TECTONICS, VOL. 30, TC2001, doi:10.1029/2010TC002735, 2011

The role of the Bay of Biscay Mesozoic extensional structure in the configuration of the Pyrenean orogen: Constraints from the MARCONI deep seismic reflection survey

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Received 28 April 2010; revised 18 October 2010; accepted 4 November 2010; published 3 March 2011.

[1] Seismic interpretation of the MARCONI deep seismic survey enables recognition of the upper crustal structure of the eastern part of the Bay of Biscay and the main features of its Alpine geodynamic evolution. The new data denotes that two domains with different Pyrenean and north foreland structures exist in the Bay of Biscay. In the eastern or Basque-Parentis Domain, the North Pyrenean front is located close to the Spanish coast, and the northern foreland of the Pyrenees is constituted by a continental crust thinned by a north dipping fault that induced the formation of the Early Cretaceous Parentis Basin. In the western or Cantabrian Domain, the North Pyrenean front is shifted to the north and deforms a narrower and deeper foreland basin which lies on the top of a transitional crust formed from the exhumation of lithospheric mantle along a south dipping extensional low-angle fault during the Early Cretaceous. The transition between these two domains corresponds to a soft transfer zone linking the shifted North Pyrenean fronts and a north- to WNW-directed thrust that places the continental crust of the Landes Plateau over the transitional crust of the Bay of Biscay abyssal plain. Comparison between this structure and regional data enables characterization of the extensional rift system developed between Iberia and Eurasia during the Late Jurassic and Cretaceous and recognizes that this rift system controlled not only the location and features of the Pyrenean thrust sheets but also the overall structure of this orogen. **Citation:** Roca, E., J. A. Muñoz, O. Ferrer, and N. Ellouz (2011), The role of the Bay of Biscay Mesozoic extensional structure in the configuration of the Pyrenean orogen: Constraints from the MARCONI deep seismic reflection survey, *Tectonics*, *30*, TC2001, doi:10.1029/2010TC002735.

1. Introduction

[2] The Pyrenees form a doubly vergent mountain belt as a result of Late Cretaceous-Cenozoic convergence between Iberian and Eurasian plates [*Choukroune and ECORS Team*, 1989; *Muñoz*, 1992, 2002]. They developed over a previous intracontinental thinned lithosphere in the east, and, more to the west, over the passive continental margin fringing the Bay of Biscay oceanic to transitional crust to the south [e.g., *Gallastegui*, 2000; *Gallastegui et al.*, 2002; *Thinon et al.*, 2003; *Sibuet et al.*, 2004a, 2004b]. As a result of this significant change in the preorogenic lithospheric configuration, the Pyrenees display different characteristics along strike (Figure 1) [*Muñoz*, 2002]. Between France and Spain, the Pyrenees s.s. form a continental collisional orogen with a limited subduction of the continental Iberian lithospheric mantle and lower crust underneath the Eurasian plate [*Muñoz*, 1992; *Beaumont et al.*, 2000; *Pedreira et al.*, 2003]. More to the west, the Cantabrian Pyrenees constitute a crustal wedge of continental rocks above the north directed subducting Iberian continental lithosphere, but facing northward the transitional to oceanic Bay of Biscay lithosphere, which shows a south directed subduction [*Pulgar et al.*, 1996; *Fernández-Viejo et al.*, 2000].

[3] Recording this structural change, the North Pyrenean thrust front shows different characteristics along strike. In the Pyrenees s.s., it is made up by basement-involved thrusts developed from the inversion of the extensional faults that bounded most of the main Early Cretaceous intracontinental basins inherited from the opening of the Bay of Biscay [*Muñoz*, 1992; *Gómez et al.*, 2002] to the north. In contrast, in the Cantabrian Pyrenees, the North Pyrenean front belongs to an imbricate thrusts system located at the toe of the present-day Cantabrian continental slope which deforms the sediments overlying the Bay of Biscay transitional to oceanic crust [*Derégnaucourt and Boillot*, 1982; *Alvarez-Marrón et al.*, 1995, 1996; *Gallastegui*, 2000; *Ayarza et al.*, 2004].

[4] In order to improve the structural knowledge of the transition between these two well-differentiated Pyrenean sectors, a deep seismic reflection survey was carried out in the eastern part of the Bay of Biscay aboard the Spanish R/V *Hesperides* in September 2003 (Figure 2) [*Gallart et al.*, 2004]. This survey, named MARCONI (North Iberian Continental Margin), included the acquisition of 11 multichannel deep seismic reflection profiles of a total length of 2000 km, and the signal record in a network of 24 OBS/OBH instruments and 36 land stations (see methodology and results of the analyses of both data by *Ruiz* [2007] and *Fernández-Viejo et al.* [2010]). These data, together with the ESCIN and ECORS-Bay of Biscay deep seismic profiles and the available industrial seismic lines, permit to recognize the structure of this area and characterize its evolution since the latest Jurassic opening of the Bay of Biscay.

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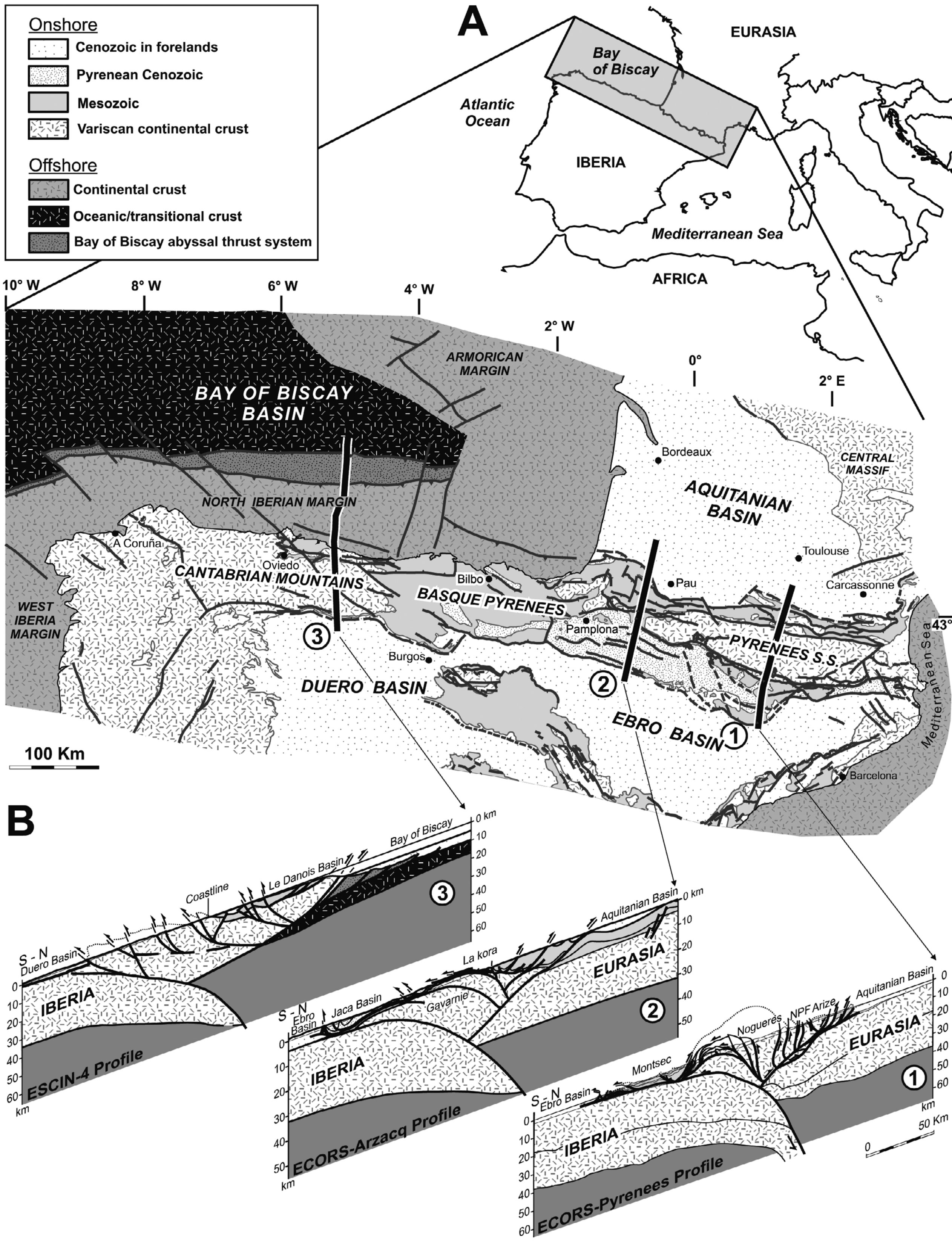

Figure 1. (a) Location map of the Pyrenees. (b) Simplified structural map of the Pyrenees and adjoining areas with three crustal sections that depict the main structural features and changes along this collisional belt: (1) ECORS-Pyrenees, (2) ECORS-Arzacq, and (3) ESCIN-4 (modified from *Muñoz* [2002]).

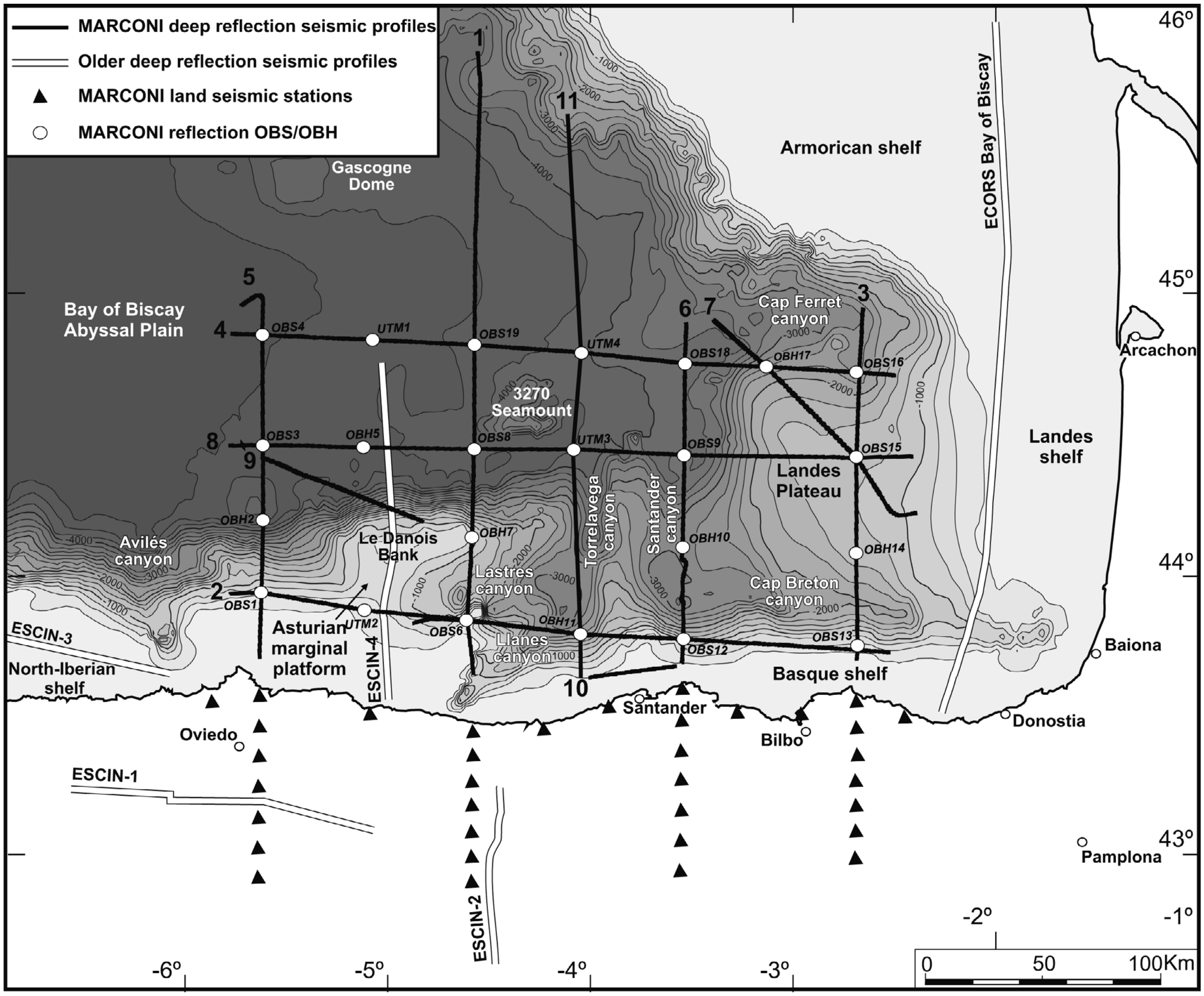

Figure 2. Bathymetric map of the eastern part of the Bay of Biscay with the location of the reflection seismic experiments carried out in the MARCONI project.

[5] The present paper aims to illustrate the main features of the North Pyrenean front and related foreland along the eastern Bay of Biscay from the description and interpretation of the MARCONI seismic profiles. With this, we seek not only to improve the knowledge of the structure of this area but also to decipher the origin of the change in the structural style from the Pyrenees s.s. to the Cantabrian Pyrenees, and thus, better understand the processes that governed the evolution of the Pyrenees. On the basis of this analysis, a kinematic and geodynamic scenario is proposed highlighting the role played by the preexisting Mesozoic structures not only in the geometrical and deformation features of the North Pyrenean front but also in the overall configuration of the Pyrenees.

