) the Axial Zone and iii) the South Pyrenean Zone (e.g. Mattauer, 1968; Choukroune & Séguret, 1973; Choukroune, 1992). The North Pyrenean Zone is bounded to the north by the North Pyrenean Frontal Thrust and to the south by the North Pyrenean Fault, which has often been considered to mark the limit between the Iberian and Eurasian plates (e.g. Mattauer, 1968; Choukroune & Mattauer, 1978; Choukroune & ECORS Team, 1989; see Section 2.6.2 for details on the controversies regarding this latter structure). This structural domain is the one that best preserves traces of the Early Cretaceous rifting phase, whose main expression corresponds to the North Pyrenean basins (see Section 2.6.1). It also includes a series of amygdaloidal Paleozoic massifs, 10s of kilometres wide, known as North Pyrenean massifs (see Section 2.6.3). The pre- and syn-rift sedimentary fill of the southern North Pyrenean basins, enclosed between the North Pyrenean massifs and the North Pyrenean Fault, recorded widespread syn- to post-rift (~110-85 Ma) HT/LP metamorphism (the Internal Metamorphic Zone, e.g. Ravier, 1959; Azambre & Rossy, 1976; Albarède & Michard-Vitrac, 1978; Montigny *et al.*, 1986; Golberg *et al.*, 1986; Golberg & Maluski, 1988; Golberg & Leyreloup, 1990; Clerc *et al.*, 2015) related to severe crustal thinning and mantle exhumation during the Cretaceous (e.g. Vielzeuf & Kornprobst, 1984; Golberg & Maluski, 1988; Dauteuil & Ricou, 1989; Golberg & Leyreloup, 1990; Clerc &

Lagabrielle, 2014 ; Clerc *et al.*, 2015; Lagabrielle *et al.*, 2016) (Fig. 9). The Axial Zone represents the Paleozoic basement of the Pyrenean belt and lies between the North and South Pyrenean zones. Traces of the Cretaceous rifting are substantially lacking in the Axial Zone. The South Pyrenean Zone is bounded to the south by the South Pyrenean Frontal Thrust and is mainly composed of Mesozoic series covered by syn- to post-orogenic Cenozoic deposits. In the central part of this domain, the South Pyrenean Central Unit exposes several allochthonous Cretaceous basins (such as the Organyà and Cotiella basins), which are decoupled at their base at the level of the Triassic evaporites and thrusted southward onto the Cenozoic foreland (see Section 2.6.4).
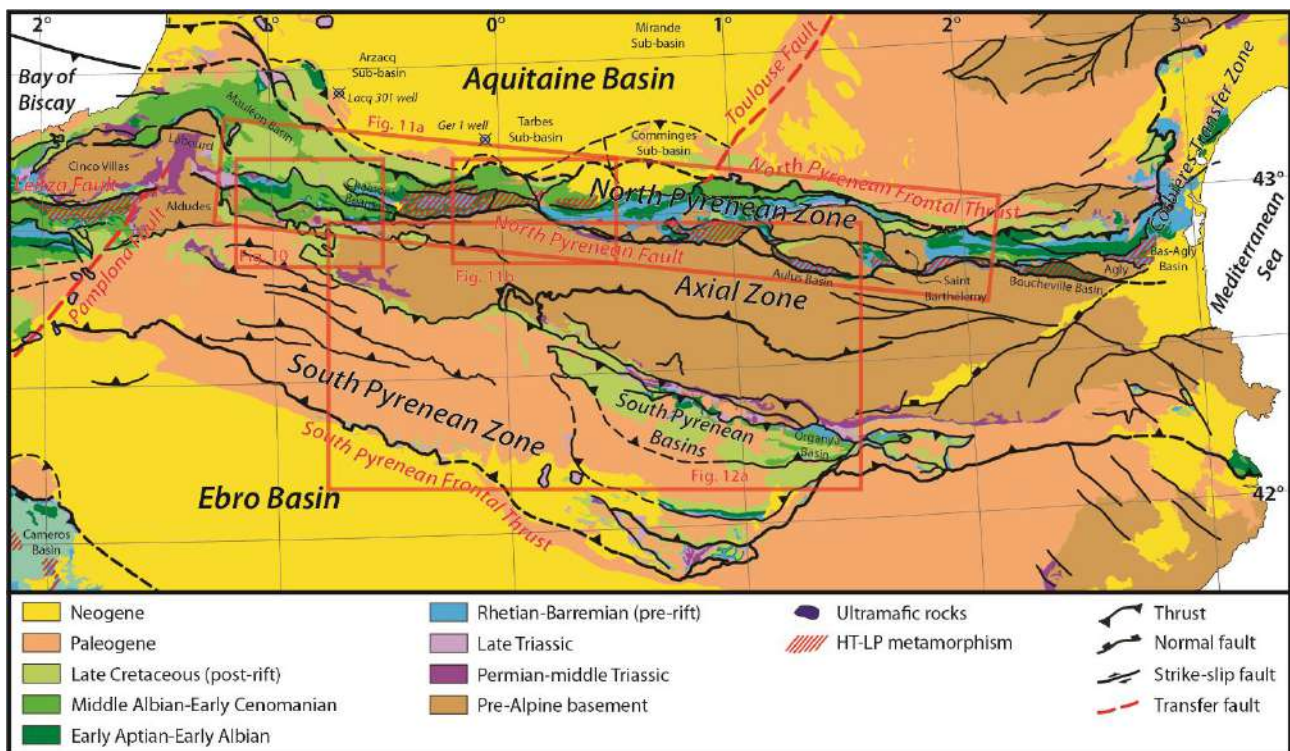
Fig. 9: Geological map of the Pyrenean Belt with locations of the structural maps and schematic reconstructions presented in Figs. 10, 11 and 12 as the boxes outlined in red and of the wells from the Southern Aquitaine basins presented in Fig. 14 (modified after Ducoux et al., 2019).

2.6.1. *North Pyrenean basins*

The North Pyrenean basins consist of a series of inverted Cretaceous sedimentary basins developed along an E-W trend, from the westerly Mauléon Basin to the easterly Boucheville/Bas-Agly basins (Fig. 9). Extension and crustal thinning in the North Pyrenean basins started during the latest Carboniferous-Early Permian, related to the late- to post-orogenic evolution of the Variscan

Belt (e.g. Lucas, 1985; Bixel and Lucas, 1987; de Saint Blanquat *et al*., 1990; Bouhallier *et al*., 1991; de Saint Blanquat, 1993; Cochelin *et al*., 2017; Saspiturry *et al*., 2019a; Asti *et al*., 2021). This initial phase of thinning was followed by a Late Permian-Triassic rifting phase culminating in the deposition of ~1-2 km thick Keuper evaporitic series (e.g. Curnelle, 1983; Calvet & Lucas, 1995; Canérot *et al*., 2005; Biteau *et al*., 2006; Saspiturry *et al*., 2019a). These first episodes of crustal stretching resulted in a significant thinning of the continental crust prior to the Early Cretaceous hyperextensional stage (Espurt *et al*., 2019; Asti *et al*., 2021). Subsequently, a quiet period of thermal subsidence led to the development of stable transgressive carbonate platforms and to the deposition of up to ~2000 m thick post-rift Jurassic sediments (e.g. Delfaud and Henry, 1967; Delfaud and Villanova, 1967; Arnaud-Vanneau *et al*., 1979; Canérot, 1991; Lenoble, 1992; James, 1998; Debrand-Passard *et al*., 1995 and references therein). The late period of this phase is marked by a generalized emersion of the carbonate platforms, with local development of bauxites, between the Late Jurassic and the earliest Barremian (e.g. Souquet *et al*., 1995; Combes *et al*., 1998; James, 1998; Canérot *et al*., 1999; Saspiturry *et al*., 2019b). A subsequent generalized transgression allowed the deposition of Barremian-Aptian carbonate platform sediments above the Neocomian erosional surface (Delfaud & Villanova, 1967; Arnaud-Vanneau *et al*., 1979; Souquet *et al*., 1995 and references therein). These deposits onlapped on a syncline-shaped Jurassic substratum, which probably acquired this geometry during the Neocomian emersion phase (Saspiturry *et al*., 2019b). This might indicate that the onset of Cretaceous extension in the North Pyrenean realm occurred at some time during the Neocomian emersion, leading to the formation of a syncline-shaped basin like in early phases of other smooth-slope basins across the Iberia-Eurasia transition (see Lagabrielle *et al*., 2020; Saspiturry *et al*., 2021), but without the deposition of proper syn-rift deposits due to intense bauxitization of the Jurassic substratum. However, the main rifting phase in the North Pyrenean Zone is considered to have occurred during the Early Albian-Early Cenomanian period and led to the deposition of 4000-5000 m thick syn-rift turbiditic series (Souquet *et al*., 1985, 1995; Debroas, 1990; Debroas *et al*., 2010). This phase culminated in crustal breakup resulting in

lithospheric mantle exhumation below the Cretaceous basins (e.g. Vielzeuf & Kornprobst, 1984; Fabriès *et al*., 1991, 1998; Jammes *et al*., 2009; Lagabrielle *et al*., 2010; Teixell *et al*., 2016; Espurt *et al*., 2019; Garcia-Senz *et al*., 2019b; Labaume & Teixell, 2020; Saspiturry *et al*., 2020a, 2021b; Angrand *et al*., 2021). Syn-rift series are covered by Middle Cenomanian-Santonian post-rift calci-turbidites, which predate the onset of the Late Santonian to Eocene convergence phase (e.g. Puigdefàbregas & Souquet., 1986; Razin, 1989; Puigdefàbregas *et al*., 1992; Debrand-Passard *et al*., 1995 and references therein). During the Cretaceous rifting phase, the thick Keuper evaporites at the base of the pre-rift stratigraphic pile played a fundamental decoupling role and strongly controlled the architecture and the thermal regime of the Cretaceous basins (Canérot, 1988; Jammes *et al*., 2009, 2010a; Lagabrielle *et al*., 2010, 2020; Corre *et al*., 2018; Duretz *et al*., 2019; Saspiturry *et al*., 2021). From the beginning of the Albian extension to the Santonian onset of rift inversion, the pre-rift and syn-rift series recorded a significant thermal anomaly which resulted in HT-LP metamorphism (Albarède and Michard-Vitrac, 1978; Montigny *et al*., 1986; Golberg *et al*., 1986; Golberg and Maluski, 1988; Clerc *et al*., 2015; Lescoutre *et al.*, 2019; Saspiturry *et al.*, 2020b) controlled by an abnormally elevated thermal gradient, i.e. ~60-80°C/Km (Vacherat *et al*., 2014; Corre, 2017; Hart *et al*., 2017; Asti *et al*., 2021; Saspiturry *et al*., 2020b).
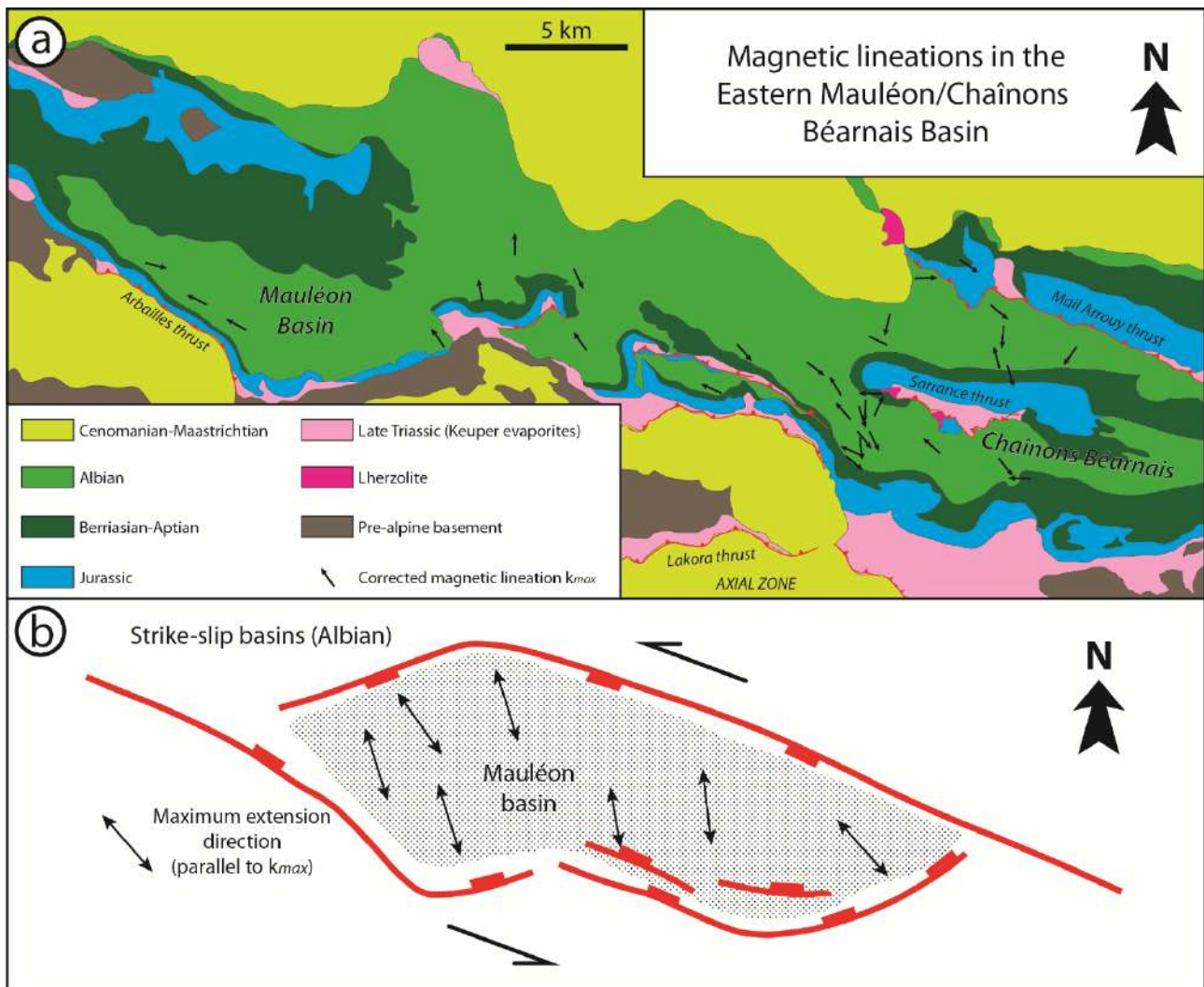
Fig. 10: a) Geological map of the Eastern Mauléon/Chaînons Béarnais Basin with orientations of the magnetic lineations after bedding correction retrieved by AMS studies. b) Schematic model proposing the development of the Albian rifting phase in the Mauléon Basin under a sinistral strike-slip regime and illustrating the possible orientation of the magnetic lineations before the Cenozoic compression. Both images are modified after Oliva-Urcia et al., 2010.