2. Geological Setting

2.1. The Bay of Biscay

[6] The Bay of Biscay is an east-west oriented subsidiary arm of the North Atlantic Ocean located between the Iberian Peninsula and the western coast of France (Figure 1). It is a deep basin with a V shape opened to the west, this is to the Atlantic Ocean, formed between the latest Jurassic and Santonian times [*Montadert et al.*, 1979a; *Boillot*, 1984; *Vergés and García-Senz*, 2001; *Sibuet et al.*, 2004b].

[7] The bathymetry (Figure 2) and the geophysical studies carried out in this basin show that it is made up of: a western part represented by a 4–5 km deep abyssal plain floored by a 13–16 km thin crust of transitional or oceanic nature [*Gallastegui*, 2000; *Gallastegui et al.*, 2002; *Thinon et al.*, 2003; *Sibuet et al.*, 2004a, 2004b; *Pedreira*, 2004; *Ruiz*, 2007]; and an eastern part integrated by a shallower plateau (1–2 km deep) developed over a 15–25 km thick continental crust [*Pinet et al.*, 1987; *Ruiz*, 2007]. Both parts are bounded northeastward and southward by two well-differentiated continental margins. The northeast one, the Armorican Margin, is a Mesozoic extensional passive margin with southwest directed listric extensional faults [*Montadert et al.*, 1974; *Derégnaucourt and Boillot*, 1982; *Le Pichon and Barbier*, 1987; *Thinon*, 1999; *Thinon et al.*,

2003]. In contrast, the southern one (the North Iberian Margin) is part of the Pyrenean orogen. It consists of a Late Cretaceous-Cenozoic thrust system that overthrusts the continental crust of the Iberian plate on top of the transitional to oceanic crust of the Bay of Biscay abyssal plain westward [*Sibuet et al.*, 1971; *Boillot*, 1984; *Alvarez-Marrón et al.*, 1996; *Gallastegui et al.*, 2002; *Ayarza et al.*, 2004, *Pedreira*, 2004], and on top of the thinned continental crust of the Landes Plateau eastward [*Cámara*, 1997; *Gómez et al.*, 2002; *Ferrer et al.*, 2008]. This thrust system developed on the former southern passive margin of the Early Cretaceous Bay of Biscay basin and mainly consists of north directed basement-involved thrusts and related folds [*Alvarez-Marrón et al.*, 1995; *Cuevas et al.*, 1999; *Gallastegui*, 2000; *Gómez et al.*, 2002; *McDougall et al.*, 2009].

2.2. Alpine Plate Kinematics of the Pyrenean Realm

[8] The Bay of Biscay basin together with the Pyrenean orogen formed during Alpine times at the boundary between the Iberian and Eurasian plates. Their formation and later evolution, therefore, is strongly controlled by the relative motion of these plates which is itself governed by the opening history of the Atlantic Ocean. This Alpine history began during Late Permian-Triassic times with the breakup of Pangea which led to the development of a rift along the future Atlantic Ocean, the rifting and opening of the Tethys, and the formation of a ESE trending rift (Pyrenean Rift) that connected these two rifts along the future Iberia-Eurasia boundary [*García-Mondejar*, 1989; *Ziegler*, 1990]. The ongoing extension along the Atlantic Rift resulted in the opening of the Central Atlantic Ocean which started to form in its southern parts during the Jurassic and propagated northward through the Cretaceous times [*Savostin et al.*, 1986; *Srivastava et al.*, 1990; *Tucholke et al.*, 2007]. In the course of this propagation, the development of the Newfoundland-Iberia Rift between Iberia and North America resulted, first, into the thinning of the continental lithosphere (Late Jurassic-earliest Cretaceous), then, the exhumation of continental mantle (Barremian-Aptian) and finally, the breakup of the continental lithosphere by the Aptian-Albian boundary [*Tucholke et al.*, 2007]. In the western Bay of Biscay mantle exhumation to oceanic accretion occurred since 125 Ma (Barremian) as evidenced by the M3 and younger identified magnetic lineations [*Williams*, 1975; *Le Pichon and Barbier*, 1987; *Thinon et al.*, 2002; *Sibuet et al.*, 2004b]. This process endured into the development of a transtensional to extensional plate boundary between Eurasia and Iberia which, uncoupled, rotated 35–37° anticlockwise with respect to Eurasia [*Van der Voo*, 1969; *Srivastava et al.*, 1990; *Olivet*, 1996; *Gong et al.*, 2008].

[9] From late Santonian, a faster opening of South Atlantic Ocean produced the northward drift of Africa and, as a consequence, the convergence between the then uncoupled Iberia and Eurasia plates [*Rosenbaum et al.*, 2002]. This drastic change in the relative motion of Iberia generated the buildup of the collisional Pyrenean orogen along the then formed Iberia-Eurasia extensional boundary [*Boillot and Malod*, 1988; *Muñoz*, 1992, *Alvarez-Marrón et al.*, 1995; *Vergés et al.*, 2002]. This orogen developed until middle Miocene when Iberia and Eurasia welded and their relative motion ceased or became imperceptible [*Srivastava et al.*, 1990; *Roest and Srivastava*, 1991; *Rosenbaum et al.*, 2002].

3. Pyrenean and North Foreland Structure in the Eastern Part of the Bay of Biscay

[10] The area covered by the MARCONI survey (study area) includes the eastern parts of the Bay of Biscay abyssal plain, the shallow continental Landes Plateau and the adjoining contractional southern margin of the Bay of Biscay (Figure 2). In this area, the MARCONI seismic profiles depict, rather well, the main structural features of the extensional fault system related to the formation of the Bay of Biscay as well as those of the North Pyrenean contractional front. From their interpretation, two domains with a different Pyrenean and north foreland structure have been recognized (Figure 3): the eastern, Basque-Parentis Domain and the western, Cantabrian Domain.

3.1. Basque-Parentis Domain

[11] This domain embraces the deep Landes Plateau and the adjoining Landes, Armorican and Basque shelves (Figures 2 and 3). In this sector, the Bay of Biscay is floored by a thinned continental crust [*Pinet et al.*, 1987; *Gallart et al.*, 2004; *Ruiz*, 2007]. Depth of the Moho decreases northward from about 30–35 km in the Basque shelf to 18–22 km bellow the east trending Cap Ferret canyon. North of this canyon, Moho depth increases suddenly reaching again 30–36 km at the Armorican shelf [*Roberts and Montadert*, 1980; *Tomassino and Marillier*, 1997; *Thinon et al.*, 2003]. Minimum crustal thickness, consequently, occurs in the northern part of the plateau where the major Late Jurassic-Albian Parentis Basin is located. Here, beneath the Parentis basin fill, the crust is less than 10 km thick and decreases westward from 7 km in the ECORS Bay of Biscay [*Pinet et al.*, 1987; *Tomassino and Marillier*, 1997] to 6–5 km in the MARCONI-3 profile [*Gallart et al.*, 2004; *Ruiz*, 2007].

3.1.1. Pyrenean Structure

[12] The North Pyrenean thrust front, in the Basque-Parentis Domain, runs along the Basque shelf close to the coast (Figure 3). It consists of north directed basement-involved thrusts and cover thrust slices [*Soler et al.*, 1981; *Sánchez*, 1991; *Bois and Gariel*, 1994; *Cámara*, 1997; *Gómez et al.*, 2002, *McDougall et al.*, 2009] whose location and geometry are controlled by the inversion of the extensional faults that bounded the Basque-Cantabrian Basin to the north (Figure 3). This basin spans over most of the Basque Pyrenees (Figure 3) and is bounded southward by the present-day South Pyrenean frontal thrust which depicts a low-angle ramp geometry [*Serrano et al.*, 1989; *Martínez-Torres*, 1993; *Hernaiz et al.*, 1994; *Pedreira*, 2004]. In the basin margins, the thick basin fill succession is pierced by Lower Cretaceous to Santonian diapirs of Upper Triassic salt [*Brinkmann et al.*, 1967; *Serrano and Martínez del Olmo*, 1990, 2004; *Klimowitz et al.*, 1999] which, squeezed during the Pyrenean contractional event, also controlled the

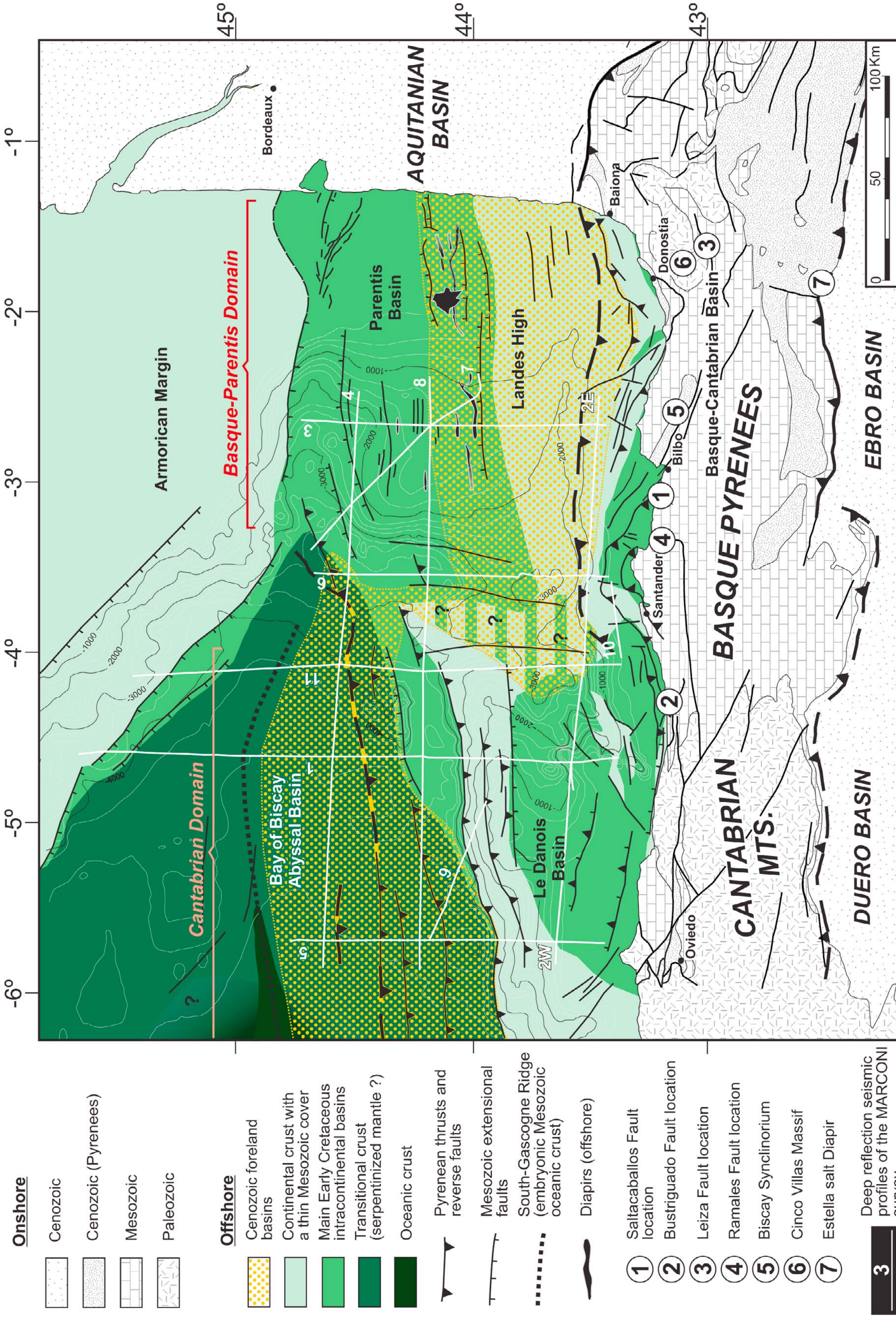

Figure 3. Structural map of the eastern part of the Bay of Biscay showing the offshore distribution of the North Pyrenean foreland basin, the major faults, and the main Cretaceous basins formed during the opening of the Bay of Biscay.

location and geometry of the Pyrenean thrusts [*Mathieu*, 1986; *García-Mondejar*, 1987; *Gómez et al.*, 2002].

[13] The Basque-Cantabrian Basin is filled by up to a 12.5 km thick succession of Upper Jurassic-Cretaceous sediments with interlayered Aptian to Santonian basic volcanic rocks [*Azambre and Rossy*, 1976; *Castañares and Robles*, 2004; *Cuevas et al.*, 1999; *García-Mondéjar et al.*, 2004; *Floquet*, 2004]. This basin was floored by an extremely thinned lithosphere in its central parts (Biscay Sinclinorium and Cinco Villas Massif) and affected by a Late Cretaceous thermal metamorphism [*Golberg and Leyreloup*, 1990; *Cuevas and Tubía*, 1999]. Lithosphere thinning resulted into the exhumation of the upper mantle at the floor of the basin as demonstrated by the outcropping mantle rocks (Leiza Fault [*Mendia and Gil-Ibarguchi*, 1991]). The prominent gravimetric and magnetic anomaly south of Bilbo [*Sibuet et al.*, 2004b, *Pedreira et al.*, 2007] is consistent with the presence of mantle rocks below the sedimentary infilling of the basin. Moreover, lower crustal rocks have been found in the Estella Diapir [*Pflug and Schöll*, 1976] which is located at the SE corner of the basin (Figure 3).