The kinematic evolution of the North Pyrenean rift system has been debated for a long time and remains a very controversial subject. No consensus exists on the stretching direction among the different North Pyrenean basins. The main consequence has been the proposal of different kinematic models explaining their formation, spanning from pull-apart transtensional tectonics (e.g. Choukroune et Mattauer., 1978; Debroas, 1978, 1987) to orthogonal rifting (e.g. Jammes *et al*., 2009; Masini *et al*., 2014). While in the western Pyrenees the Mauléon Basin mainly shows evidence for a NNE-SSW trending direction of extension during its mid-Cretaceous opening (e.g. Jammes *et al*., 2009; Masini *et al*., 2014; Lescoutre, 2019; Saspiturry *et al*., 2019b), the rest of the North Pyrenean basins display compelling evidence for a NW-SE trending transtensional opening.

Several studies have proposed that the North Pyrenean Zone would have accommodated the sinistral movement of the Iberian plate with respect to Europe as a transtensional corridor where pull-apart basins developed (e.g. Le Pichon *et al*., 1970; Choukroune *et al*., 1973; Choukroune, 1976, 1992; Choukroune & Mattauer, 1978; Peybernès et Souquet, 1984; Debroas, 1990, 1995). In the Chaînons Béarnais region, immediately to the east of the Mauléon Basin, AMS studies have revealed a main N-S to NW-SE direction of stretching during the Albian rifting resulting from left-lateral transtension (Oliva-Urcia *et al*., 2010) (Fig. 10). Accordingly, Canérot (2017) proposed that the transitional domain between the Mauléon and Chaînons Béarnais basins opened as a pull-apart basin between the latest Aptian and the Early Cenomanian, controlled by sinistral displacement along its northern and southern margins. In the central North Pyrenean Zone, studies on the tectono-stratigraphic evolution of the sedimentary basins have proposed a pull-apart mechanism with a NW-SE trending direction of stretching (Debroas, 1987, 1990, 1995, 2003) (Fig. 11). From the Baronnies Basin (to the west) to the Bessède-de-Sault region (to the east), most of the stretching lineations developed in the sedimentary cover, in relation to the syn-rift HT-LP metamorphism, point to a NW-SE direction of stretching, consistent with a transtensional opening of the Pyrenean rift system (Clerc, 2012; Clerc *et al*., 2015; Ducoux, 2017). A notable exception is represented by the eastern termination of the North Pyrenean basins (i.e. Boucheville Basin and Bas-Agly syncline), where stretching lineation directions are more scattered and trend NNE-SSW, E-W and N-S (e.g. Clerc, 2012; Vauchez *et al*., 2013; Clerc *et al*., 2015, 2016; Chelalou *et al*., 2016; Ducoux, 2017). To sum up, evidence for NW-SE trending oblique extension is rare in the western (i.e. Mauléon Basin) and easternmost (i.e. Boucheville Basin and Bas-Agly syncline) Pyrenees, where a NNE-SSW direction of extension is documented (e.g. Vauchez *et al*., 2013; Clerc *et al*., 2015, 2016; Ducoux, 2017; Lescoutre, 2019; Saspiturry *et al*., 2019b), but is widespread all along the central and eastern parts of the North Pyrenean Rift System (e.g. Clerc *et al*., 2015; Ducoux, 2017).
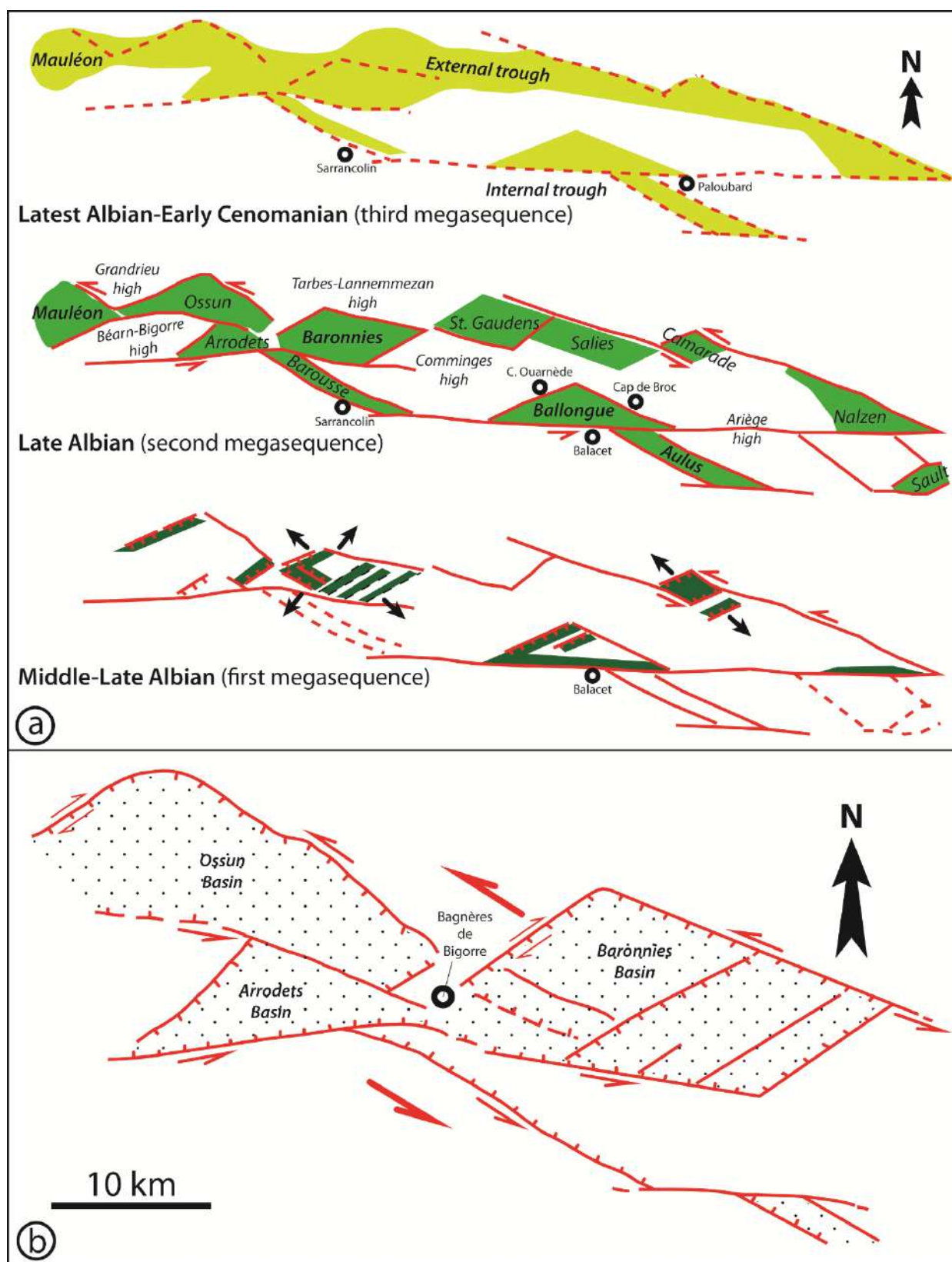
Fig. 11: *a) Schematic representation of the progressive development of the North Pyrenean basins between the Middle Albian and the Early Cenomanian by gradual connection of isolated pull-apart basins to form a single sedimentary trough in a left-lateral transtensional regime (modified after Debroas, 1995, 2003). b) Schematic structural model illustrating the opening of the rift basins exposed in the Bigorre region as sinistral pull-apart basins during the Albian-Cenomanian (modified after Debroas, 1990).*

2.6.2. The North Pyrenean Fault

The North Pyrenean Fault is a nearly E-W trending, ~300 km long, sub-vertical strike-slip structure which was inherited from late-Variscan times and has classically been considered to represent the limit between the Iberian and Eurasian plates (e.g. Mattauer, 1968; Choukroune, 1976; Choukroune & Mattauer, 1978; Choukroune & ECORS Team, 1989) (Figs. 1 and 9). Its formation has been related to a series of strike-slip shear zones that accommodated the dextral displacement between Iberia and Eurasia during the late-Variscan period (e.g. Arthaud & Matte, 1975; Matte 2001). Afterwards, it was characterized by left-lateral kinematics during the Early Cretaceous (e.g. Choukroune, 1976; Choukroune & Mattauer, 1978).

Often presented in the literature as a single and continuous structure, the North Pyrenean Fault is indeed a strike-slip fault system consisting of several subparallel segments (see historical review in Souquet *et al*., 1977). This system is more or less continuously exposed between the easternmost Pyrenees and the Chaînons Béarnais region (to the west), but its trace fades away to the west between the Chaînons Béarnais and the Labourd Paleozoic Massif (e.g. Hall & Johnson, 1986). Further west, it aligns with the strike-slip Leitza Fault of the eastern Basque-Cantabrian Basin, separating the Cinco Villas Paleozoic Massif and the Nappe des Marbres metamorphic unit (Figs. 7 and 9). This structure has often been considered as the western prolongation of the North Pyrenean Fault (e.g. Choukroune, 1976; Rat, 1988; Mendia & Ibarguchi, 1991; Combes *et al*., 1998). The transition zone between these two domains consists of a complex corridor of en-echelon strike-slip faults (Canérot *et al*., 2004; Canérot, 2016 and references therein). However, recent publications have proposed that the Leitza Fault actually corresponds to a former syn-rift detachment fault which was subsequently reactivated during the convergence (see DeFelipe *et al*., 2018; Ducoux *et al*., 2019; Lescoutre, 2019; Lescoutre & Manatschal, 2020; Manatschal et al., 2021) such that the strike-slip component is only minor.

Several plate kinematic models have considered that the North Pyrenean Fault accommodated ~400-500 km of left-lateral displacement between Iberia and Eurasia during the Late Jurassic-Early Cretaceous (e.g. Le Pichon *et al.*, 1970; Choukroune & Mattauer, 1978; Olivet, 1996; Jammes *et al.*, 2009) (Fig. 2). However, despite widespread evidence for a NW-SE transtensional component during the opening of the Pyrenean rift system, the maximum sinistral displacement observed on the North Pyrenean Fault is on the order of ~20 km (Debroas, 1987). The minor contribution of this fault to the Late Jurassic-Early Cretaceous eastward displacement of Iberia is also suggested by the insignificant displacement of the previous paleogeographic markers (e.g. Peybernès, 1978; Canérot, 2016). Some authors have also suggested that the deformation between the Iberian and Eurasian plates was not accommodated strictly by the North Pyrenean Fault, but rather by the diffused continental transtensional system of pull-apart basins represented by the North Pyrenean Zone during the Late Jurassic, which later stretched under ~NNE-SSW orthogonal extension during the Aptian-Cenomanian (e.g. Jammes *et al.*, 2009; Tugend *et al.*, 2015a).

2.6.3. North Pyrenean Basement

The Pyrenees were mainly affected by Carboniferous to Permian metamorphism due to crustal thinning and lithospheric mantle delamination (Denèle *et al.*, 2014; Cochelin *et al.*, 2017; Lemirre *et al.*, 2019 and references therein), resulting in the partial melting of the middle/lower crust, which then intruded into the base of the upper crust of the Axial Zone as gneiss domes along major ductile shear zones (Denèle *et al.*, 2007; 2009; Cochelin *et al.*, 2017, 2018a, 2018b). Within the Variscan crust of the Pyrenees, thinning was accommodated by the activation of extensional shear zones at the top of migmatitic gneiss domes (Mezger and Passchier, 2003; Cochelin *et al.*, 2017). The associated direction of stretching was both longitudinal and transverse, illustrating strain partitioning within the partially molten crust (Cochelin *et al.*, 2018b).