[14] In the MARCONI-3 profile, the North Pyrenean front has been imaged by a thrust wedge (Figures 4 and 5). This thrust wedge is located south of the Cap Breton canyon between SP 50 and 100 and demonstrates 2 km of shortening. Thickness and reflector dip variations in the related foreland syncline indicates its development between the late Eocene and early Miocene [*Ferrer et al.*, 2008].

3.1.2. North Pyrenean Foreland Structure

[15] North of the North Pyrenean front, a 65 km wide North Pyrenean foreland basin is clearly illustrated in the MARCONI-3 profile by a southward thickening wedge of synorogenic deposits (latest Cretaceous to early Miocene in age) (Figure 4). The northern boundary of this wedge is located over the Txipiroi High which represents the forebulge of the North Pyrenean foreland basin. North of the Txipiroi High, uppermost Cretaceous-Cenozoic sediments are also present but they do not follow the pattern of a flexured basin: they include several thicker depocenters whose development has been related to the inherited structure and the thermal subsidence generated during the latest Jurassic-Albian extensional phase [*Desegaulx and Brunet*, 1990, *Brunet*, 1994; *Ferrer et al.*, 2008].

[16] The MARCONI-3 profile also shows that, in this domain, the foreland uppermost Cretaceous-Cenozoic sediments are mostly undeformed. Cenozoic sediments are deeply incised by the Cap Breton and Cap Ferret submarine canyons (Figure 4) which initiated during the middle part of the Eocene [*Schoeffler*, 1965; *Cirac et al.*, 2001; *Ferrer et al.*, 2008]. Beneath the uppermost Cretaceous-Cenozoic sedimentary pile, the underlying thinned continental crust appears formed by two distinct parts: a major Mesozoic basin to the north, the Parentis Basin, and a coeval structural high to the south, the Landes High.

[17] The Parentis Basin, striking east-west, connects westward with the Bay of Biscay abyssal plain (Figure 3). This is a 100 km wide extensional basin filled by a thick (near 10 km) sequence of synrift Upper Jurassic-Lower Cretaceous carbonate to terrigenous rocks that overlie Lower to Middle Jurassic carbonates, lowermost Jurassic to Upper Triassic evaporites and Lower Triassic-Permian detrital rocks [*Dardel and Rosset*, 1971; *Mathieu*, 1986; *Bourrouilh et al.*, 1995; *Bois et al.*, 1997].

[18] Along the MARCONI-3 profile (Figure 6), the Parentis Basin appears as bounded southward by a major downward concave kinked north dipping planar fault with a normal displacement of near 20 km. This fault generates the total omission of the upper and middle crustal levels beneath the Parentis basin depocenter as attested in the ECORS-Bay of Biscay profile where the Parentis basin fill overlies a very thin lower crust (7 km thick). This lower crust shows high velocities that have been related to intruded mafic rocks [*Tomassino and Marillier*, 1997] and could be correlated to the mafic lower crust described in the Pyrenees and Newfoundland-Iberian Rift [*Reston*, 2007; *Péron-Pinvidic and Manatschal*, 2009 *Jammes et al.*, 2009].

[19] Above this crustal detachment fault, the Mesozoic Parentis basin fill is deformed by large wavelength ridges that, cored by Upper Triassic evaporites, are cut by extensional faults and, locally, pierced by squeezed diapirs made up by the same Triassic rocks [*Curnelle and Marco*, 1983; *Mathieu*, 1986; *Mediavilla*, 1987; *Masse*, 1997; *Biteau et al.*, 2006; *Ferrer et al.*, 2008]. These salt tectonic structures also affect the overlying uppermost Cretaceous to lower Miocene synorogenic deposits which are folded upon them (Figure 6). On the basis of the geometric relationships between the sedimentary successions and thickness variations, the formation of the Parentis Basin salt structures has been dated as Albian-Late Cretaceous and their later contractional squeezing as Oligocene-early Miocene [*Ferrer et al.*, 2008]. In addition to this minimal contractional deformation, the profile also shows a minor pop-up affecting the pre-upper Miocene deposits close to the southern basin boundary (Figure 6). This pop-up records the last contractional Pyrenean deformations and resulted from the reactivation of faults and diapirs already formed in this basin boundary [*Ferrer et al.*, 2008].

[20] The Landes High is located between the Parentis Basin and the Basque-Cantabrian Basin (Figure 4). It belongs to a plateau uplifted and eroded during Early Cretaceous [*Robles et al.*, 1988], in which the thick infill of the North Pyrenean foreland basin unconformably overlies the Variscan basement or a thin older Mesozoic cover [*Gariel et al.*, 1997; *McDougall et al.*, 2009]. North of the Pyrenean front, this thin cover does not show evidence of significant deformation except in some localized areas where it appears deformed by gentle synclines truncated by the overlying sediments of the North Pyrenean foreland basin infill [*Derégnaucourt*, 1981; *Ferrer et al.*, 2008].

[21] As a summary, the MARCONI-3 profile shows the main structural features of the easternmost part of the Bay of Biscay (Basque-Parentis Domain) which is characterized by the existence of the deep Late Jurassic-Early Cretaceous extensional basin (the Parentis Basin) bounded southward by a crustal detachment fault. This fault dips to the north and truncated the whole crust thinning it down to 7 km. The North Pyrenean front is located close to the Spanish coast and the only evidences of the Pyrenean contractional deformation

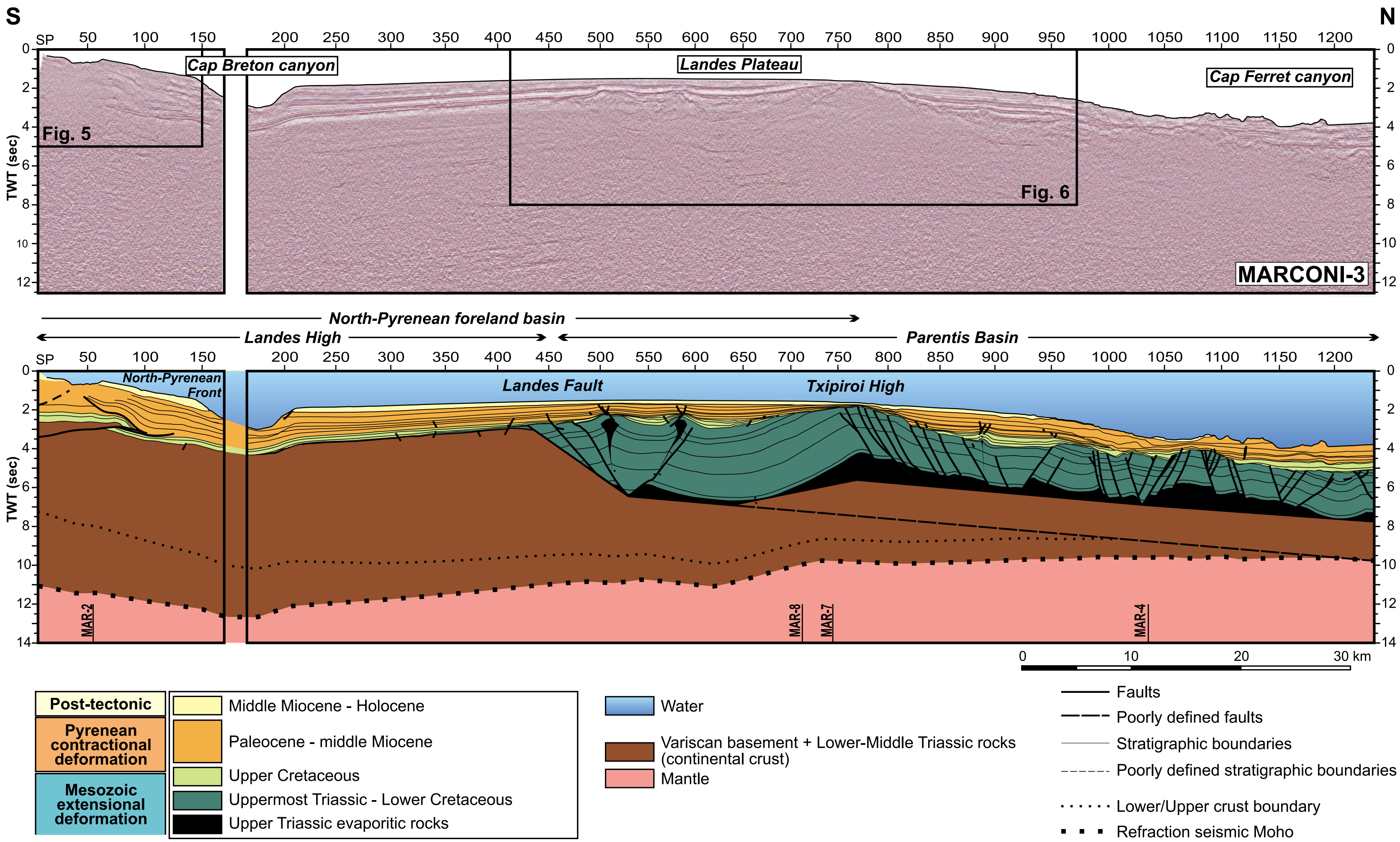

Figure 4. Unmigrated MARCONI-3 deep seismic profile with line drawing interpretation showing the main geological features of the upper crust in the Basque-Parentis Domain (modified from *Ferrer et al.* [2008]). Refraction seismic Moho boundary is compiled from *Muñoz* [2002], *Pedreira* [2004] and *Ruiz* [2007]. See Figure 2 for location.

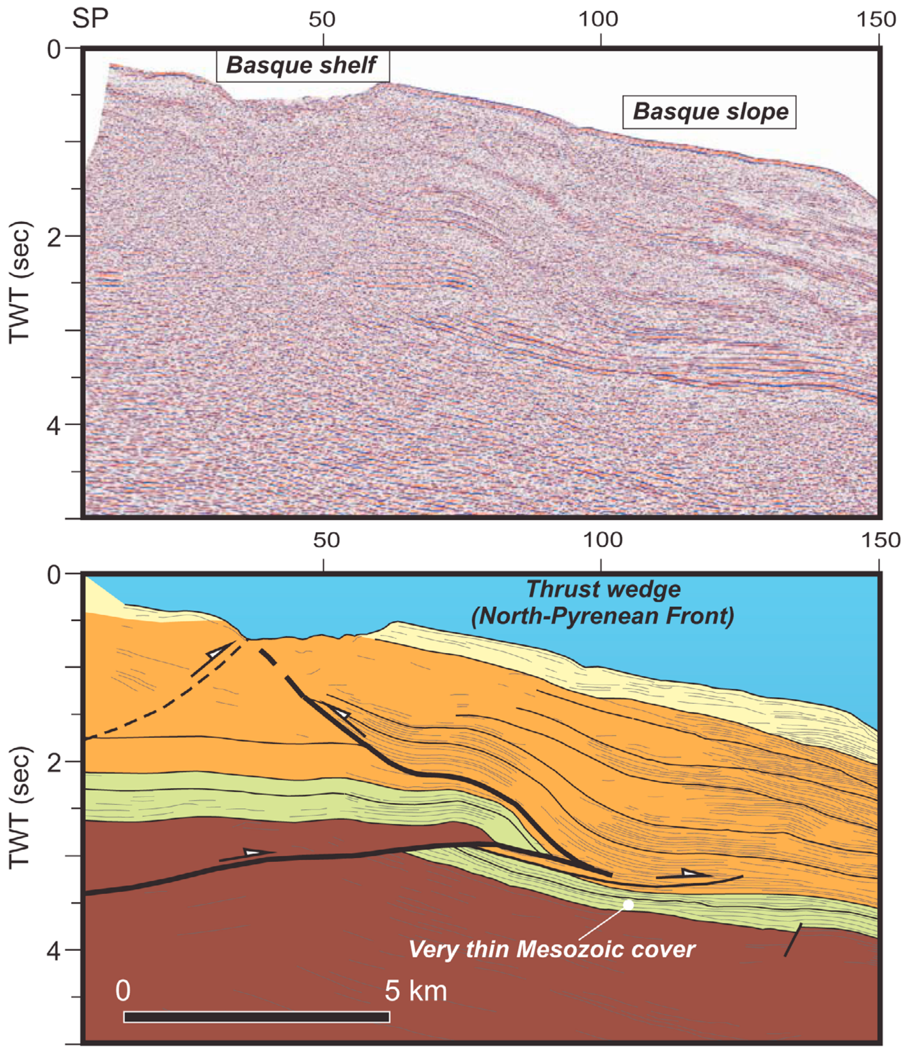

Figure 5. Southernmost part of the unmigrated MARCONI-3 seismic profile with line drawing interpretation illustrating the structure of the North Pyrenean front (modified from *Ferrer et al.* [2008]). See Figure 4 for location and legend.

in its foreland are related to the squeezing of the salt structures of the Parentis Basin.

3.2. Cantabrian Domain

[22] This domain includes the Bay of Biscay areas floored by the deep abyssal plain and the adjoining Armorican and North Iberian shelves (Figures 2 and 3). This last shelf is significantly wider than in the Basque-Parentis Domain and embraces, from South to North: a shallow coastal shelf, a relatively deep marginal platform (the Asturian marginal platform) and a striking morphological high named Le Danois Bank (Figure 2). North of this bank, there is a narrow and steep continental slope that depicts a height difference close to 4000 m and gradients always greater than 15°, values clearly higher than in the Basque-Parentis Domain (Figures 2 and 7).

[23] The continental crust beneath the North Iberian shelf has been affected by the north Pyrenean contractional structures, but yet its thickness decreases from near 27–30 km along the coastline to 20 km at the toe of the continental slope [*Gallart et al.*, 1997; *Pulgar et al.*, 1996; *Fernández-Viejo et al.*, 1998; *Ruiz*, 2007], emphasizing a crustal thinning of the Cantabrian margin during the Late Jurassic-Early Cretaceous rifting. More to the north, in the Bay of Biscay abyssal plain, the crust of the abyssal plain is much thinner with crustal velocities of an oceanic or transitional crust [*Limond et al.*, 1974; *Fernández-Viejo*, 1997; *Ruiz*, 2007]. In this area, the refraction seismic Moho is located at a depth of 10–16 km and increases its depth up to 20 km at the toe of the Cantabrian continental slope [*Bacon et al.*, 1969; *Limond et al.*, 1974; *Roberts and Montadert*, 1980; *Ruiz*, 2007]. If the water and the basin-fill sediments are removed, the crust appears slightly tilted to the south with a rather constant crustal thickness of 5–8 km [*Gallastegui*, 2000; *Ruiz*, 2007].

3.2.1. Pyrenean Structure

[24] In the Cantabrian Domain, the structure of the northern frontal part of Pyrenees is well-recorded by the

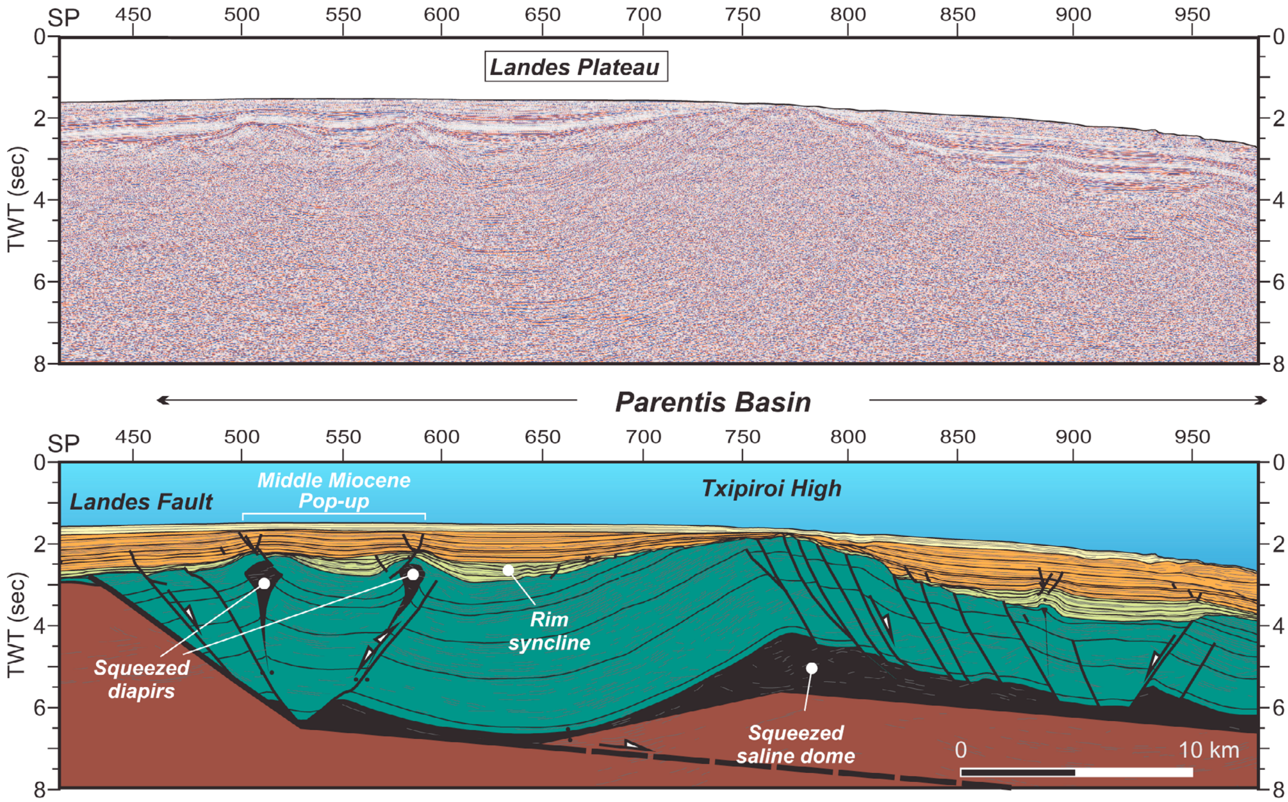

Figure 6. Main geological features of the Parentis Basin along the central part of the unmigrated MARCONI-3 seismic profile. The seismic profile and underlying line drawing interpretation depict the Mesozoic structure of the southern margin of the Parentis Basin as well as the squeezing of the diapirs and later inversion of some preexistent faults during the Pyrenean compression (modified from *Ferrer et al.* [2008]). See Figure 4 for location and legend.

MARCONI-1, 5 and 11 deep seismic reflection profiles. These profiles show that the North Pyrenean front is located 40–70 km north of the toe of the continental slope (Figures 7 and 8), and, therefore, clearly shifted to the north in relation to its position in the Basque-Parentis Domain (Figure 3). South of this front, the submerged part of the Pyrenean orogen appears to be constituted by two zones with a well-differentiated contractional structure: the North Iberian shelf plus the Cantabrian slope and the thrust system developed in the abyssal plain (Figures 7 and 8).

3.2.1.1. North Iberian Shelf and Cantabrian Continental Slope

[25] Along the MARCONI profiles, the reflectivity in the North Iberian shelf as well as in the continental slope is very poor: the seismic noise is noticeable and there are a lot of multiple reflections of the shallow and irregular seabed. However, the MARCONI-1 profile shows a 2–3 TWT sec thick reflective package beneath the platform that has been interpreted as the eastern prolongation of the Le Danois Basin (Figure 7). The structure of this basin just west of this profile has been imaged by hydrocarbon exploration seismic data [*Riaza Molina*, 1996; *Gutiérrez Claverol and Gallastegui*, 2002; *Pérez-García et al.*, 2009]. These data show that Le Danois Basin is a Late Jurassic-Early Cretaceous half-graben bounded southward by north dipping extensional faults. North of its depocenter the synrift sediments progressively thin northward and are affected by south dipping extensional faults. At present the Le Danois Basin shows a pop-up structure which resulted from the inversion of the extensional faults [*Boillot et al.*, 1979; *Riaza Molina*, 1996; *Alvarez-Marrón et al.*, 1997; *Gallastegui*, 2000; *Gallastegui et al.*, 2002]. This inversion initiated during the Senonian and mainly developed during the Eocene-early Miocene as attested by the formation of an intra-Senonian unconformity and the contractional deformation of the pre-middle Miocene sediments on the basin margins [*Gutiérrez Claverol and Gallastegui*, 2002; *Pérez-García et al.*, 2009].

[26] In the MARCONI-1 profile, the reflective package, correlated with the Le Danois basin infill, also thins toward the north. At the northern basin margin, the basin infill is deformed by north verging folds and south dipping thrusts placed in the footwall of a south dipping extensional fault (Figure 7). This extensional fault only cuts the lower part of the Mesozoic succession. In contrast, the folds and thrusts also affect the lower part of the overlying Cenozoic sediments but not the upper part of them which, more to the west, have been dated as upper Miocene-Holocene [*Gallastegui*, 2000; *Gutiérrez Claverol and Gallastegui*, 2002]. These features together with the thinner Mesozoic thickness in the footwall of the extensional fault suggest that

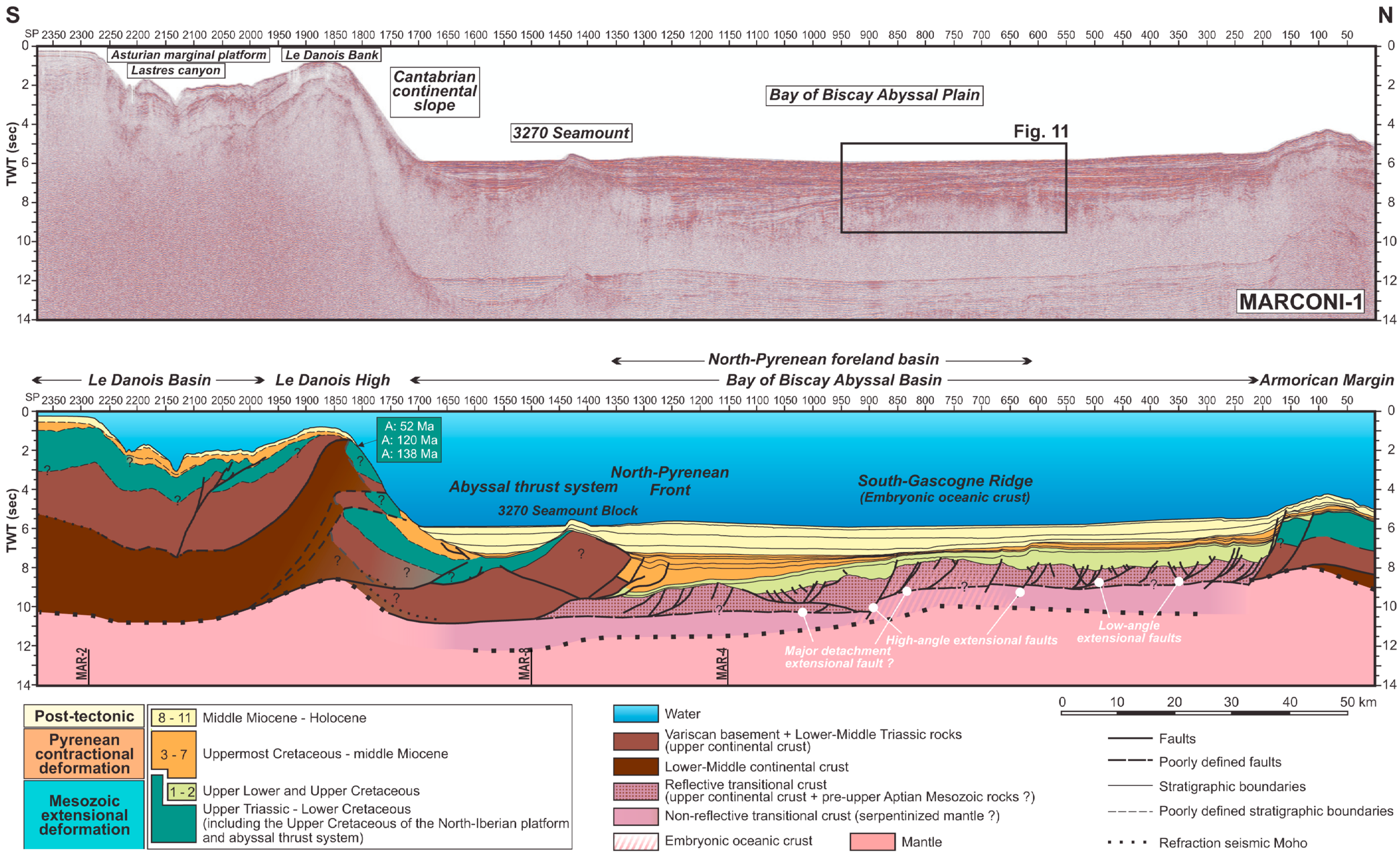

Figure 7. Unmigrated MARCONI-1 deep seismic profile with line drawing interpretation showing the main geological features of the crust in the eastern part of the Cantabrian Domain. Cantabrian continental slope structure is derived from samples collected during dives and dredges [*Capdevila et al.*, 1980; *Malod et al.*, 1982]. The green rectangle located at the top of the Cantabrian continental slope indicates the approximate location of the granulite clasts with the apatite fission track ages (A) determined by *Fügenschuh et al.* [2003]. Refraction seismic Moho boundary and nature of the crust are derived from *Ruiz* [2007] data. See Figure 2 for location.