Several ductile extensional shear zones are exposed in the Paleozoic basement of the North Pyrenean Zone. Nevertheless, the interpretation of the shear zones exposed in the North Pyrenean massifs is not always straightforward and their periods of activity have been proposed to be either Carboniferous-Permian or Cretaceous in age. However, some ductile extensional structures undoubtedly related to the Cretaceous rifting event exist. In the Urdach Massif of the Chaînons Béarnais (western Pyrenees), crustal rocks associated with the exhumed lithospheric mantle display an ENE-WSW trending stretching lineation, associated with a top-to-the-ENE sense of shear (Asti *et al*., 2019; Lagabrielle *et al*., 2019). This extensional deformation has been constrained to the Late Albian by ^{40}Ar/^{39}Ar dating on muscovite (Asti *et al*., 2019). Moreover, in the Turon de la Técouère ultramafic massif (Chaînons Béarnais) stretching lineations trending NW-SE are associated with a top-to-the-NW extensional shear zone developed during the Cretaceous mantle exhumation (Vissers, *et al*., 1997; Newman *et al*., 1999). This direction of extension is also consistent with observations from the Lherz ultramafic massif exposed in the Aulus Basin (e.g. Lagabrielle & Bodinier, 2008).

In the Saint Barthélemy North Pyrenean Massif, a south-verging extensional shear zone is exposed (Passchier, 1982). Even though early studies have considered a Cretaceous age for the formation of this shear zone (e.g. Passchier, 1984; de Saint Blanquat *et al*., 1986), studies on its tectono-metamorphic evolution suggest rather a late/post-Variscan age (Permian) for its activation (de Saint Blanquat *et al*., 1990; de Saint Blanquat, 1993; Delapierre *et al*., 1994; Lemirre, 2018). However, some minor reactivation of this shear zone during the Cretaceous rifting cannot be ruled out (Costa & Maluski, 1988; de Saint Blanquat *et al*., 1990; Odlum & Stockli, 2020). A similar situation is reported for the Agly North Pyrenean Massif (eastern Pyrenees), where a top-to-the-NNE late Variscan extensional shear zone (Bouhallier *et al*., 1991; Olivier *et al*., 2008) has been proposed to also have been reactivated during the Cretaceous rifting event (Vauchez *et al*., 2013; Clerc *et al*., 2016; Odlum & Stockli, 2019).

2.6.4. Organyà Basin (South Pyrenean Zone)

The Organyà Basin is a WNW-ESE trending allochthonous Late Jurassic-Early Cretaceous sedimentary basin exposed in the central part of the South Pyrenean Zone (Berástegui *et al*., 1990; Muñoz, 1992; Vergés & García-Senz, 2001) (Figs. 9 and 12). It is thrusted ~100 km southward with respect to its original position and it is decoupled at its base at the level of the Triassic Keuper evaporites (Muñoz, 1992; Puigdefàbregas *et al*., 1992; Mencos *et al*., 2015). In its present-day position, it resides over Permian-Triassic basins, which in turn lie unconformably over the Variscan basement or over Carboniferous-Permian sediments deposited during the collapse of the Variscan belt (Vergés & García-Senz, 2001 and references therein). The total thickness of the sedimentary fill is ~5000-6000 m, including about 4000-5000 m of latest Tithonian-Middle Albian syn-rift deposits (Berástegui *et al*., 1990; Dinarès-Turell & García-Senz, 2000; Vergés & García-Senz, 2001; García-Senz & Muñoz, 2019 and references therein). The rifting phase includes a sedimentary hiatus between the uppermost Valanginian and the Early Barremian (e.g. Dinarès-Turell & García-Senz, 2000), followed by an increase in sedimentation rates during the Aptian-Early Albian, culminating in a Middle Albian-Middle Cenomanian unconformity (Berástegui *et al*., 1990; Dinarès-Turell & García-Senz, 2000; García-Senz & Muñoz, 2019 and references therein). This uppermost unconformity is also coeval with rifting and mantle exhumation in the North Pyrenean basins (e.g. Lagabrielle *et al*., 2010). Paleomagnetic studies of the syn-rift sedimentary fill have highlighted a ~35° counter clockwise vertical axis rotation during the Albian, which has been interpreted as reflecting the coeval rotation of the whole Iberian plate (Gong *et al*., 2008). Deposition of Late Cenomanian-Early Santonian post-rift series precedes the onset of tectonic inversion in the Late Santonian (Mencos *et al*., 2015 and references therein).

At present, there is no consensus on the kinematic evolution of the Organyà Basin during the Early Cretaceous rifting phase. AMS studies in the syn-rift series provided scattered directions of the magnetic fabric (Gong *et al*., 2009; Oliva-Urcia *et al*., 2011) (Fig. 12a). In particular, Gong *et*

al. (2009) proposed that this complex pattern is the result of a N-S direction of extension (in present-day coordinates) during the Early Cretaceous (orthogonal) rifting, partially modified during the following compressional phase. Nevertheless, Oliva-Urcia *et al*. (2011), whose interpretation is strengthened by the presentation of coherent mineralogical and meso-structural data and by a comparison with the adjacent Cotiella Basin, proposed that the AMS pattern resulted from NW-SE oriented extension in a left-lateral transtensional tectonic regime (Fig. 12a). By contrast, Tavani *et al*. (2011a) propose a NNE-SSW direction of extension (i.e. orthogonal rifting) for the opening of the Organyà Basin, based on macro- and meso-structural observations on the southern margin of the basin, although this study also reports the existence of NE-SW trending normal faults and tension gashes developed during the rifting phase (Fig. 12b).
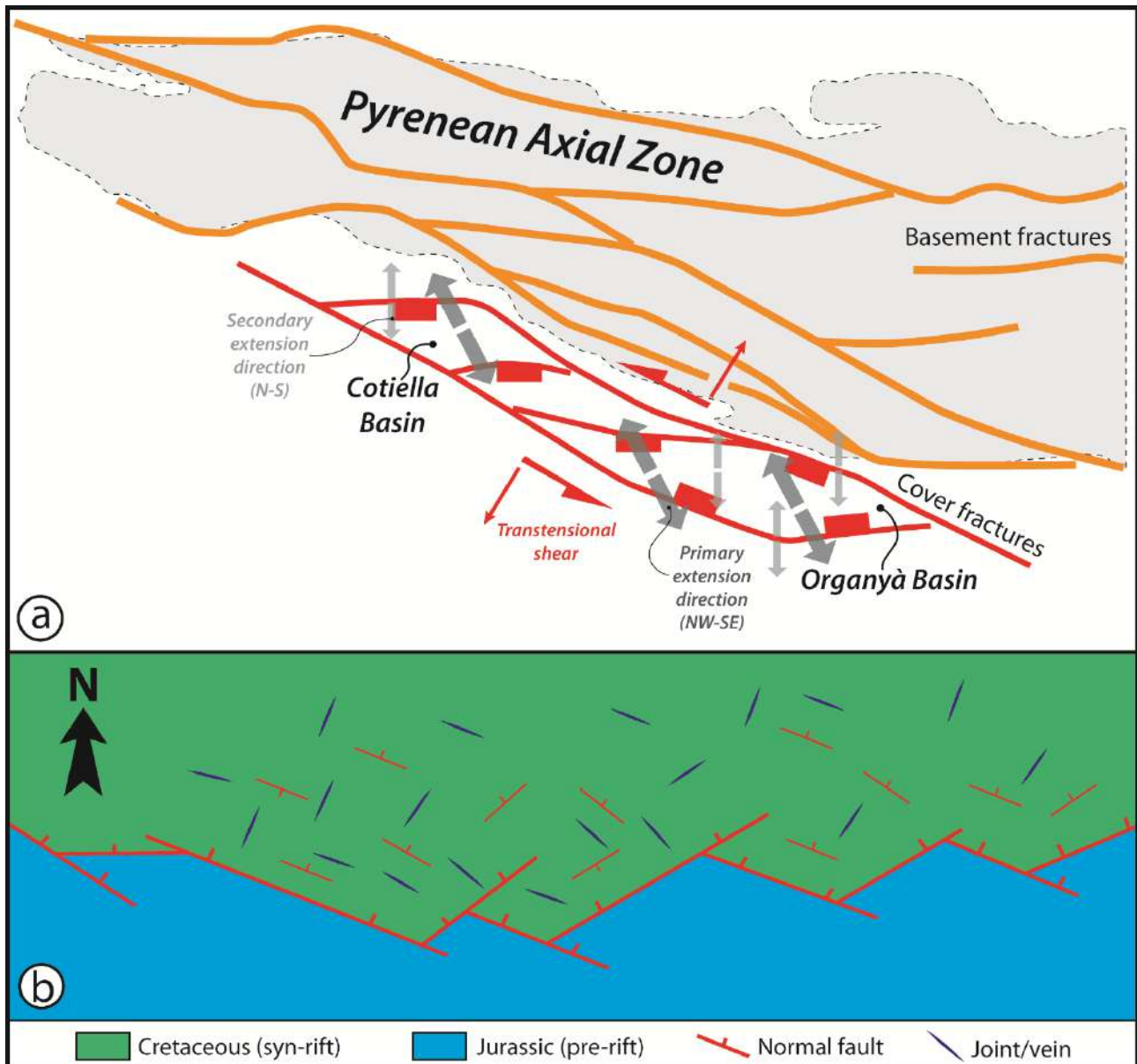
Fig. 12: *a) Schematic representation of faulting relationships in the Pre-Alpine basement of the Pyrenean Axial Zone and in the Mesozoic series of the South Pyrenean basins, with illustration of the extension directions (grey arrows) during the Aptian-Albian transtensional interval retrieved by structural and AMS studies (redrawn and modified after Oliva-Urcia et al., 2011). b) Schematic map view representation of the structural patterns related to the Early Cretaceous extensional phase on the southern margin of the Organyà Basin, showing the general orientation of the main extension-related structures such as normal faults and joints/veins (modified after Tavani et al., 2011a).*

2.7. Parentis Basin

The Parentis Basin is a slightly inverted, E-W trending offshore basin. It represents the eastern termination of the V-shaped Bay of Biscay rift system and is located between the Southern Armorican Shelf (to the north) and the Landes Plateau (to the south) (e.g. Bois *et al.*, 1997) (Figs. 1,

13a and 14a). It is filled by up to 10-11 km of Permian-Mesozoic sediments accumulated in response to ~40 km of crustal stretching (Brunet, 1994; Bourrouilh *et al*., 1995; Bois *et al*., 1997; Vergés & García-Senz, 2001 and references therein). The continental crust has been progressively thinned to a total thickness of only 5-6 km below the base of the basin (e.g. Pinet *et al*., 1987; Bois *et al*., 1997) (Figs. 13b and c). An initial phase of Permian-Triassic rifting culminated in the deposition of thick Keuper evaporites and was followed by the deposition of up to ~2 km thick Jurassic post-rift series (e.g. Mathieu, 1986; Desegaulx & Brunet, 1990; Brunet, 1994; Bourrouilh *et al*., 1995). A subsequent rifting phase took place between the latest Jurassic (Kimmeridgian/Tithonian boundary?) and the Early Cretaceous (Early Albian) resulting in the accumulation of up to ~4-5 km thick syn-rift deposits (e.g. Mathieu, 1986; Brunet, 1984, 1994; Curnelle, 1995). This phase includes a Neocomian interval of reduced subsidence (e.g. Brunet, 1997) and the development of a top-Barremian unconformity (e.g. Ferrer *et al*., 2012 and references therein). Several studies have highlighted the important decoupling role played by the Keuper evaporites in the lower part of the pre-rift series during the Early Cretaceous rifting phase (e.g. Jammes *et al*., 2010a, b, c; Ferrer *et al*., 2012, 2014; Lagabrielle *et al*., 2020). After the Late Albian, the Parentis Basin underwent post-rift thermal subsidence (e.g. Mathieu, 1986; Roca *et al*., 2011) and thickness variations observed in the Late Cretaceous deposits are mainly controlled by post-rift salt diapirism (Roca *et al*., 2011).
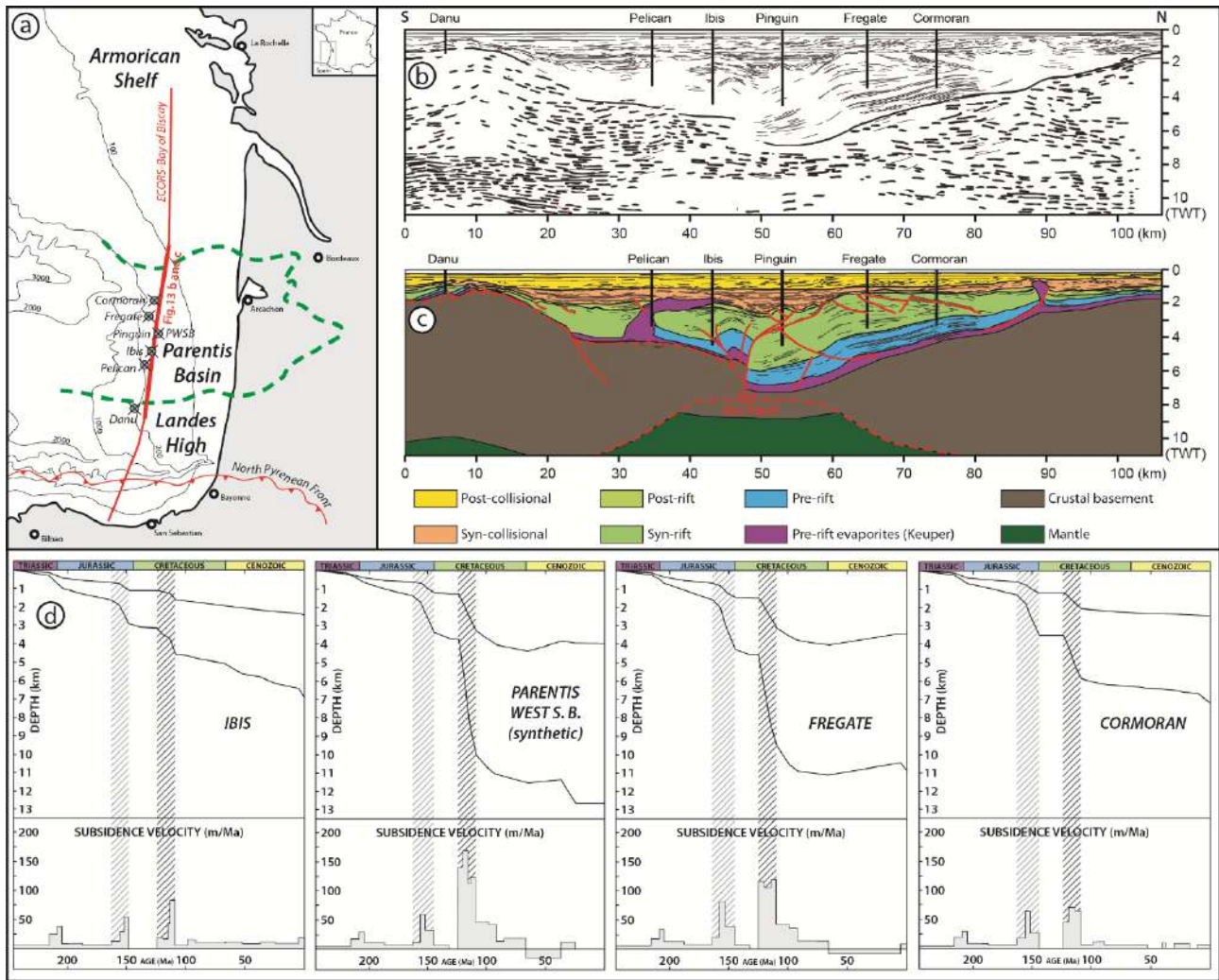
Fig. 13: *a) Locations of the ECORS-Bay of Biscay seismic reflection profile across the Parentis Basin and of the boreholes used for its calibration (redrawn and modified after Bois et al., 1997). b) Line drawing and c) interpretation of the central portion of the ECORS-Bay of Biscay seismic reflection profile (modified after Jammes, 2009). Note the overall flower structure of the Parentis Basin in cross section, with most of the (reactivated) extensional structures converging towards the Ibis Fault. d) Tectonic subsidence curves and tectonic subsidence velocities corrected for sea-level and water depth variations for three oil wells (Ibis, Fregate and Cormoran) and one synthetic well (Parentis West S. B.; see Brunet, 1991) (redrawn and modified after Brunet, 1997). Upper curves represent tectonic subsidence under water with local isostatic compensation, while lower curves represent basement deepening curves with decompacted sediments (see Brunet, 1997 for details on applied corrections). Note that during the Late Jurassic-Early Cretaceous rifting, tectonic subsidence is noticeably faster in the Early Cretaceous interval than in the Late Jurassic interval.*

Even though sinistral strike-slip pull-apart mechanisms have been proposed for the Cretaceous rifting phase (e.g. Choukroune & Mattauer, 1978; Mathieu, 1986; Biteau *et al*., 2006; Jammes *et al*., 2010a, b, c), deep seismic investigations did not provide irrefutable evidence for such a tectonic scenario (e.g. Bois *et al*., 1997). In particular, Mathieu (1986) proposes a first phase of N-S extension during the Late Jurassic-Aptian and a second Late Aptian-Early Albian phase controlled by left-lateral transtension. In agreement with this interpretation, Biteau *et al*. (2006)

suggest a NW-SE direction of extension during the main Aptian-Albian phase of development of the Parentis depocenter, after a first Late Jurassic phase of NE-SW trending extension. By contrast, Jammes *et al*. (2010a, b, c) propose rather an initial Late Jurassic-Early Aptian phase of sinistral transtension, accommodated by the strike-slip Ibis Fault below the Parentis Basin (Fig. 13c), and a subsequent Late Aptian-Early Albian phase of orthogonal rifting. However, it is not clear which geological data support this scenario, which also seems to contrast with interpretations of the ECORS Parentis seismic reflection profile proposed by Tugend *et al*. (2015b). On this section, a major variation in the Neocomian sediments is described either side of the Ibis Fault, which is not the case for the Albian deposits. Whatever the kinematic interpretation, it is worth noting that these two phases are characterized by substantial differences in subsidence rates, with the Aptian-Albian phase showing the highest subsidence rates (Brunet, 1983, 1984, 1994). Such an evolution of the subsidence rates might be more in agreement with a tectonic evolution from orthogonal rifting to sinistral transtension.

2.8. Aquitaine Basin

The Aquitaine Basin is a wide Late Cretaceous-Cenozoic retro-foreland basin developed during the Pyrenean Orogeny (e.g. BRGM *et al*., 1974; Désegaulx & Brunet, 1990; Christophoul *et al*., 2003; Ford *et al*., 2016; Rougier *et al*., 2016; Angrand *et al*., 2018; Espurt *et al*., 2019; Ortiz *et al*., 2020). It is bounded to the south by the North Pyrenean Frontal Thrust, to the west by the Bay of Biscay passive margins, to the east by the French Central Massif and to the north by the Armorican Massif (Figs. 1 and 14a). While the Late Cretaceous-Cenozoic flexural sequences are deposited in a vast single basin, below the syn-orogenic Cenozoic cover several Late Paleozoic-Mesozoic basins are preserved (e.g. BRGM *et al*., 1974; Curnelle *et al*., 1982; Brunet, 1984; Biteau *et al*., 2006; Serrano *et al*., 2006; Angrand *et al*., 2018; Issautier *et al*., 2020) (Figs. 1 and 14a). The evolution of these basins was strongly influenced by Variscan inheritance (e.g. Autran & Cogné,

1980; Debelmas, 1986; Bourrouilh *et al*., 1995). The pre-orogenic basement of the Aquitaine Basin is divided in a southern and a northern domain by the WNW-ESE trending Celtaquitaine Flexure (BRGM *et al*., 1974) (Fig. 14a). This inherited Variscan structural lineament marks a significant paleogeographic limit, representing the northern boundary of the depositional environment of the Late Triassic-Early Lias Keuper evaporites around the Iberia/Eurasia plate boundary (e.g. BRGM *et al*., 1974; Biteau *et al*., 2006; Serrano *et al*., 2006). To the NW, this structure aligns with the steep scarp between the South Armorican shelf and the Bay of Biscay abyssal plain. The southern Aquitaine domain includes the mildly inverted ~WNW-ESE trending Arzacq, Tartas, Tarbes, Comminges and Mirande basins, while the northern domain includes the NW-SE trending Charente and Périgord-Quercy basins (Figs. 1 and 14a). In the northern domain the Mesozoic sedimentary fill does not exceed 2000 m, while in the southern domain it can locally be thicker than 5000 m (BRGM *et al*., 1974; Serrano *et al*., 2006; Issautier *et al*., 2020). Nevertheless, Late Cretaceous uplift of the northern Aquitaine domain resulted in southward tilting of the Aquitaine platform and subsequent erosion of the older deposits, thus contributing to the relative erosional thinning of the sedimentary cover preserved in the northern domain (Platel, 1987, 1996).
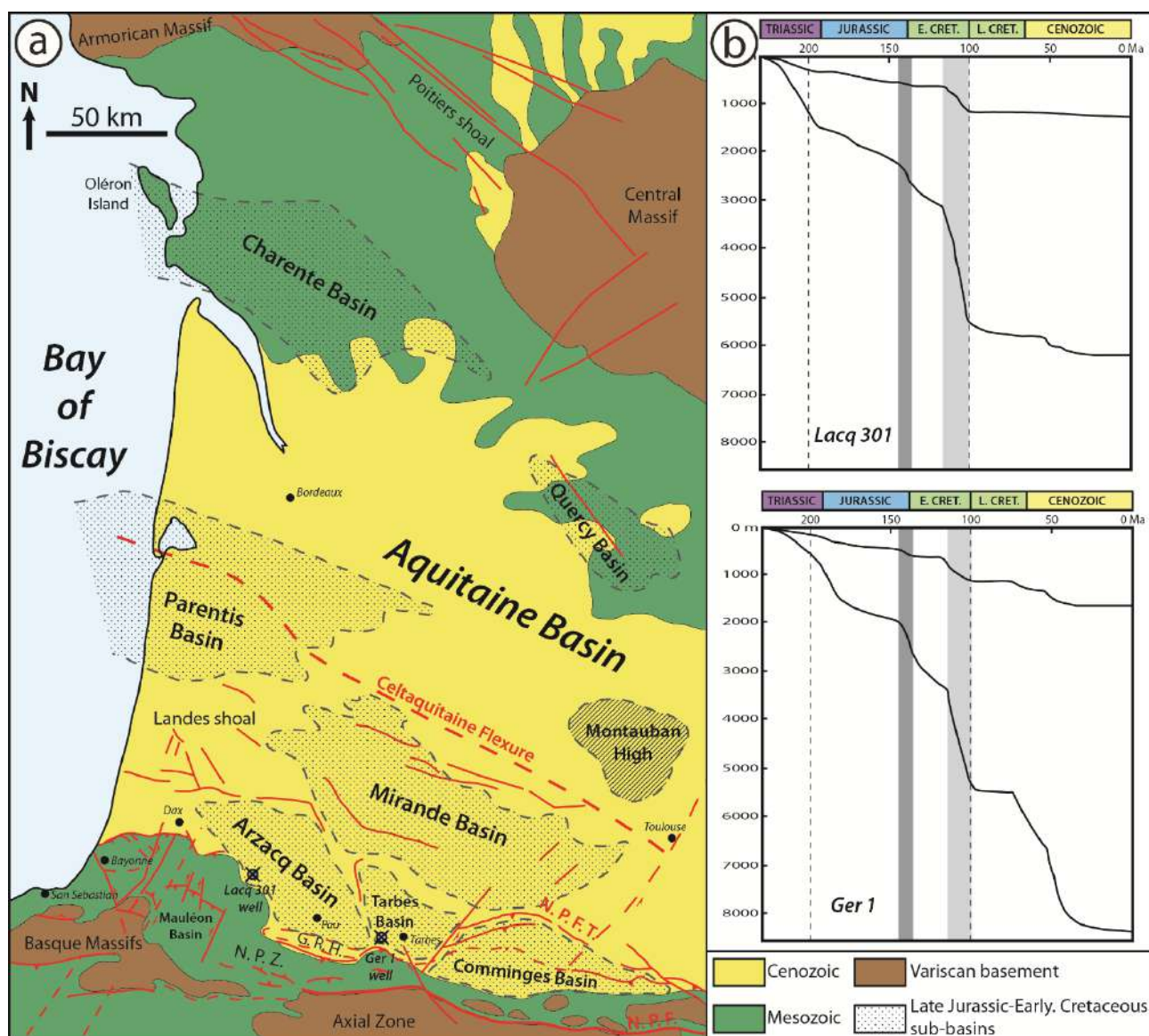
Fig. 14: *a) Simplified geological map of the Aquitaine Basin with locations of the main Late Jurassic-Early Cretaceous sub-basins reviewed in this study and of the wells presented in (b) (redrawn and modified after Bourrouilh et al., 1995). Abbreviations: G.R.H. = Grand-Rieu High; N.P.F. = North Pyrenean Fault; N.P.F.T. = North Pyrenean Frontal Thrust; N.P.Z. = North Pyrenean Zone. b) Subsidence curves for the Lacq 1 (Arzacq Basin) and Ger 1 (Tarbes Basin) wells from the Southern Aquitaine Basin (redrawn and modified after Brunet, 1984; see Fig. 9 for location). Upper curves are tectonic subsidence curves with local isostatic compensation; lower curves represent progressive deepening of the Paleozoic basement with decompacted sediments (see Brunet, 1984 for details on the applied corrections). Note that during the Late Jurassic-Early Cretaceous rifting, tectonic subsidence curves are steeper for the Early Cretaceous interval than for the Late Jurassic one.*