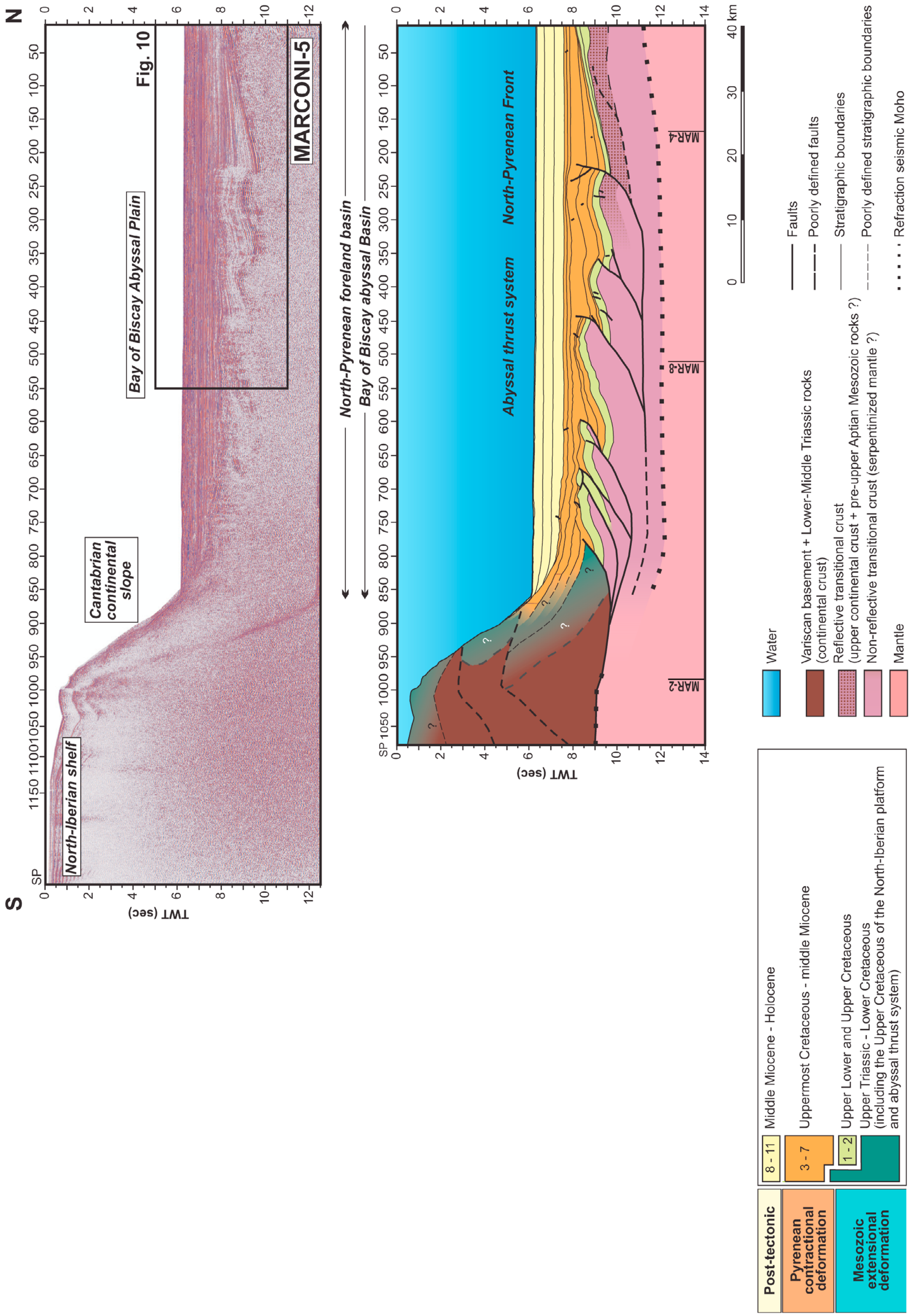
S
N
TWT (sec)
SP
North-Iberian shelf
Cantabrian continental slope
Bay of Biscay Abyssal Plain
Fig. 10
MARCONI-5
North-Pyrenean foreland basin
Bay of Biscay abyssal Basin
Abyssal thrust system
North-Pyrenean Front
MAR-2
MAR-8
MAR-4
40 km
Water
Variscan basement + Lower-Middle Triassic rocks (continental crust)
Reflective transitional crust (upper continental crust + pre-upper Aptian Mesozoic rocks ?)
Non-reflective transitional crust (serpentinized mantle ?)
Mantle
Faults
Poorly defined faults
Stratigraphic boundaries
Poorly defined stratigraphic boundaries
Refraction seismic Moho
Post-tectonic
Pyrenean contractional deformation
Mesozoic extensional deformation
8 - 11
3 - 7
1 - 2
Middle Miocene - Holocene
Uppermost Cretaceous - middle Miocene
Upper Lower and Upper Cretaceous
Upper Triassic - Lower Cretaceous (including the Upper Cretaceous of the North-Iberian platform and abyssal thrust system)

Figure 8

thrusts and related folds resulted from the Late Cretaceous-early Miocene (?) tectonic inversion of the extensional fault that developed at the northern margin of the Le Danois Basin during the Early Cretaceous.

[27] North of the North Iberian shelf, in the MARCONI profiles, the reflectivity in the Cantabrian continental slope is very poor and restricted to the first TWT sec (Figures 7 and 8). It mainly consists on short and discontinuous reflectors that dip less than the continental slope toward the north. Samples collected with submersibles denote, however, that this slope includes north directed thrusts affecting the Variscan basement and the synrift Upper Jurassic-Lower Cretaceous series [*Capdevila et al.*, 1980; *Malod et al.*, 1982]. Thermochronology studies performed in the synrift samples have supplied additional significant results on the structure and tectonic evolution of this area [*Fügenschuh et al.*, 2003]. Thus, granulites sampled in the Le Danois Bank have yielded apatite fission track (AFT) ages of 138, 120 and 52 Ma. These results demonstrate that Le Danois Bank was located in the footwall of major north dipping extensional faults that produced a significant Early Cretaceous (Berriasian-early Aptian) thinning of the continental crust. Thinning is attested by the exhumation of lower crustal rocks at the basin floor to supply pebbles to the Aptian-Albian breccia where the samples were collected [*Fügenschuh et al.*, 2003]. The young 52 Ma AFT age of one of these samples of granulites would suggest resetting of the Early Cretaceous ages recorded in the other collected samples which could be related with burial during earlier stages of thrusting, probably during Late Cretaceous, as described in the Central Pyrenees [*Metcalf et al.*, 2009].

[28] The existence of an Early Cretaceous north dipping extensional fault exhuming lower crust north of the Le Danois Basin could also be inferred from comparison of their location at the southern margin of the Bay of Biscay basin with the preserved West Iberian Margin [*Péron-Pinvidic and Manatschal*, 2009]. Therefore, thrusts in the Cantabrian slope would cut a preexistent major north dipping extensional fault and would passively transport the Le Danois Basin to the north. The age of these thrusts is difficult to discern because Cenozoic sediments have not been identified in the slope. However, we can infer they developed during Late Cretaceous-Oligocene to account for the AFT ages and the lithospheric flexure recorded by the southward thickening prism of the uppermost Cretaceous-Oligocene foreland sediments (units 3–5, Figures 7 and 8).

3.2.1.2. Abyssal Thrust System

[29] North of the Cantabrian continental slope, there is a 40–70 km wide system of north directed thrusts which, striking parallel to the slope, appears well-recorded in the MARCONI-1, 5 and 11 profiles. These profiles show that the structural features of this thrust system change substantially from east to west.

[30] Thus, in the east (profiles 1 and 11), it is constituted by a major thrust sheet and a narrow imbricate thrust system in its frontal part (Figures 7 and 9). The major thrust sheet is 8–14 km thick and 30–40 km wide and involves the lower part of the sedimentary infill of the Bay of Biscay abyssal basin as well as the underlying basement which has typical velocities of upper continental crust [*Ruiz*, 2007]. In the MARCONI-1 profile (Figure 7), the frontal part of this thrust sheet correlates with an east-west elongated high, the so-called 3270 Seamount, which is separated from the continental slope by a flat trough. The few diffuse reflectors observed inside the thrust sheet suggest that it corresponds to major south tilted basement block (the 3270 Seamount Block) overlain by south dipping Mesozoic sediments. These sediments thicken toward the Cantabrian continental slope and are affected by some minor north dipping extensional faults and, at the toe of the continental slope, by a system of north directed thrusts. According to these features, the major thrust sheet observed in the MARCONI-1 profile is interpreted as a Mesozoic extensional rider of upper continental crustal rocks passively transported by the Pyrenean sole thrust. This thrust is detached at the top of the high-velocity lower layer of transitional crust (serpentinized continental mantle) [*Ruiz*, 2007] and truncates in its front the autochthonous lower part of the Bay of Biscay basin fill and the upper part of the underlying basement.

[31] North of this major thrust, the MARCONI profiles also show, one (profile 1) or a set (profile 11) of small anticlines affecting the sedimentary units 3 to 5 (Figures 7 and 9). These anticlines verge northward and correspond to fault-bend folds related with an imbricate thrust system. The floor thrust of this system dips to the south and branches downward into the tip of the main thrust.

[32] The emplacement age of this imbricate thrust system as well as that of the major thrust is well defined by the growth strata and unconformities present in the sediments infilling the Bay of Biscay abyssal basin (Figures 7 and 9). From them, the emplacement of these thrusts can be dated as coeval with the sedimentation of the units 5 to 7 (upper Eocene?–middle Miocene); although the tilting and upward narrowing growth triangle geometry shown by the younger sediments close to the 3270 Seamount Block crest, suggest that the major frontal thrust continued being active up to the present, but involving a very small displacement. The growth strata and unconformities, on the other hand, also allow us to define a break-back thrusting sequence for the imbricate frontal thrusts in the profile 11 (Figure 9).

[33] On the contrary, in the western parts of the Cantabrian Domain (profile MARCONI-5; Figure 8) the abyssal thrust system consists of a north directed imbricated fan system of thrusts and related folds and not by a major thrust sheet. The width of the thrust sheets ranges between 5 and 20 km and involve the lower and middle units of the Bay of Biscay basin infill (units 1 to 7) as well as the upper part of a thinner

Figure 8. Unmigrated MARCONI-5 deep seismic profile with line drawing interpretation showing the main geological features of the crust in the western part of the Cantabrian Domain. Cantabrian continental slope structure is derived from samples collected during dives and dredges eastward in Le Danois Bank [*Capdevila et al.*, 1980; *Malod et al.*, 1982] and westward in the Ortegal Spur [*Boillot et al.*, 1987]. Refraction seismic Moho boundary and nature of the crust are derived from *Ruiz* [2007] data. See Figure 2 for location.

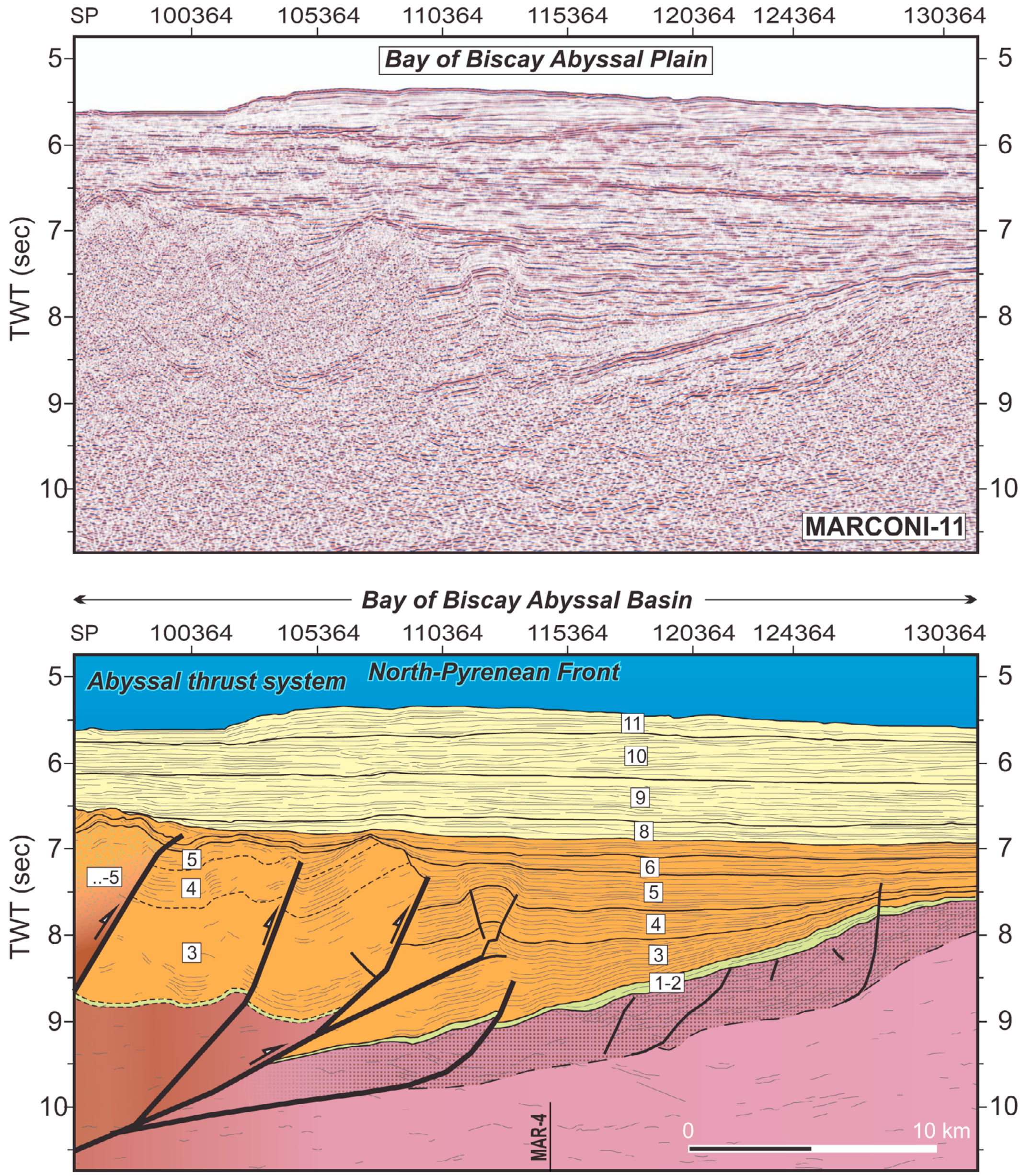

Figure 9. Detail of the unmigrated MARCONI-11 seismic profile with line drawing interpretation depicting the structure of the North Pyrenean front in the eastern part of the Cantabrian Domain. Labels 1–11 are Bay of Biscay abyssal basin fill differentiated units. See Figure 7 for legend.

transitional crust (5–8 km) (Figure 10). Regardless of these structural differences with the eastern parts of the Cantabrian Domain, the unconformities and growth strata geometries (Figure 10) denote that the contractional deformation has the same age as in the east (unit 5 to 7). However, these geometries display here a piggyback thrusting sequence; although the folding of the youngest syntectonic sediments close to the Cantabrian slope toe (Figure 8) evidences a reactivation of the most inner thrusts at the last stages of the contractional deformation.

3.2.2. North Pyrenean Foreland Structure

[34] The northern foreland of the Pyrenees in the Cantabrian Domain includes the central parts of the Bay of Biscay abyssal plain north of the thrust front (Figure 3). It is a foreland with a thick Upper Cretaceous-Cenozoic sedimentary succession that overlies a very thin (5–8 km) transitional to oceanic crust and, to the north, the thinned continental crust of the adjacent Armorican Margin.

3.2.2.1. The Crust of the Bay of Biscay Abyssal Plain

[35] The nature of the thin crust of the Bay of Biscay abyssal plain is not clear and has been widely debated. East of 6°W, the magnetic anomaly maps do not show any clear oceanic signature. Furthermore, the crust is quite thick, appears deformed by faults and includes an upper seismic layer with anomalous low velocities (<5.5 km/s) that overlies a lower seismic layer in which the velocities (7.2–7.3 km/s)

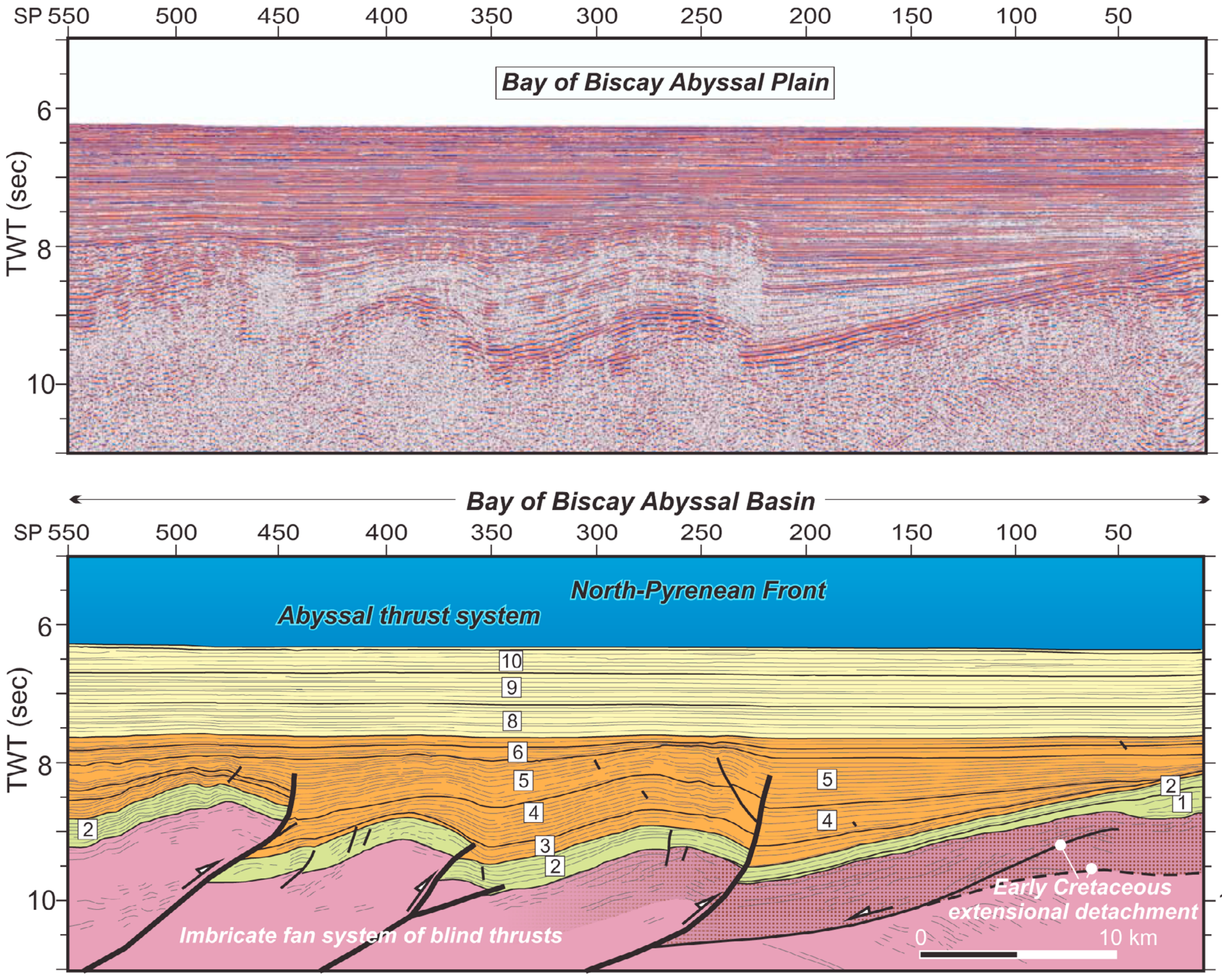

Figure 10. Northernmost part of the unmigrated MARCONI-5 seismic profile with line drawing interpretation illustrating the structure of the North Pyrenean front in the central part of the Cantabrian Domain. Labels 1–11 are Bay of Biscay abyssal basin fill differentiated units. See Figure 8 for location and legend.

[*Ruiz*, 2007] are higher than those observed westward in the oceanic layer 3 [*Limond et al.*, 1974]. For some authors, these features are compatible with a crust which continues being partial or totally oceanic [*Limond et al.*, 1974; *Boillot et al.*, 1979; *Sibuet and Collette*, 1993; *Alvarez-Marrón et al.*, 1995, 1996, 1997]; but, for others, they denote a transitional crust [*Derégnaucourt and Boillot*, 1982; *Gallastegui*, 2000; *Ruiz*, 2007]. This transitional crust would be in continuation with the transitional crust at the toe of the Armorican slope west of 6°W and would define a transition zone separating the oceanic crust from the Armorican thinned continental crust. An equivalent configuration has been recently described between the Atlantic oceanic crust and the thinned continental crust of the Newfoundland and West Iberian margins on the basis of ODP and seismic data [*Lau et al.*, 2006; *Tucholke et al.*, 2007; *Péron-Pinvidic and Manatschal*, 2009]. In these areas the transitional zone is formed by a "thin" (4–9 km) crust with riders of Mesozoic prerift and synrift sediments and continental crustal rocks that are extensionally detached over an exhumed continental mantle with seismic velocities comprised between 7.2 and 8 km/s [*Chian et al.*, 1999; *Lau et al.*, 2006; *Hopper et al.*, 2007; *Péron-Pinvidic and Manatschal*, 2009].

[36] Following these interpretations, the Bay of Biscay abyssal crust in the eastern part as well as in front of the surrounding margins is transitional. The positive magnetic lineation at the toe of the North Iberian continental slope would be related to the serpentinization of a subcontinental mantle [*Ruiz*, 2007], similar to the magnetic lineation observed at the toe of the Armorican shelf where the crust has been defined as transitional [*Montadert et al.*, 1979a; *Thinon et al.*, 2003].

[37] On the contrary, west of 6°W latitude, the recorded crustal seismic velocities and thickness look characteristic of an oceanic crust [*Limond et al.*, 1972, 1974] with a spreading center located between the symmetrical A34 magnetic lineations [*Verhoef et al.*, 1996; *Srivastava et al.*, 1990; *Sibuet et al.*, 2004b]. However, the region with low-amplitude magnetic anomalies, without linearity, at the bottom of the Armorican slope has been attributed to a transitional crust on

the basis of the high seismic velocities [*Barbier et al.*, 1986; *Thinon et al.*, 2003] and dredged granulites on the Goban Spur [*Didier et al.*, 1977]. Between this transitional crust and the oceanic one at the A34 anomaly, the basement flooring the region with the well-defined magnetic lineations with the V shape opened to the west could be normal oceanic crust or alternatively transitional crust, consisting of infiltrated exhumed continental mantle. Such a crust would not be possible to distinguish from a normal oceanic crust with the available data.

[38] Reconstructions of the North Atlantic at Chron M0 (125 Ma, earliest Aptian) illustrate the northward propagation of an oceanic spreading center from the Southeast Newfoundland Ridge [*Boillot et al.*, 1989; *Srivastava et al.*, 2000; *Péron-Pinvidic et al.*, 2007; *Tucholke et al.*, 2007]. The northern tip of an incipient oceanic crust would be located south of Galicia Bank and Flemish Cap and northward, the Newfoundland-Iberia Rift would be floored by thin continental crust and exhumed continental mantle [*Tucholke and Sibuet*, 2007, *Péron-Pinvidic et al.*, 2007]. In these reconstructions, the amount of extension in the Bay of Biscay also appears significantly smaller than in the portion of the Newfoundland-Iberia Rift located between Galicia Bank and Flemish Cap [*Sibuet et al.*, 2007a]. As a consequence, the crust of the Bay of Biscay showing magnetic anomalies older than M0 would probably also be transitional and mainly made up of exhumed continental mantle. This mantle would have experienced serpentinization during exhumation, a process that could even account for the formation of magnetic lineations [*Sibuet et al.*, 2007b]. If normal oceanic crust would have not formed until early Albian times, as proposed for the West Iberia Margin [*Jagoutz et al.*, 2007; *Tucholke et al.*, 2007] the Bay of Biscay oceanic crust would be restricted to a relatively narrow east-west strip centered in the A34 magnetic lineations at both sides of the spreading center.

3.2.2.2. Sedimentary Architecture

[39] In the Bay of Biscay abyssal plain, eleven seismic sedimentary units have been defined above the seismic acoustic basement based on the interpretation of the MARCONI profiles. They have a Cretaceous-Cenozoic age and some of them are bounded by unconformities as demonstrated by onlap surfaces and truncation of underlying reflectors (Figures 7, 9, and 11). These units can be grouped into four packages with well differentiated features.

[40] 1. A lowermost package (unit 1) of very variable thickness that belongs to the first identifiable sedimentary sequence overlying the acoustic basement. The lower boundary of this package is not well defined and it fills two major basins separated by a basement ridge striking along the axis of the Bay of Biscay abyssal plain (the South Gascogne Ridge of *Montadert et al.* [1971] and *Derégnaucourt* [1981]) (Figure 7). Within these two basins, the unit 1 appears formed by rather continuous reflectors that onlap the basement highs resulting from the motion of low-angle extensional faults. These faults cut the lower part of this unit (Figure 7) which depicts reflector sequences thickening into their hanging wall as well as internal unconformities and onlap surfaces (Figure 11) which denote a syndepositional faulting.

[41] 2. A lower thin package (unit 2) of rather constant thickness along the whole profiles that covers the basement highs. Seismically, it is made up of moderate to high-amplitude reflectors with a good continuity (Figures 7–12). The boundary between unit 2 and unit 1 is concordant except on the flanks of the South Gascogne Ridge where the near horizontal reflectors of unit 2 truncate the tilted reflectors of unit 1 which are parallel to the top of the acoustic basement (Figure 11).

[42] 3. A middle package, integrated by the labeled 3 to 7 units, which thickens toward the south (Figures 7–9). This thickening is very pronounced in unit 3 and decreases progressively in the overlying units. The northern boundary of this southward thickening wedge of synorogenic deposits corresponds to the main bulge of the North Pyrenean foreland basin which coincides with the location of the preexistent South Gascogne Ridge (Figures 7 and 11). North of this forebulge, the thickness of the package is significantly thinner and relatively constant (Figure 7). Accordingly to these geometric features, the intermediate package is interpreted as the main phase of the infill of the North Pyrenean foreland basin developed by the flexure of the Bay of Biscay lithosphere beneath the overriding Iberian plate. A question arises about the location of the depocenter of unit 3, and consequently, about the position of the North Pyrenean thrust front during its deposition. The leading thrust imbricate fan and related folds at the North Pyrenean thrust front developed during sedimentation of units 5 to 7 (Figures 7, 9, and 10). If the trailing thrust of such imbricate fan, which is the floor thrust of the main thrust sheet of the abyssal thrust system, has the same age, its hanging wall would involve a thick succession of units 3–4 and their depocenter should be located in the footwall of the Cantabrian slope thrust system, which would be the thrust front at that time (Figure 7).