2.8.1. *Southern Aquitaine basins*

Sedimentary basins in the southwestern domain are separated by two major WNW-ESE trending Variscan inherited structures, namely (from south to north) the Grand-Rieu High and Audignon-Pécorade-Antin Maubourguet ridges (Fig. 14a). The former separates the North Pyrenean basins (e.g. Mauléon Basin) from the Arzacq Basin; the latter isolates the Arzacq Basin to the south

from the Tartas/Mirande Basin to the north (Serrano *et al*., 2006; Issautier *et al*.; 2020). This domain of the Aquitaine Basin recorded an initial phase of Permian-Early Lias rifting, which culminated in the Late Triassic-Hettangian deposition of up to ~1 km thick Keuper evaporites (e.g. Curnelle *et al*., 1982; Curnelle, 1983, 1995; Biteau *et al*., 2006; Serrano *et al*., 2006) in a likely transtensional pull-apart setting related to the early breakup of Pangea (e.g. Issautier *et al*., 2020). This phase was followed by an Early-Middle Jurassic phase of post-rift thermal subsidence which led to the development of stable carbonate platforms across the whole Aquitaine Basin realm, resulting in the accumulation of up to ~2 km thick post-rift series (e.g. Curnelle *et al*., 1982; Brunet, 1984; Biteau *et al*., 2006; Serrano *et al*., 2006). Rifting resumed in the Late Jurassic-Early Cretaceous interval with a complex topographic pattern (e.g. Brunet, 1984). This resulted in a patchwork of structural highs (e.g. Grand Rieu and Audignon-Pécorade-Antin Maubourguet highs), which underwent erosion and/or non-deposition between the Kimmeridgian/Tithonian and the Neocomian (i.e. Berriasian-Hauterivian), and localized subsiding domains allowing a sedimentary record of this tectonic event (e.g. Bourrouilh *et al*., 1995; Issautier *et al*., 2020). This phase was followed by a progressive transgression of Barremian-Aptian sedimentation over the structural highs (Issautier *et al*., 2020), connecting both Arzacq and Tartas basins into a single and vast Early Aptian basin. Then, sedimentation rates increased in the Late Aptian-Albian interval (e.g. Brunet, 1983, 1984; Désegaulx & Brunet, 1990; Fig. 14b), with an extensional tectonic regime strongly influenced by the decoupling layer at the base of the pre-rift series represented by the Keuper evaporites (e.g. Curnelle *et al*., 1982; Bourrouilh *et al*., 1995; Issautier *et al*., 2020; Saspiturry *et al*., 2021). The onset of this latter increase in the sedimentation rate seems to be slightly diachronous between these different basins, starting in the Barremian in the Mirande Basin (as in the Parentis Basin) and in the Aptian in the Arzacq and Tartas basins (Brunet, 1983; Issautier *et al*., 2020). Finally, since the Cenomanian, the southern domain of the Aquitaine Basin underwent post-rift thermal subsidence (e.g. Brunet, 1983; Fig. 14b), which continued during the early orogenic phase (Desegaulx *et al*., 1991; Angrand *et al*., 2018). Thick-skinned deformation related to the Pyrenean

compression started during the latest Santonian-Campanian (e.g. Désegaulx & Brunet, 1990; Christophoul *et al*., 2003; Biteau *et al*., 2006; Ford *et al*., 2016; Rougier *et al*., 2016; Espurt *et al*., 2019).

Since the Mesozoic deposits are buried beneath thick Cenozoic series, there are not many firm constraints on the kinematic evolution of the southern Aquitaine sub-basins during the Late Jurassic-Early Cretaceous rifting phase. However, several authors have suggested that this domain accommodated part of the deformation related to the left-lateral displacement between Iberia and Eurasia (e.g. Choukroune & Mattauer, 1978; Curnelle *et al*., 1982; Boillot, 1984; Biteau *et al*., 2006). In particular, most of them proposed an initial Late Jurassic-Neocomian phase of orthogonal rifting under ~NNE-SSW stretching followed by a subsequent phase of Aptian-Albian sinistral transtension resulting in ~NW-SE extension (Boillot, 1984; Bourrouilh *et al*., 1995; Biteau *et al*., 2006; Serrano *et al*., 2006). Nonetheless, other studies based on the Arzacq Basin have rather suggested orthogonal (NNE-SSW) extension during the Aptian-Albian (e.g. Masini et al., 2014; Tugend et al., 2015a; Issautier et al., 2020). However, as discussed above, the Aptian-Albian phase interestingly corresponds to an increase in sedimentation rates (Fig. 14b; Brunet, 1983, 1984; Désegaulx & Brunet, 1990). This is more consistent with a kinematic scenario where the Late Jurassic-Neocomian phase is dominated by orthogonal rifting, whereas the Aptian-Albian phase is dominated by sinistral transtension. In fact, transtensional pull-apart systems typically show higher sedimentation rates than orthogonal rift systems (e.g. Mann *et al*., 1983; Hempton & Dunne, 1984; Gürbüz, 2014 and references therein).

2.8.2. *Northern Aquitaine basins*

The Charente and Périgord-Quercy basins of the northern Aquitaine domain (Fig. 14a) accumulated only a few tens of meters of sediments during the Permian-Early Lias phase (in contrast with the ~2000 m thick pile of sediments locally deposited in the southern domain; BRGM *et al*., 1974). After an Early-Middle Jurassic tectonic quiescence, these two basins recorded

localized subsidence starting in the Kimmeridgian, allowing the deposition of Late Jurassic evaporite-bearing series (BRGM *et al*., 1974; Cubaynes *et al*., 1989; Pélissié & Astruc, 1995). During the Early Cretaceous, the whole northern Aquitaine domain underwent emersion and erosion (e.g. BRGM *et al*., 1974; Cubaynes *et al*., 1989; Combes *et al*., 1998; James, 1998; Canérot *et al*., 1999; Serrano *et al*., 2006; Husson *et al*., 2015), erasing most of the sedimentary record of any tectonic event during this period. However, some Early Cretaceous continental deposits are described in the Charente Basin, aligned along a NW-SE axis (e.g. Platel, 1987; Néraudeau *et al*., 2012; Husson *et al*., 2015 and references therein). The Cenomanian transgressive deposits unconformably overlie the Jurassic series (BRGM *et al*., 1974; Moreau, 1995).

Due to the Early Cretaceous erosion and to major uncertainties on the architecture of these basins, very little is known about what controlled the localization of subsidence during the Late Jurassic. However, Delfaud *et al*. (1995) proposed that this phase could be related to the activation of transtensional systems. Moreover, considering the distribution of the Early Cretaceous continental deposits in very narrow and relatively deep depocenters in Charente, Husson *et al*. (2015) suggested that this could be the result of a tectonically controlled paleogeography, structured in narrow horsts and grabens or in pull-apart basins. In this framework, it is worth noting that the Northern Aquitaine basins are aligned along a NW-SE trend, which is parallel to the Variscan strike-slip structures of the Southern Armorican Massif (e.g. Matte, 1986; Ballèvre *et al*., 2009; Fig. 14a). This trend also continues more to the NW, with major strike-slip structures on the South-Armorican continental shelf (e.g. Thinon, 1999; Barrier *et al*., 2004; Thinon *et al*., 2008, 2018; Lemaire et al., 2021).

3. DISCUSSION: THE MESOZOIC IBERIA/EURASIA PLATE BOUNDARY: A WIDE DOMAIN OF TRANSTENSION PROGRESSIVELY LOCALIZING THE PLATE DISPLACEMENT

The review of the tectono-stratigraphic history of the basins presented in this work has major implications on the timing and distribution of the deformation in the wide transform domain forming the Iberia/Eurasia plate boundary (Fig. 15). Our compilation of basin histories shows that, along both sides of their theoretical limits, the Iberia and Eurasia plates are not rigid but fractured into multiple pieces that may deform independently. Therefore, the displacement of the Iberian plate related to the northward propagation of the Southern North Atlantic and to the opening of the V-shaped Bay of Biscay margins is accommodated within a ~400 km wide NW-SE trending corridor, spanning from the Asturian/Iberian Chain (in the SW) to the Armorican Shelf/Northern Aquitaine domains (in the NE).
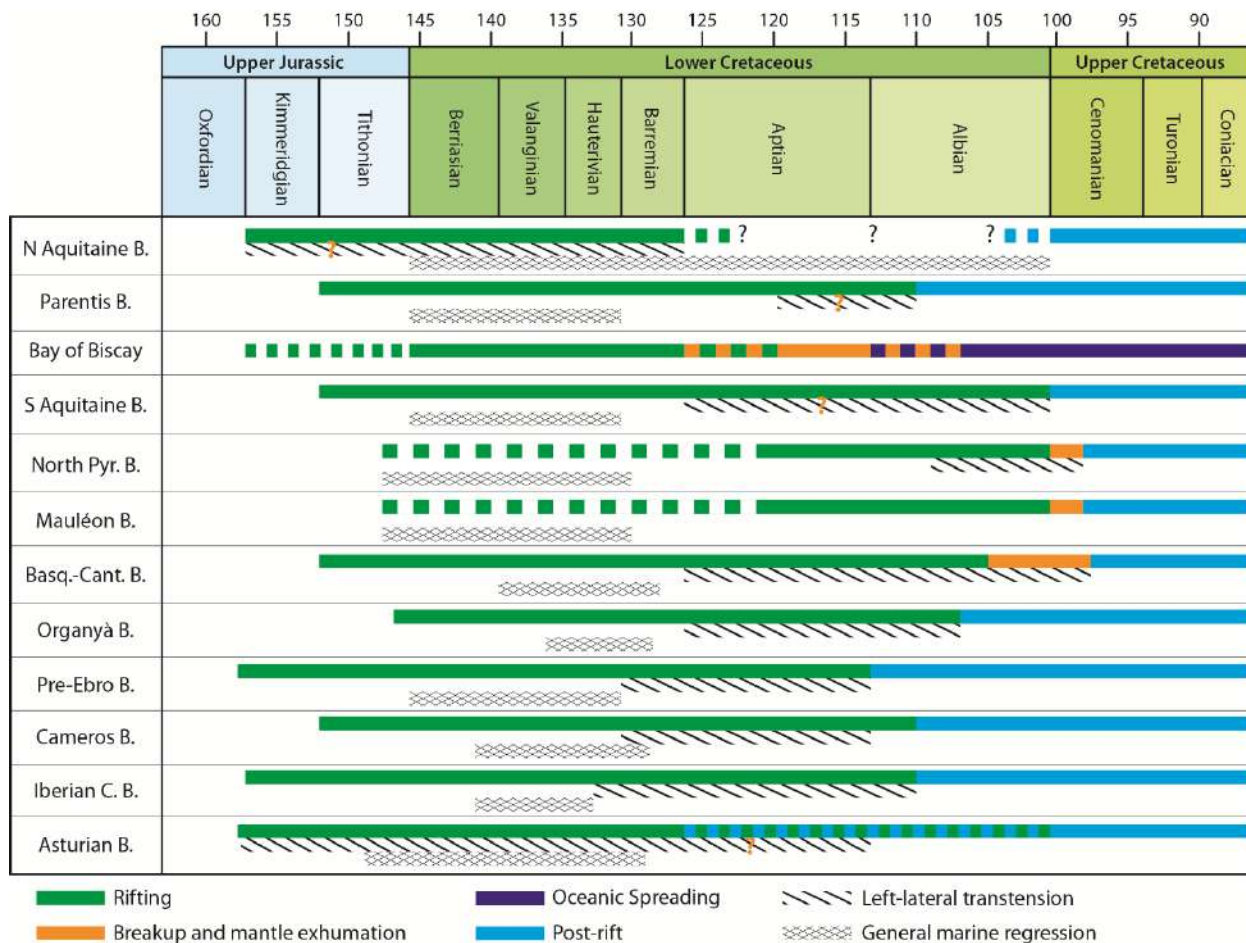

Fig. 15: Schematic diagram showing the main tectonic phases affecting the basins reviewed in this work during the Late Jurassic-Early Cretaceous interval. During the Late Jurassic, extension seems to have initially affected the most external parts of the system and then progressively affected the inner portions. During this phase, the Northern Aquitaine basins and the Asturian Basin, located at the boundaries of the system, underwent oblique rifting, while the rest of the system underwent orthogonal rifting. All the basins were affected by an episode of emersion and/or reduced subsidence during the Neocomian (~145 to 130 Ma). After the Neocomian, most of the rift basins experienced a phase of renewed subsidence controlled by oblique transtensional rifting, with the notable exception of the North

Pyrenean basins. This phase coincided with crustal breakup in the inner Bay of Biscay margins. When oceanic spreading affected the Bay of Biscay domain during the Albian, most of the rift basins became inactive and transtensional rifting accommodating the displacement between Iberia and Eurasia localized along the Basque-Cantabrian and North Pyrenean systems. Rifting finally terminated during the Cenomanian.

3.1. Distribution and timing of deformation along the diffuse plate boundary

To first order, all of the reviewed Mesozoic sedimentary basins recorded the same tectonic phases. However, the identified geodynamic stages are slightly diachronous and get younger from the edges of the system towards its central part (Fig. 15). In the following section, we examine how such differences reveal the behavior and evolution of the Iberia/Eurasia plate boundary.