[43] 4. An upper and almost undeformed package that unconformably drapes the preexistent structural topography. This package includes the units 8 to 11 (Figure 7) and, unlike the previous one, has a rather constant thickness except the lowermost part (units 8 and 9) that still depicts a progressive but gentle thinning toward the north. The reflectors of this upper package, much more continuous, are horizontal except in the northern parts of the Bay of Biscay abyssal plain where the lowermost sequences of the package appear deformed, mainly by the major normal fault that bounds the Bay of Biscay abyssal plain to the north (Figure 7). Also they onlap the folds developed during the sedimentation of the underlying package. As far as the thickness of this last posttectonic package is concerned, two considerations can be made entailing a set of implications on the nature and thermal behavior of the lithosphere in the Bay of Biscay. The first is that the thickness of the posttectonic package in the Cantabrian foreland does not differ to the one observed over the abyssal thrust system (Figure 7); and the second is that this thickness is 5 to 6 times thicker than over the Landes Plateau (Basque-Parentis Domain). From the first, we can deduce that, after Pyrenean deformation, subsidence has been similar in the whole abyssal plain which, considering that crustal thickness is not so different [*Ruiz*, 2007], implies a similar nature and thickness of the lithosphere underlying both the “undeformed” foreland and the

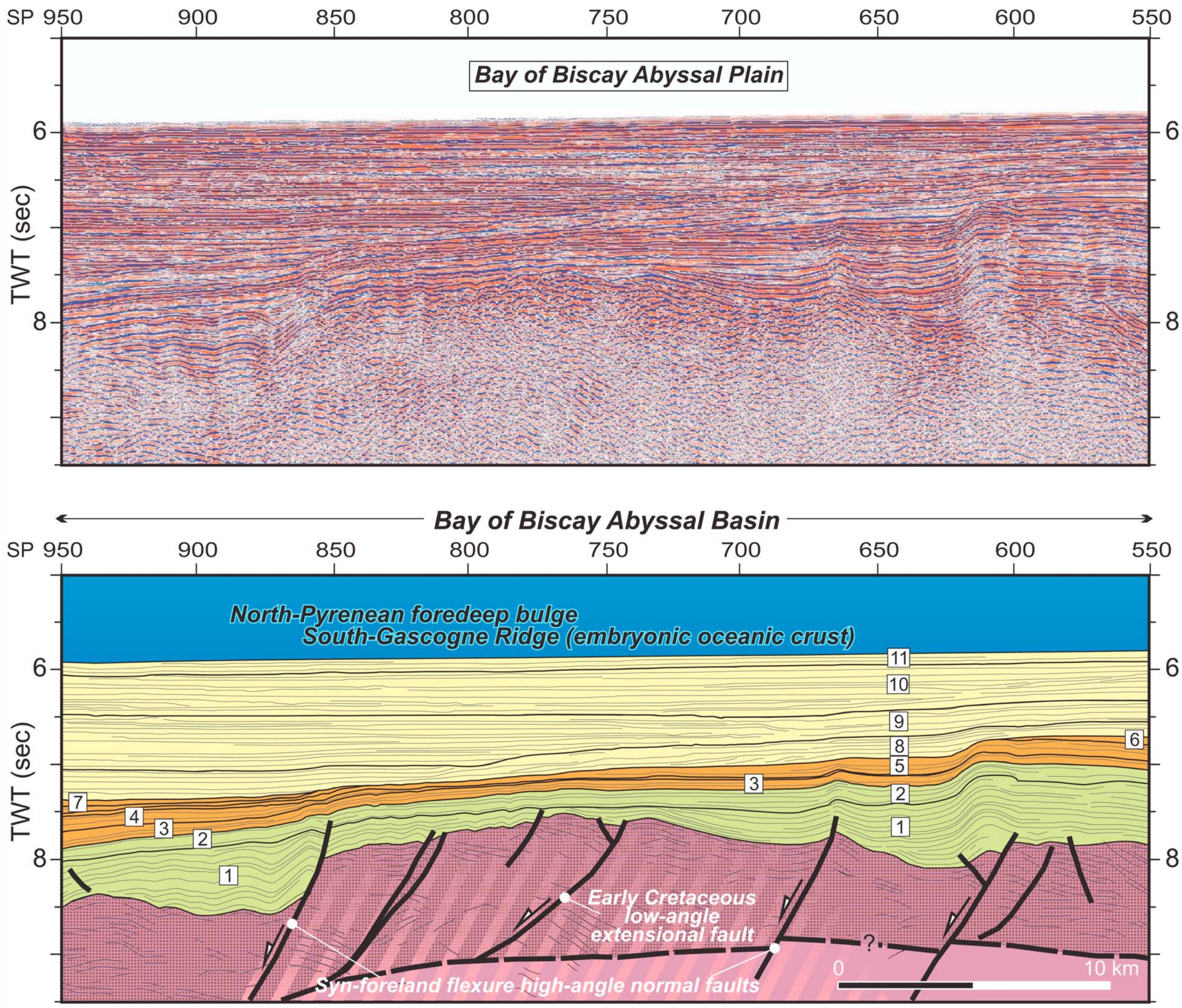

Figure 11. Detail of the unmigrated MARCONI-1 seismic profile showing the structure of the transitional crust and overlying Upper Cretaceous-Cenozoic sediments at the northern edge of the North Pyrenean foreland basin. Labels 1–11 are Bay of Biscay abyssal basin fill differentiated units. See Figure 7 for location and legend.

abyssal thrust system. From the second, we can infer that the lithosphere in the abyssal plain has been much hotter than in the Landes Plateau from the onset of the Pyrenean deformation.

[44] The age of these four packages is difficult to precise due to the lack of wells in the study area. However, from seismic facies and reflector correlations of MARCONI profiles with seismic surveys comprising the wells drilled westward (DSDP118 and DSDP-119 [*Laughton et al.*, 1972]) and northward at the toe of the Armorican Margin (IPOD 400, 401 and 402 of Leg 48 [*Montadert et al.*, 1979b]), we can infer that: the lowermost package would probably be Albian to Cenomanian in age, the lower package would be Late Cretaceous (Santonian or Campanian) in age, the middle package would be latest Cretaceous-middle Miocene in age, and the upper one post middle Miocene in age.

3.2.2.3. Structure of the North Pyrenean foreland

[45] The main features of the North Pyrenean foreland in the Cantabrian Domain are well depicted in the MARCONI-1 profile (Figure 7), although some of them are also visible in profiles 11 and 5 (Figures 8, 9, and 11). In these profiles, the basement of the North Pyrenean foreland is integrated by a transitional crust that depicts the presence of two major east trending basement highs bounding large-scale basins filled by the Albian to Cenomanian successions of the unit 1: one fringing the North Pyrenean thrust front to the north (Figures 7 and 8) and the other running along the axis of the Bay of Biscay abyssal basin (Figure 7). This last ridge, the South Gascogne Ridge, is well illustrated in the MARCONI-1 profile where it appears as a large wavelength but smooth high cored by an acoustic basement depicting a more chaotic and poorer reflectivity seismic signature. Over this basement high, the Albian to Cenomanian sediments of the unit 1 are nearly parallel to the top of the acoustic

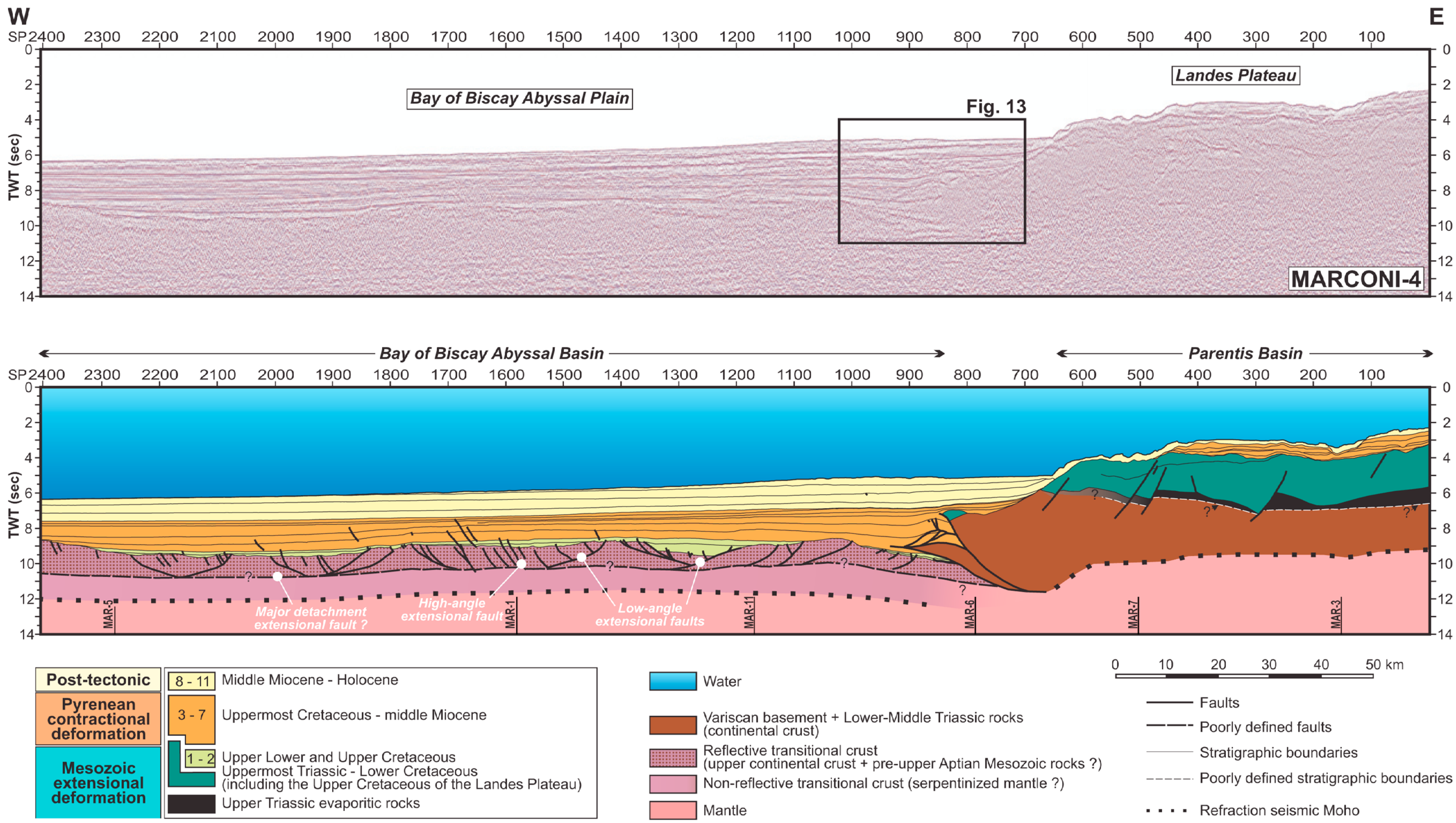

Figure 12. Unmigrated MARCONI-4 deep seismic profile with line drawing interpretation showing the main geological features of the boundary between the transitional crust of the Bay of Biscay abyssal basin (Cantabrian Domain) and the thinned continental crust of the Parentis Basin (Basque-Parentis Domain). Refraction seismic Moho boundary and nature of the crust are derived from *Ruiz* [2007] data. See Figure 2 for location.

basement and are erosionally truncated by the overlying near-horizontal reflectors of the Santonian or Campanian unit 2 (Figure 11). Such a geometric relationship indicates that, at least here, the South Gascogne Ridge belongs to a basement uplift formed between the Cenomanian and the Santonian. This ridge disappears progressively east of the MARCONI-1 profile although it is still identifiable in the MARCONI-11 profile. On the contrary, west of the MARCONI-1 profile, published seismic profiles [*Derégnaucourt*, 1981; *Alvarez-Marrón et al.*, 1996; *Ayarza et al.*, 2004] depict that it becomes more prominent until connect with the Bay of Biscay oceanic ridge. This lateral continuity, together with the chaotic and poor reflectivity of the basement and the geometry of the overlying sediments allow us to interpret the South Gascogne Ridge as an embryonic oceanic crust that formed and warped up during the Cenomanian-Santonian times. The nature of this embryonic oceanic crust is difficult to define due to the lack of wells drilling the South Gascogne Ridge. However, the similarity between the seismic features of this ridge and those observed in the North Atlantic ridges drilled by the 1070 and 1277 ODP sites [*Jagoutz et al.*, 2007; *Tucholke et al.*, 2007] suggest that it was probably also integrated by exhumed continental mantle with variable volumes of E- to N-MORB- type to alkaline melts.

[46] South of the South Gascogne Ridge, the transitional crust of the North Pyrenean foreland also appears tilted toward the south near the North Pyrenean thrust front. The portion of the flexed crust is narrower (<50 km) than in the Basque-Parentis Domain (68 km) and shows a steeper dip (8–12, 5°) in comparison with the 4° of this last domain (Figures 4, 7, 8, and 9). This denotes that the rigidity of the flexed lithosphere is different in the two domains with a Te (elastic thickness) clearly lower in the Cantabrian Domain. The lower crustal rigidity of this last domain could be related to the different nature of the flexed lithosphere but also to its lithospheric temperature that has been much hotter than in the Basque-Parentis Domain from the onset of the Pyrenean deformation (see above discussion comparing the thickness of the sedimentary posttectonic package basin infill in both domains).

[47] In this flexed crust as well as north of the South Gascogne Ridge, the transitional crust is integrated by an upper slightly reflective level with low seismic velocities (5.4–5.7 km/s [*Fernández-Viejo et al.*, 1998; *Ruiz*, 2007]) and a lower level of chaotic seismic signature in which the seismic velocity is much higher (~7.2 km/s [*Fernández-Viejo et al.*, 1998; *Ruiz*, 2007]). Comparing these crustal features with those observed for the transitional crusts of the Armorican, West Iberia and Newfoundland passive margins [*Thinon et al.*, 2003; *Péron-Pinvidic et al.*, 2007], it is inferred that: the upper layer is integrated by upper continental crustal rocks and pre-Albian prerift and synrift sediments; and the lower layer by serpentinized continental mantle rocks. The reflectivity of the upper layer is defined by undulated and very discontinuous reflectors that mostly dip to the north forming large wavelength braided bodies. The geometry of these reflectors and bodies as well as the displacements observed at top of the basement denote that the transitional crust underlying the Bay of Biscay abyssal plain is cut by two sets of south dipping extensional faults (Figures 7, 11, and 12). One not so well defined, integrated by low-angle extensional faults that affect the lower part of the Albian to Cenomanian unit 1 and other formed by younger high-angle extensional faults.