In particular, most of these basins seem to have experienced two distinct tectonic/kinematic phases, the first one during the Late Jurassic-Berriasian and the second one during the Barremian-Early Cenomanian. These two parts of their tectonic history are separated by a generally Neocomian (~Valanginian-Hauterivian) interval of emersion and/or reduced subsidence, which does not necessarily correspond to a quiet tectonic phase (for example, the Basque-Cantabrian Basin is tectonically active during this phase, but the Valanginian-Barremian syn-rift series are essentially continental; e.g. Camara [2020] and references therein). During the initial stages of plate reorganization, the deformation seems to affect the system edges simultaneously at first, and then progressively migrates towards the central part of the system. In fact, the first basins that started to develop are the Northern Aquitaine basins (in the NE; aligned with the Armorican Shelf) and the Asturian and Central and Southern Iberian Chain basins (in the SW), which underwent rifting since the latest Oxfordian/earliest Kimmeridgian (Fig. 15). The deformation then stepped inward with the activation, since the Tithonian, of the Cameros, Basque-Cantabrian, Southern Aquitaine and Parentis basins. Between the latest Tithonian and the Barremian, rift activation progressivelly reached the inner Bay of Biscay margins (Armorican Basin), the South Pyrenean Rift System (Organyà Basin) and finally the North Pyrenean basins. Indeed, rifting is recorded in the North Pyrenean Zone since the Aptian or even earlier, if the creation of a synform-shaped basin during the Neocomian emersion is considered as the onset of the rifting (e.g. Saspiturry *et al*., 2019b; see

Section 2.6.1). From a kinematic point of view, it is worth noting that this first phase of rifting is characterized by geological evidence suggesting a left-lateral transtensional regime in the Northern Aquitaine and Asturian basins at the edges of the system, while all of the other basins seem to be affected by orthogonal rifting (Figs. 15 and 16). Thus, during this event, the plate boundary behaves as a wide extending domain accommodating the onset of left-lateral displacement of Iberia at its external edges and undergoing orthogonal extension in its interior (Fig. 16a). From a geodynamic point of view, this phase is coeval with rifting along the West Iberian Margin of the Southern North Atlantic (e.g. Alves *et al*., 2002, 2006, 2009).

Fig. 16: Kinematic reconstruction of the Iberia/Eurasia diffuse plate boundary between the Late Jurassic and the Early Cretaceous, with representation of the rift domains reviewed in this study and their kinematic evolution through time. During the Late Jurassic (a) left-lateral displacement between the two plates concentrates on the edges of the system, while the rest of the plate boundary undergoes mainly orthogonal rifting. During the Neocomian (b) the whole plate

The final part of this first rifting phase is characterized by a generalized emersion/regression that affected all of the sedimentary basins within the Iberia/Eurasia plate boundary zone during the Late Tithonian-Neocomian interval (Figs. 15 and 16b). This phase is slightly diachronous between the different basins, but its onset seems to be geographically scattered and thus did not follow the same edges-to-center trend of the onset of rifting. Indeed, the emersion/regression started in the Late Tithonian in the Asturian and North Pyrenean basins while the other reviewed basins were affected between the Berriasian and the Valanginian. The end of this phase coincides globally with the end of the Neocomian, spanning between the Late Hauterivian and the Early Barremian, with the notable exception of the Northern Aquitaine basins, which remained emerged until the Cenomanian (Fig. 15). From a geodynamic perspective, this phase of generalized regression is coeval with northward propagation of crustal necking and mantle exhumation along the West Iberian Margin (e.g. Soares *et al.*, 2012; Mohn *et al.*, 2015; Jeanniot *et al.*, 2016; Alvès & Cunha, 2018; Nirrengarten *et al.*, 2018).

The second phase of rift-related subsidence generally started after the Neocomian (Fig. 15). The boundaries of the system (Asturian and North Aquitaine basins) are less kinematically constrained than the rest of the plate boundary zone, where evidence suggesting a progressive migration of left-lateral transtension towards the center (Pyrenees) is observed. In fact, left-lateral transtension first affected the Iberian Chain and the Pre-Ebro basins from the Late Hauterivian-Early Barremian, then activated in the Basque-Cantabrian, South Pyrenean, Southern Aquitaine and Parentis domains in the Aptian (Fig. 16c). Transtension finally reached the North Pyrenean basins

during the Albian (Fig. 16d). The end of rifting followed the same geographic trend from the Aptian-Albian boundary to the Early Cenomanian and was coeval with the breakup phase of the inner Bay of Biscay (Thinon *et al.*, 2002; Roca *et al.*, 2011) (Figs. 15 and 16) and NW Iberia margins (Soares *et al.*, 2012; Alvès & Cunha, 2018; Alvès *et al.*, 2020). As oceanic spreading began along these margins, the transtensional deformation finally reached the North Pyrenean Rift System, which became, together with the Basque-Cantabrian Basin, the locus of new crustal breakup and lithospheric mantle exhumation (Figs. 15 and 16d). On a larger scale, this phase coincides with continental breakup and progressive onset of oceanic accretion in the Southern Nord Atlantic (e.g. Jagoutz *et al.*, 2007; Tucholke *et al.*, 2007; Alvès & Cunha., 2018).

While the entire system underwent extension since at least the Late Jurassic, it is noteworthly that only the North Pyrenean/Basque-Cantabrian system localized the plate boundary (Fig. 16) and reached mantle exhumation in Albian-Cenomanian times. This observation is consistent with several analogue and numerical oblique rifting models which show that at the onset of stretching the deformation is accommodated over a more or less large region and then progressively localizes in a narrower domain at the center of the extending region (e.g. Autin *et al.*, 2010; Brune, 2014; Duclaux *et al.*, 2020). Onset and localization of rifting in the North Pyrenean/Basque-Cantabrian domain might represent an oblique continental breakup propagation resulting from the eastward propagation of the Bay of Biscay oceanic basin, as suggested by recent numerical models of crustal breakup propagation (see Jourdon *et al.*, 2020). Such a process could have favored the sudden localization of the plate boundary along the North-Pyrenean/Basque-Cantabrian corridor. In fact, in almost all of the basins reviewed in this work, rifting stopped as soon as the Basque-Cantabrian/North Pyrenean corridor underwent crustal breakup (Figs. 15 and 16). The orthogonal component of extension across this trough has likely been important. Indeed, 2D crustal balanced cross-section restorations, across the Pyrenees, show: 1) a pre-orogenic width of exhumed mantle domain ranging from ~10-20 km (Lagabrielle *et al.*, 2010; Teixell *et al.*, 2016, 2018; Espurt *et al.*, 2019; Ternois *et al.*, 2019; Angrand *et al.*, 2021), to ~50/70 km (Mouthereau *et*

al., 2014; Grool *et al.*, 2018; Jourdon *et al.*, 2019), and 2) a N-S width of the synrift turbiditic basins spanning between ~60/90 km (Mouthereau *et al.*, 2014; Clerc *et al.*, 2016; Ternois *et al.*, 2019; Saspiturry *et al.*, 2020a) and ~145 km (Espurt *et al.*, 2019).

To sum up, the Early Cretaceous transtensional deformation at the plate boundary was first distributed and accommodated over a large domain and then localized in a narrower corridor in the central part of the system during the Albian-Early Cenomanian, following lithospheric breakup in the Bay of Biscay (Fig. 16). This episode marks the progressive focusing of the plate boundary along the narrower North Pyrenean rift system. This is thus a good example of progressive localization of deformation, from a wide and distributed plate boundary to a narrow tectonic corridor.

3.2. Timing of the counterclockwise rotation of Iberia

One of the main ongoing debates on the Mesozoic kinematic evolution of Iberia concerns the timing of its counterclockwise rotation in relation to the opening of the North Atlantic and of the Bay of Biscay margins. Despite most authors agreeing on the fact that a rotation of about 35°-40° must have occurred, no consensus exists on its timing (e.g. Van der Voo, 1969; Galdeano *et al.*, 1989; García-Mondéjar, 1996; Gong *et al.*, 2008; Barnett-Moore *et al.*, 2016 and references therein). The most solid reconstructions suggest that this rotation must have taken place sometime between the Triassic and the Late Cretaceous (Van der Voo, 1969). However, it has not yet been clarified if this rotation occurred as a single tectonic event or if it was polyphase and distributed in time. Gong *et al.* (2008), on the basis of paleomagnetic data from the Early Cretaceous sediments of the Organyà Basin, argued for an Aptian 35° counterclockwise rotation. These authors used this argument to state that the rotation of Iberia occurred in a single Aptian event. However, as we have suggested in Section 2.6.4, the Aptian represents a phase of left-lateral transtension within the Organyà Basin. Moreover, the pre- and syn-rift sedimentary cover are strongly decoupled from their basement due to the thick Keuper evaporites at their base and this may have favored the cover

mobility. Thus, it is likely that the rotation evidenced by Gong *et al.* (2008) is, at least partially, related to local rotations in the Organyà Basin due to the transtensional tectonic regime and to the strong basement/cover decoupling, rather than to the rotation of the whole Iberian plate. This is also consistent with local rotations observed during the Early Cretaceous rifting phase of the Organyà Basin by Dinarès-Turell & García-Senz (2000). The reliability of Gong *et al.* (2008)'s hypothesis had already been recently questioned by Barnett-Moore *et al.* (2016). Indeed, these latter studies highlight the difficulty in reconciling the idea of a rotation of Iberia limited to a single Aptian event and the evidence for northward propagation of rifting along the Iberia-Newfoundland margins since the Late Triassic–Early Jurassic (Tucholke *et al.*, 2007; Alvès *et al.*, 2009).

Based on the spatial correlation of the Armorican and the North Iberian margins across the restored Bay of Biscay system, García-Mondéjar (1996) argued for a counterclockwise rotation of about 37° of the Iberian plate and proposed that it took place during the Aptian-Albian interval. However, the review presented in this work shows that the opening of the Bay of Biscay margins was polyphased and occurred over a long time between the Late Jurassic (Kimmeridgian) and the Albian/Cenomanian. Recent works have also shown that displacement of Iberia with respect to North America and Eurasia also occurred during the Permian-Triassic period (Angrand *et al.*, 2020 and references therein). Moreover, paleomagnetic data from the Lisbon region (Portugal; Galdeano *et al.*, 1989) suggest that most of the counterclockwise rotation of Iberia took place since the Barremian, with a larger amount (~27°) between 125 and 110 Ma, and a smaller amount (~14°) after ~110 Ma. This could be compatible with a model implying widespread left-lateral transtension at the scale of the whole plate boundary zone during the Barremian-Early Albian (Figs. 15 and 16c), and a more localized transtension in the Basque-Cantabrian/North Pyrenean corridor during the Albian-Early Cenomanian interval (Figs. 15 and 16d). These arguments thus favor the hypothesis that the counterclockwise rotation of Iberia was polyphase and occurred in response to a series of rifting events at the northern and western plate margins distributed throughout the whole Mesozoic.

3.3. Questioning the existence of the Ebro Block

As discussed above in Section 2.4, several studies concerning the Late Jurassic-Early Cretaceous kinematic evolution of the Iberia-Eurasia plate boundary have postulated that the Ebro realm behaved as a rigid, undeforming continental ribbon while rifting was occurring along the Iberian Chain and the Pyrenean basins at its boundaries (e.g. Tugend *et al*., 2015a; Nirrengarten *et al*., 2018; Tavani *et al*., 2018; Angrand *et al*., 2020; Angrand & Mouthereau, 2021; Frasca *et al*., 2021; King *et al*., 2021) (Fig. 3). However, the review presented in this work shows that some traces of the Late Jurassic-Early Cretaceous rifting exist below the Cenozoic Ebro Basin (Fig. 6). Interestingly, the few existing data on the Mesozoic subsidence of the Ebro domain show that this region, like the other rift basins within the broad Iberia/Eurasia plate boundary zone, recorded two distinct subsidence phases during the Kimmeridgian-Tithonian(?) interval and during the Barremian-Aptian(?), which are separated by a Neocomian interval of reduced subsidence (Vargas *et al*., 2009). The Barremian-Aptian rifting phase shows a higher subsidence rate than the Kimmeridgian-Tithonian one. This could be due to kinematic differences between an initial Late Jurassic phase of orthogonal rifting and a subsequent Early Cretaceous phase of sinistral transtension, as suggested for other sedimentary basins within the plate boundary zone (e.g. Mathieu, 1986; Serrano *et al*., 2006; Martín-Chivelet *et al*., 2019). In light of these considerations, the existence of the Ebro Block as an undeforming continental ribbon during the opening of the Bay of Biscay should be questioned. In our opinion, it is much more likely that the Ebro realm behaved as a deforming domain undergoing rifting during at least the Late Jurassic-Early Cretaceous and accommodating a certain amount of left-lateral displacement between Iberia and Eurasia, as suggested for the other rift basins in its surroundings (Figs. 15 and 16).

3.4. A conceptual model for the Mesozoic Iberia/Eurasia plate boundary zone

In this work we show that the deformation related to the left-lateral displacement between Iberia and Eurasia is accommodated over a large domain (~400 km wide). The transtensional

corridors therein did not react synchronously and each of them accommodated a portion of the deformation between the Late-Jurassic and the Early Cretaceous. Most of the plate kinematic reconstructions agree on the fact that during this period the Iberian plate should have moved several hundreds (~400) of kilometers eastward (e.g. Olivet, 1996; Jammes *et al*., 2009; Nirrengarten *et al*., 2018) (Figs. 2 and 3). However, there has always been an inconsistency between these kinematic models and the lack of a single wrench structure accommodating the displacement between the two plates. This paradox can thus be resolved by looking to the wider domain that accommodated this displacement (Fig. 16).

Some strike-slip faults involved in the transtensional phases of some sedimentary basins within this domain (such as the North Pyrenean Fault, the Leitza Fault, the Ventaniella/Ubierna Fault system, the Hesperian Fault, the North Iberian Fault, the Ibis Fault) accommodated strike-slip displacement on the order of a few tens of kilometers (e.g. Souquet *et al*., 1977; Peybernès, 1978; Debroas, 1987; Canérot, 2016). It is worth noting that, during the transtensional phases, the basins described in this work likely developed as negative flower structures (e.g. Jammes *et al*., 2010c) (see the example of the Parentis Basin in Fig. 13c). This implies that the emerging strike-slip structures that may be observed at the boundaries of these basins (such as the North Pyrenean Fault for example) accommodated only a portion of the total strike-slip displacement accounted for by each flower structure. Hence, amounts of horizontal displacement quantified on oblique/strike-slip structures bounding these basins likely represent a substantial underestimation of the whole horizontal displacement accommodated by each transtensional basin. If we assume that each transitional corridor reviewed in this work (about ten in total) accommodated the same order of magnitude of left-lateral displacement (few tens of kilometers) during the Late Jurassic-Early Cretaceous, then this would imply that the whole domain enclosed between the Iberian Chain and the Armorican Shelf could have accommodated the several hundreds of kilometers of left lateral displacement between Iberia and Eurasia proposed by the plate kinematic models (Fig. 17). It does not strictly mean that each one these rift systems accommodated exactly the same amount of left-

lateral displacement, but at least that this component of the deformation has been distributed over a wide region. This scenario could resolve a long-lasting paradox in Western European Mesozoic tectonics.
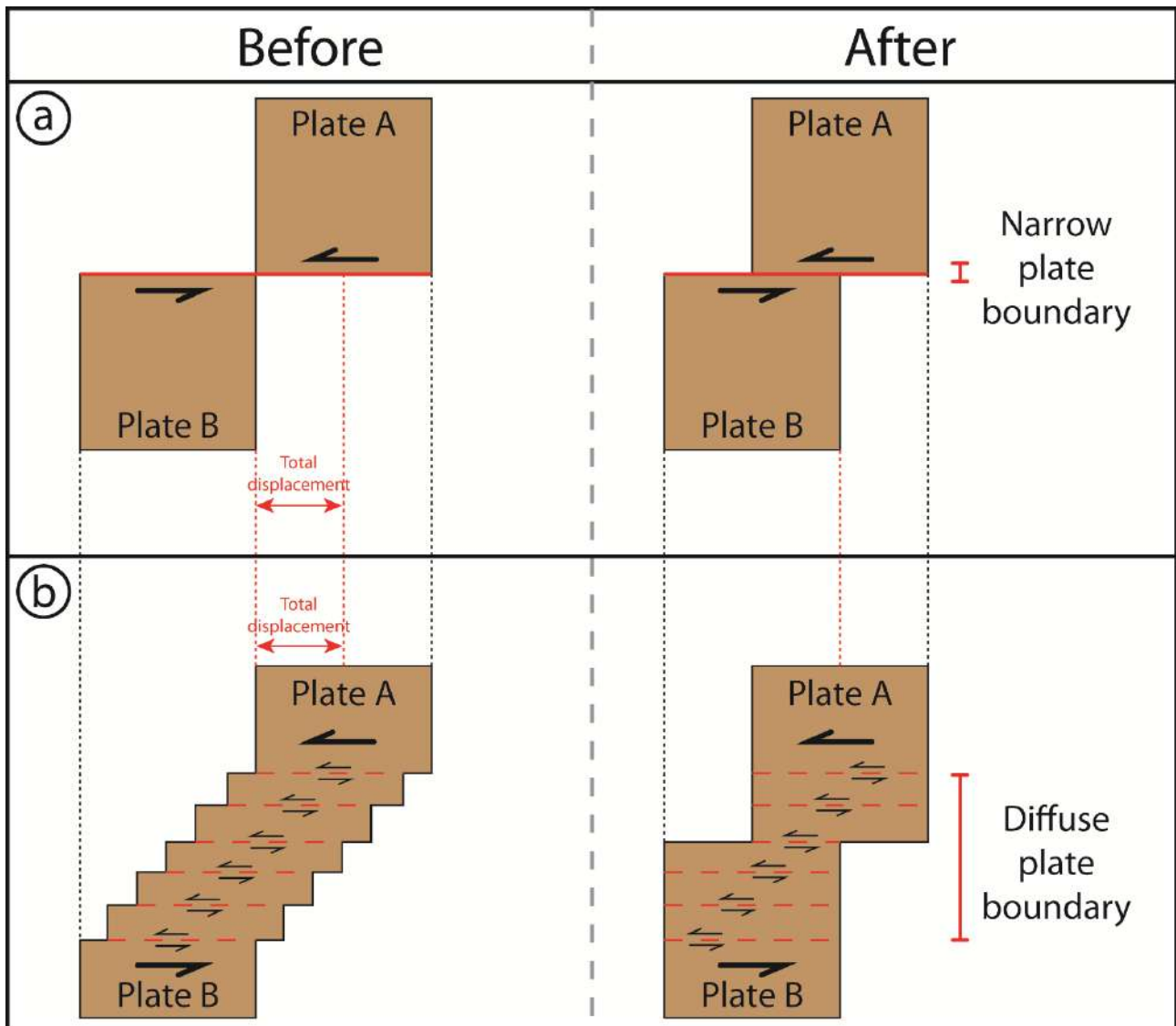
Fig. 17: *Schematic representation of (a) a narrow transform plate boundary and of (b) a diffuse transform plate boundary. In (a) the total left-lateral displacement between the two plates is accommodated by a single wrench structure. In (b) the very same amount of total displacement is partitioned between a series of sub-parallel strike-slip structures, each of which accommodates a portion of the total left-lateral displacement between the stable parts of the plates.*

3.5. What causes the kinematic ambiguity?

The Late Jurassic-Early Cretaceous kinematic evolution of the different rift basins within the Iberia/Eurasia plate boundary zone and their role in the overall plate kinematic context have been largely debated over the last decades. Several reasons might be behind this long lasting controversy.

Particularly, some authors have put forward models arguing for a dominantly orthogonal rifting at the scale of the plate boundary (e.g. Manatschal & Müntener, 2009; Tavani *et al*., 2018), while others have supported models in favor of dominantly left-lateral transtensional mechanisms (e.g. Choukroune & Mattauer, 1978; Olivet, 1996; Stampfli & Borel, 2002). The review of the kinematic evolution of all of the rift domains presented in this paper shows indeed that both tectonic regimes affected these sedimentary basins, but during distinct time intervals. In fact, most of the sedimentary basins in the inner domain of the plate boundary zone were affected by an initial orthogonal rifting phase (generally during the Late Jurassic-Early Neocomian, with ~N-S to ~NE-SW directions of extension), but also by a subsequent sinistral transtensional rifting event (generally between the Barremian and the Albian, with a ~NW-SE direction of extension), although with a potentially significant orthogonal component. Thus, in order to correctly interpret the extension directions recorded in these sedimentary basins in the frame of the overall plate kinematics, one should be careful and properly identify to which tectonic phase the observed geological markers are related. In other words, since both orthogonal and transtensional regimes affected the rift basins within the Iberia/Eurasia plate boundary zone during the Late Jurassic-Early Cretaceous rifting, with more or less pronounced diachronicity between the different basins, the timing and the kinematics of a tectonic event described in a single rift basin cannot be extrapolated to the whole plate boundary domain.

As already pointed out by Oliva-Urcia *et al*. (2011), another possible cause of the controversy regarding the kinematic evolution of the rift basins reviewed in this work is the coexistence of different stretching directions during the transtensional phases, and the confusion that may arise between regional and local extension directions. In fact, in transtensional tectonic settings the deformation is often partitioned between a strike-slip component and an extensional component (e.g. Sanderson & Marchini, 1984; Umhoefer & Dorsey, 1997; Dewey *et al*., 1998; Dewey, 2002; Wilson *et al*., 2006; Autin *et al*., 2010; Brune, 2014). In the reviewed basins, the relative importance of these two components is highly difficult to quantify. In some cases, such as

for the Pyrenean realm (e.g. Oliva-Urcia *et al*., 2010; Tavani *et al*., 2011; Mouthereau *et al*., 2014; Espurt *et al*., 2019; Saspiturry *et al*., 2019b), the orthogonal component of extension might have been important during the transtensional phase. This may increase the complexity of the relative extension directions and of the resulting overall tectonic pattern. Moreover, it has been largely illustrated that, close to major strike-slip faults, stress deflection may cause local deviations of the extension direction (e.g. Casas *et al*., 1992; Simón *et al*., 1999; Homberg *et al*., 2004; Nüchter & Ellis, 2011). This may thus lead to misinterpretations between the regional far-field extension direction and the locally deflected ones. For this reason, the most reliable studies to unravel the kinematic evolution of the basins within the Iberia-Eurasia plate boundary zone are those that consider the overall structural and/or AMS patterns at the scale of entire basins (e.g. Pulgar *et al*., 1999; Soto *et al*., 2007, 2008, 2011; Oliva-Urcia *et al*., 2010, 2011; Camara, 2017), rather than those based on local observations at particular sites within the basins. Moreover, analogue and numerical modeling studies have shown that in transtensional tectonic settings the directions of extension may be scattered and vary though time (e.g. Withjack & Jamison, 1986; Autin *et al*., 2010; Brune, 2014; Duclaux *et al*., 2020; Jourdon *et al*., 2020).

3.6. What favors the distribution of deformation?

As we have discussed above, the deformation related to the left-lateral displacement between Iberia and Eurasia during the Late-Jurassic-Early Cretaceous is distributed over a ~400 km wide region, rather than localized on a single wrench structure. The main reason is probably that this domain was already deformed and strongly fragmented as a consequence of the rift events that preceded the Late Jurassic-Early Cretaceous phase, such the Late Carboniferous-Early Permian extension during the early dismantling of the Variscan belt and the Late Permian-Triassic rifting related to the breakup of the Pangea supercontinent (e.g. Arthaud & Matte, 1975; Autran & Cogné, 1980; Ziegler, 1988; Matte, 2001; Ziegler & Dèzes, 2006; Ziegler *et al*., 2006). These events strongly influenced the transition between Iberia and Eurasia with the formation of large scale

strike-slip faults and shear zones, such as the Ventaniella/Ubierna fault system (e.g. Tavani *et al*., 2011b; Fernández-Viejo *et al*., 2014), the North Pyrenean Fault (e.g. Mattauer, 1968; Choukroune, 1976), the Celtaquitaine Flexure (e.g. BRGM *et al*., 1974) and the South Armorican Shear Zone (e.g. Jégouzo, 1980). Moreover, most of the sedimentary basins in this domain also experienced an important Permian-Triassic rifting phase (e.g. Sopeña *et al*., 1988; Doblas *et al*., 1994; Serrano *et al*., 2006). This could imply that part of the left-lateral motion between the Iberian and Eurasian plates had already been achieved before the Jurassic and not only during the opening of the North Atlantic and Bay of Biscay oceanic basins (Angrand *et al*., 2020; Angrand & Mouthereau, 2021). Therefore, this strong compartmentalization of the wide domain at the transition between the two plates favored the reactivation of all of these systems during the Late Jurassic-Early Cretaceous rifting phase, rather than the localization of the strain along a single narrow plate boundary.

4. CONCLUSION

The review presented in this work proposes a coherent tectonic Mesozoic framework of the wide domain accommodating the deformation between the Iberian and Eurasian plates. The main conclusions can be summarized as follows:

- The Mesozoic deformation at the Iberia/Eurasia plate boundary, related to the opening of the southern North Atlantic and the Bay of Biscay margins, was accommodated across a ~400 km wide region spanning from the Iberian Chain System (in the SW) to the Armorican Shelf/Northern Aquitaine systems (in the NE). Thus, this region should be considered as a diffuse plate boundary zone (*sensu* Gordon, 1998) as opposed to previous interpretations considering a narrow plate boundary along the North Pyrenean Fault only.

- During the Late Jurassic-Early Cretaceous interval, some domains that have been previously overlooked, such as the Pre-Ebro Basin and the Northern Aquitaine basins, have likely accommodated a certain amount of the deformation between Iberia and Eurasia.

- To first order, the rift basins within the diffuse plate boundary zone experienced the same Late Jurassic-Early Cretaceous tectonic history, but with: (i) slight diacronicities between the different systems and (ii) two main subsidence phases characterized either by orthogonal or transtensional rifting.