[48] Above this last set of extensional faults, reflectors of the overlying sedimentary units are folded forming monoclines with a south dipping limb or, in some cases, tight anticlines with a steeper and longer southern limb (Figure 11). The geometric relationships between these folds and the underlying extensional faults allow interpretation of them as extensional fault-propagation folds [*Jin and Groshong*, 2006; *Withjack and Schlische*, 2006]. In the fault located at the SP 860, this propagation would begin during the Late Cretaceous as attested by the progressive decrease of the dip of the unit 2 reflectors in the anticline limbs (Figure 11). In the other identified fault-propagation folds, the dip variations and crosscutting relationships of the reflectors in their limbs denote that they developed later during the sedimentation of the upper Eocene (?) to middle Miocene units 5 to 9 with an age that is progressively younger toward the north (Figures 7 and 11). In particular, these growth strata geometries show that these folds and related extensional faults formed coevally to the sedimentation of units 5 to 7 at the South Gascogne Ridge and coevally to the sedimentation of the "posttectonic" units 8 and lower of the 9, northward, at the ones located at the SP 610, 660 and 290 (Figures 7 and 11). Consequently, the development of the North Pyrenean thrust front appears synchronous to the extensional motion of high-angle faults at the northern boundary of the flexed North Pyrenean foreland (bulge) which apparently migrated toward the north during the sedimentation of the middle Miocene units 8 and 9. Besides these plate flexure related extensional faults, there is another high-angle fault that bounds transitional crust of the Bay of Biscay abyssal plain to the north. This fault has a large (up to 2 TWT sec) dip separation and affects the 8 to 10 units but not the overlying unit 11 (Figure 7).

3.3. The Boundary Between the Basque-Parentis and Cantabrian Domains

[49] Interpretation of the MARCONI seismic profiles, therefore, shows that the Bay of Biscay consists of two distinct domains as far as the Mesozoic extensional grain of the deformed crust, its lithospheric nature, and the relationship of the North Pyrenean front with respect to the extensional system are concerned: the Basque-Parentis Domain and the Cantabrian Domain. In the Basque-Parentis Domain, the North Pyrenean front is located close to the Spanish coast and the foreland consists of a thinned continental crust. In the Cantabrian Domain the North Pyrenean front is shifted to the north and the foreland is characterized by mantle exhumation.

[50] In the linkage between the Basque-Parentis and the Cantabrian North Pyrenean thrust fronts, the boundary between both domains coincides with a bathymetric low located between the Landes Plateau and the Asturian Platform in which there are two deep north trending canyons (Santander and Torrelavega canyons) separated by a rela-

tively deep ridge (Santander spur, Figures 2 and 3). The structure of this 60 km wide and north trending bathymetric low is difficult to characterize due to the scarcity and quality of the seismic profiles acquired in this boundary zone. Nevertheless, the available profiles (including the MARCONI-4 and 8 profiles) show that the continental crust appears deformed by northeast to east trending thrusts limited by N to NNE trending faults [*Derégnaucourt*, 1981; *Sánchez*, 1991]. The northeast to east trending thrusts and related folds verge northward and mainly developed along strike of both North Pyrenean fronts. They affect the basement and the overlying pre-middle Miocene sedimentary sequences [*Cámara*, 1989; *Sánchez*, 1991]. The greater thickness of the Mesozoic successions in their hanging wall suggests they resulted from the inversion of Early Cretaceous extensional faults during the Pyrenean deformation [*Sánchez*, 1991]. The N to NNE trending faults, on the other hand, are present along the whole linking zone and do not show any clear vergence [*Derégnaucourt*, 1981]. They have been interpreted as Jurassic-Paleogene extensional faults which would have controlled the location of the Santander and Torrelavega canyons [*Derégnaucourt*, 1981]. However, onshore, at the southern continuation of this linking zone, N to NNE trending faults connecting east-west to northwest-southeast extensional faults (i.e., Bustriguado, Saltacaballos and Ramales faults; see location in Figure 3) have been identified as Early Cretaceous extensional transfer zones that have been subsequently inverted obliquely during the Pyrenean shortening [*Quintana et al.*, 2006, 2009; *López-Mir and Roca*, 2008; *García-Senz and Robador*, 2009].

[51] Then, the linkage between the Basque-Parentis and Cantabrian North Pyrenean fronts appears as a synthetic partially overlapped transfer zone with predominating N to NNE trending faults. These faults would have been active during the Paleogene, synchronously with north directed thrusts and would have inverted previously developed transfer faults of the Early Cretaceous extensional fault system. The lack of well developed strike-slip faults connecting both fronts as well as the diffuse distribution of the N to NNE trending faults in a 60 km wide zone indicates that this transfer zone belongs to a soft fault linkage.

[52] North of this soft transfer zone, at the NW corner of the Landes Plateau (Figure 3), the MARCONI-4 profile shows that the boundary between the Basque-Parentis and the Cantabrian domains corresponds to a northeast trending thrust system that places the continental basement of the Landes Plateau over the transitional crust of the abyssal plain (Figure 12). This thrust system includes a complex triangle zone with a folded intracutaneous thrust wedge truncated backward by a steeper WNW-directed thrust (Figure 13). Over the passive roof thrust of this thrust wedge, the upper part of unit 3 and the bottom of unit 4 are involved in a duplex of back thrusts which, together with the overlying units 4, 5 and 6, are folded forming an open anticline.

[53] Thickness and reflector dip variations of the growth package related with this thrust system demonstrates that it developed during the sedimentation of the upper Eocene-lower Miocene units 5 and 6 (Figure 13), synchronously with the frontal structures of the abyssal thrust system (Figure 9). This age also coincides with the incision and development of the erosive Cap Breton and Cap Ferret canyons on top of the Landes Plateau [*Ferrer et al.*, 2008] which could be explained by the emplacement of the described thrust system which sank the easternmost parts of the abyssal plain in relation to the overthrusting Landes Plateau.

4. Kinematics of the Pyrenean Deformation in the Eastern Part of the Bay of Biscay

[54] The lack of wells cutting the lower and middle part of the sedimentary infill of most of the Bay of Biscay's deeper parts makes it difficult to date the evolution of the Pyrenean contractional deformation. However, from the description and geometric analysis of the MARCONI profiles as well as from the information provided by the samples collected by submersibles [*Capdevila et al.*, 1980; *Malod et al.*, 1980, 1982; *Boillot et al.*, 1987] and wells drilled in the neighboring areas (i.e., wells in the western Bay of Biscay, Parentis Basin and Armorican and North Iberian margins [*Laughton et al.*, 1972; *Montadert et al.*, 1979b; *Derégnaucourt*, 1981; *Lanaja*, 1987; *Sánchez*, 1991; *Bois et al.*, 1997; *Gutiérrez Claverol and Gallastegui*, 2002; *Biteau et al.*, 2006; *McDougall et al.*, 2009; *Pérez-García et al.*, 2009]), we can infer the presence of three deformational stages in the evolution of the Bay of Biscay linked to the development of the Pyrenean orogen (Figure 14).

[55] 1. An initial stage, Late Cretaceous (Senonian) in age, when the eastern part of the Bay of Biscay was submitted to a compressive tectonic regime that produced the development of a widespread erosive unconformity mainly visible in the Basque-Parentis Domain (Figures 4–6) and at the margins fringing the Bay of Biscay abyssal plain.

[56] 2. An intermediate stage, latest Cretaceous to early Eocene in age, characterized by the formation of the North Pyrenean foreland basin accompanied, in the Cantabrian Domain, by the activation of south dipping extensional faults close to its peripheral bulge (Figures 7 and 11). Such fault locations as well as the synchrony between foreland basin development and extensional faulting denote layer-parallel stretching in the outer arc of the flexed lithosphere hinge zone. The flexure of the North Pyrenean foreland basin during this stage demonstrates growing of the Pyrenean orogen. However, no significant contractional deformation has been recorded in the study areas except some minor thrusts and folds that, described in the Le Danois Basin [*Pérez-García et al.*, 2009], are insufficient to explain the foreland flexure. As a consequence active contractional structure should be expected to occur at the present continental slope. These inner and older Pyrenean structures would continue to the east into the northern part of the Basque-Cantabrian Basin and onshore into the Western Pyrenees. There, north directed thrusts at the thrust front controlled a significant flexure of the north foreland at Late Cretaceous-Eocene times that resembles the geometry observed in the Bay of Biscay [*Debroas*, 1990, *Déramond et al.*, 1993].

[57] 3. A final stage, middle Eocene to middle Miocene in age, characterized by the development of the major con-

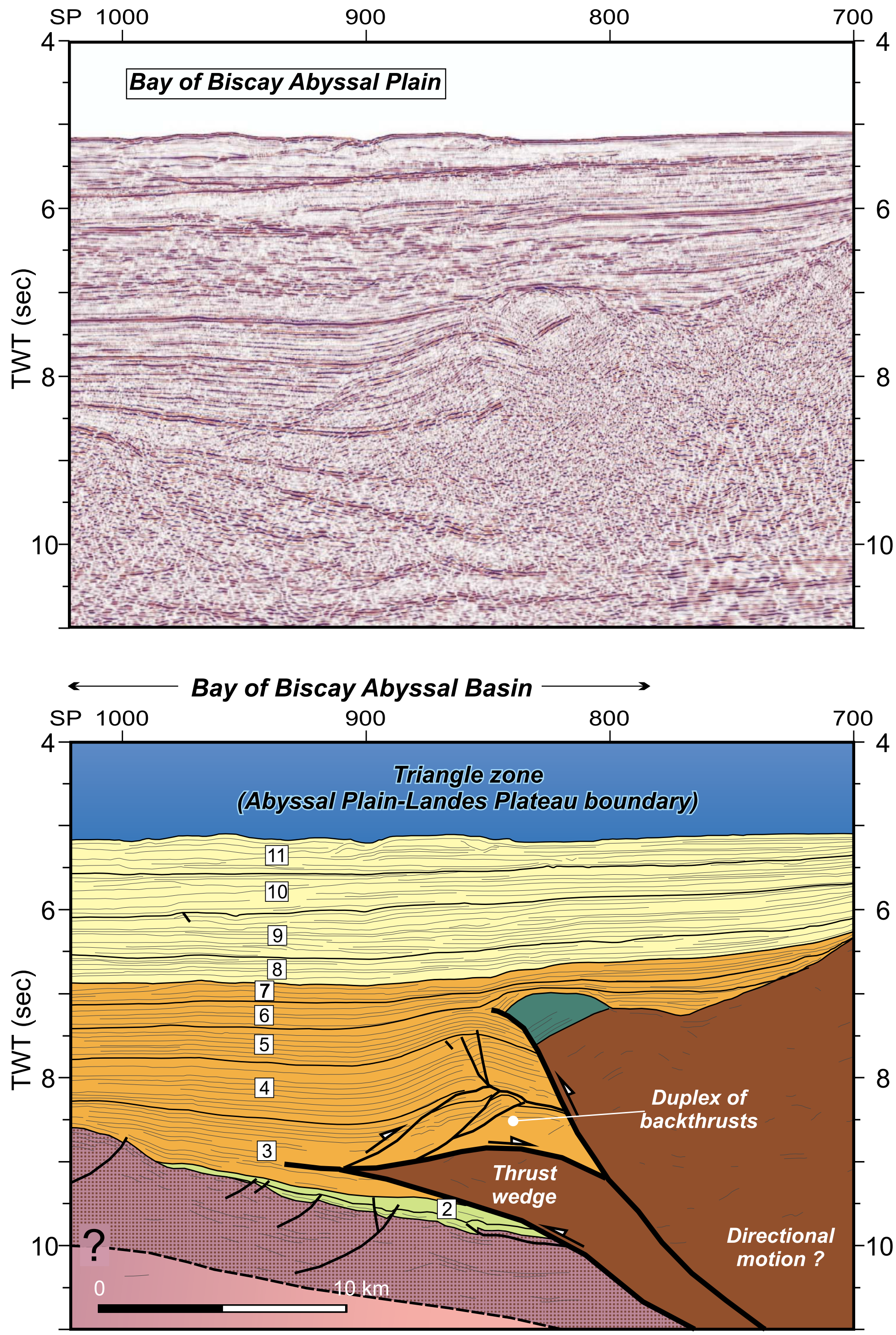

Figure 13. Detail of the thrust structure in the boundary between the transitional crust of the Bay of Biscay abyssal basin (Cantabrian Domain) and the thinned continental crust of the Parentis Basin (Basque-Parentis Domain) along the unmigrated MARCONI-4 profile. Labels 1–11 are Bay of Biscay abyssal basin fill differentiated units. See Figure 12 for location and legend.

tractional structures observed along the North Pyrenean thrust fronts at the eastern part of the Bay of Biscay. These structures consist of imbricate thrust systems and triangle zones following a piggyback thrusting sequence up to the late Oligocene. Afterward, deformation diminished and migrated toward the hinterland following a break-back thrusting sequence. North of North Pyrenean front, the flexure of the North Pyrenean foreland initiated during the