- The Late Jurassic-Early Cretaceous evolution of the Iberia/Eurasia diffuse plate boundary zone can be summarized in four steps: (i) a Late Jurassic phase of transtensional rifting localized at the northeastern and southwestern margins of the system, while the rest of the sedimentary basins underwent orthogonal rifting; (ii) a Neocomian phase of generalized regression, with emersion of several domains within the diffuse plate boundary zone; (iii) a Barremian-Aptian phase of distributed transtensional rifting, with the notable exception of the North Pyrenean System which was undergoing orthogonal rifting; (iv) an Albian-Cenomanian phase of localization of the transtensional deformation along the Basque-Cantabrian/North Pyrenean corridor after the onset of oceanic spreading in the Bay of Biscay margins, while the rest of the rift basins became tectonically inactive.

- The distribution of the deformation over the wide Iberia/Eurasia diffuse plate boundary zone is favored by the structural pre-rift inheritance, as this region has a strong late/post-Variscan pre-rift heterogeneity.

- The regional tectonic template developed during the Late Jurassic-Early Cretaceous phase represents a fundamental structural inheritance for the subsequent Late Cretaceous-Cenozoic convergence phase, since most of the reviewed rift basins have been more or less inverted during the Pyrenean Orogeny and thus accommodated parts of the Iberia-Eurasia convergence.

More generally, this review has major implications for future plate kinematic reconstructions of Iberia during the Mesozoic. Indeed, we show that the limits of the stable plates should be placed to the SW of the Iberian Chain System (on the Iberian side) and along the

Armorican Shelf/Northeastern Aquitaine boundary (on the Eurasian side). The deforming domain between these two boundaries should be represented as a fuzzy region where exact locations of the different tectonic structures and rift basins therein are uncertain. The inconsistency between the deformation inferred from plate kinematic models of the Iberia/Eurasia plate boundary and geological observations in the Pyrenees has been highlighted for a long time (e.g. Souquet *et al*., 1977). However, kinematic models published in the last four decades have never quite satisfied constraints derived from field observations. Going beyond the assumption that plate boundaries are necessarily narrow and determining the real width of the domains that may accommodate the deformation between moving plates may resolve controversies that often arise between plate kinematic models and geological constraints.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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FIGURE CAPTIONS

Fig. 1: Simplified structural map of the Iberia/Eurasia diffuse plate boundary zone, with locations of the basins presented in this work (modified after Lagabrielle *et al*., 2020). Acronyms: NPFT, North Pyrenean Frontal Thrust; SPFT, South Pyrenean Frontal Thrust.

Fig. 2: Main plate kinematic models proposing a different Late Jurassic-Cretaceous evolution of the Iberia/Eurasia plate boundary. a) Model by Olivet (1996) proposing an Albian-Cenomanian eastward displacement of the Iberian plate accommodated along the North Pyrenean Fault (as redrawn by Mouthereau et a., 2014). b) Model by Jammes *et al*. (2009) proposing left-lateral displacement between Iberia and Eurasia accommodated along the North Pyrenean Fault between the Late Jurassic and the Aptian, followed by Albian-Cenomanian orthogonal (N-S) extension. c) Model by Srivastava *et al*. (2000) and Sibuet *et al*. (2004) proposing a scissor opening of the Bay of Biscay, implying subduction of a large oceanic domain during the Albian in the Pyrenean realm (as redrawn by Mouthereau et a., 2014).

Fig. 3: Recent plate kinematic models by (a) Tugend *et al*. (2015a), (b) Nirrengarten *et al*. (2018), and (c) Angrand & Mouthereau (2021) proposing a partitioning of the Late Jurassic-Early Cretaceous deformation between Iberia and Eurasia in two or three transtensional corridors located in the Iberian Chain Rift System, in the Pyrenees, and in the Bay of Biscay/Parentis system, separated by one or two intervening micro continental blocks (i.e. the Landes High and the Ebro Block).

Fig. 4: Geological map and structural pattern of the Asturian Basin. a) Geological map of the Asturian Basin (including the offshore Le Danois sub-basin) and of the surrounding Cantabrian

region (redrawn and modified after Pulgar *et al.*, 1999). b) Simplified structural map of the onshore Asturian Basin with age of (re)activation of the main structures (redrawn and modified after Lepvrier & Martínez-García, 1990).

Fig. 5: a) Simplified geological map of the Iberian Chain System with locations of the main Permian-Mesozoic rift basins (modified after Guimerà, 2018); isolines in the Ebro Basin refer to base of the Cenozoic series. b) Evolution of extension directions (gray arrows) between the Late Jurassic and the Albian in the Cameros and in the Basque-Cantabrian basins as retrieved by anisotropy of magnetic susceptibility (AMS) analysis (modified after Soto *et al.*, 2008). c) Backstripped tectonic subsidence curves for the southeastern part of the Iberian Chain System (Maestrat and South Iberian basins); see (a) for location of the wells (redrawn and modified after Salas *et al.*, 2001). Note that during the Late Jurassic-Early Cretaceous rifting, tectonic subsidence curves are steeper for the Early Cretaceous interval than for the Late Jurassic one.

Fig. 6: a) Age map of the sedimentary series underlain below the Cenozoic cover of the Ebro Basin as derived from interpolation of well log data (modified after Filleaudeau, 2011). b) Tectonic subsidence curve for the Bujaraloz-1 well (see location in (a); modified after Vargas *et al.*, 2009).

Fig. 7: Geological map of the Basque-Cantabrian Basin with locations of the structural maps presented in Fig. 8 outlined in red boxes (modified after Pedrera *et al.*, 2017).

Fig. 8: a) Simplified geological map of the Basque-Cantabrian Basin with structural patterns related to the Late Jurassic-Early Cretaceous rifting phase and with the strain ellipse (in the inset) related to the inferred tectonic regime (redrawn and modified after Cámara, 2017). In green: Northern rift

fault system; in magenta: southern rift fault system; in blue: listric faults and salt-related structures.

b) Geological map of the western Basque-Cantabrian Basin showing the structural pattern of the main faults controlling the mid-Cretaceous transtensional phase (modified after Soto *et al.*, 2011).

c) Schematic reconstruction of the relative position of Iberia and Eurasia between the Late Jurassic and the Early Cretaceous with the location of the Basque-Cantabrian Basin in the frame of the plate kinematic context; this model proposes an initial Tithonian phase of orthogonal rifting followed by a subsequent phase of left-lateral transtensional extension during the Albian (redrawn and modified after Soto *et al.*, 2007).

Fig. 9: Geological map of the Pyrenean Belt with locations of the structural maps and schematic reconstructions presented in Figs. 10, 11 and 12 as the boxes outlined in red and of the wells from the Southern Aquitaine basins presented in Fig. 14 (modified after Ducoux *et al.*, 2019).

Fig. 10: a) Geological map of the Eastern Mauléon/Chaînons Béarnais Basin with orientations of the magnetic lineations after bedding correction retrieved by AMS studies. b) Schematic model proposing the development of the Albian rifting phase in the Mauléon Basin under a sinistral strike-slip regime and illustrating the possible orientation of the magnetic lineations before the Cenozoic compression. Both images are modified after Oliva-Urcia *et al.*, 2010.

Fig. 11: a) Schematic representation of the progressive development of the North Pyrenean basins between the Middle Albian and the Early Cenomanian by gradual connection of isolated pull-apart basins to form a single sedimentary trough in a left-lateral transtensional regime (modified after Debroas, 1995, 2003). b) Schematic structural model illustrating the opening of the rift basins

exposed in the Bigorre region as sinistral pull-apart basins during the Albian-Cenomanian (modified after Debroas, 1990).

Fig. 12: a) Schematic representation of faulting relationships in the Pre-Alpine basement of the Pyrenean Axial Zone and in the Mesozoic series of the South Pyrenean basins, with illustration of the extension directions (grey arrows) during the Aptian-Albian transtensional interval retrieved by structural and AMS studies (redrawn and modified after Oliva-Urcia *et al*., 2011). b) Schematic map view representation of the structural patterns related to the Early Cretaceous extensional phase on the southern margin of the Organyà Basin, showing the general orientation of the main extension-related structures such as normal faults and joints/veins (modified after Tavani *et al*., 2011a).

Fig. 13: a) Locations of the ECORS-Bay of Biscay seismic reflection profile across the Parentis Basin and of the boreholes used for its calibration (redrawn and modified after Bois *et al*., 1997). b) Line drawing and c) interpretation of the central portion of the ECORS-Bay of Biscay seismic reflection profile (modified after Jammes, 2009). Note the overall flower structure of the Parentis Basin in cross section, with most of the (reactivated) extensional structures converging towards the Ibis Fault. d) Tectonic subsidence curves and tectonic subsidence velocities corrected for sea-level and water depth variations for three oil wells (Ibis, Fregate and Cormoran) and one synthetic well (Parentis West S. B.; see Brunet, 1991) (redrawn and modified after Brunet, 1997). Upper curves represent tectonic subsidence under water with local isostatic compensation, while lower curves represent basement deepening curves with decompacted sediments (see Brunet, 1997 for details on applied corrections). Note that during the Late Jurassic-Early Cretaceous rifting, tectonic subsidence is noticeably faster in the Early Cretaceous interval than in the Late Jurassic interval.

Fig. 14: a) Simplified geological map of the Aquitaine Basin with locations of the main Late Jurassic-Early Cretaceous sub-basins reviewed in this study and of the wells presented in (b) (redrawn and modified after Bourrouilh *et al*., 1995). Abbreviations: G.R.H. = Grand-Rieu High; N.P.F. = North Pyrenean Fault; N.P.F.T. = North Pyrenean Frontal Thrust; N.P.Z. = North Pyrenean Zone. b) Subsidence curves for the Lacq 1 (Arzacq Basin) and Ger 1 (Tarbes Basin) wells from the Southern Aquitaine Basin (redrawn and modified after Brunet, 1984; see Fig. 9 for location). Upper curves are tectonic subsidence curves with local isostatic compensation; lower curves represent progressive deepening of the Paleozoic basement with decompacted sediments (see Brunet, 1984 for details on the applied corrections). Note that during the Late Jurassic-Early Cretaceous rifting, tectonic subsidence curves are steeper for the Early Cretaceous interval than for the Late Jurassic one.

Fig. 15: Schematic diagram showing the main tectonic phases affecting the basins reviewed in this work during the Late Jurassic-Early Cretaceous interval. During the Late Jurassic, extension seems to have initially affected the most external parts of the system and then progressively affected the inner portions. During this phase, the Northern Aquitaine basins and the Asturian Basin, located at the boundaries of the system, underwent oblique rifting, while the rest of the system underwent orthogonal rifting. All the basins were affected by an episode of emersion and/or reduced subsidence during the Neocomian (~145 to 130 Ma). After the Neocomian, most of the rift basins experienced a phase of renewed subsidence controlled by oblique transtensional rifting, with the notable exception of the North Pyrenean basins. This phase coincided with crustal breakup in the inner Bay of Biscay margins. When oceanic spreading affected the Bay of Biscay domain during the Albian, most of the rift basins became inactive and transtensional rifting accommodating the

displacement between Iberia and Eurasia localized along the Basque-Cantabrian and North Pyrenean systems. Rifting finally terminated during the Cenomanian.

Fig. 16: Kinematic reconstruction of the Iberia/Eurasia diffuse plate boundary between the Late Jurassic and the Early Cretaceous, with representation of the rift domains reviewed in this study and their kinematic evolution through time. During the Late Jurassic (a) left-lateral displacement between the two plates concentrates on the edges of the system, while the rest of the plate boundary undergoes mainly orthogonal rifting. During the Neocomian (b) the whole plate boundary experiences a phase of reduced subsidence with localized exhumation events, while lithospheric breakup propagates northward on the Western Iberian margin. During the Aptian (c) most of the rift basins within the plate boundary zone are affected by left-lateral transtension controlling a phase of increased subsidence, with the notable exception of the North Pyrenean Rift System. During the Albian-Cenomanian (d) the Bay of Biscay realm undergoes oceanic accretion and the transtensional deformation localizes along the Basque-Cantabrian/North Pyrenean Rift System, which remains the only active domain while the rest of the sedimentary basins attain the post-rift stage. Plate kinematic model modified after Angrand *et al*. (2020). Acronyms: ACF, Ateca-Castellon Fault; AsB, Asturian Basin; ArS, Armorican Shelf; BCB, Basque-Cantabrian Basin; CAF, Celtaquitaine Flexure; CCRB, Catalan Coastal Range Basin; CoB, Columbrets Basin; HF, Hesperian Fault; IF, Ibis Fault; LF, Leitza Fault; MaB, Mauléon Basin; NAB, North Aquitaine basins; NIF, North Iberian Fault; NPB, North Pyrenean basins; PaB, Parentis Basin; SAB, South Aquitaine basins; SPB, South Pyrenean basins; VF, Ventaniella Fault.

Fig. 17: Schematic representation of (a) a narrow transform plate boundary and of (b) a diffuse transform plate boundary. In (a) the total left-lateral displacement between the two plates is accommodated by a single wrench structure. In (b) the very same amount of total displacement is

partitioned between a series of sub-parallel strike-slip structures, each of which accommodates a portion of the total left-lateral displacement between the stable parts of the plates.

Declaration of interests

☒The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: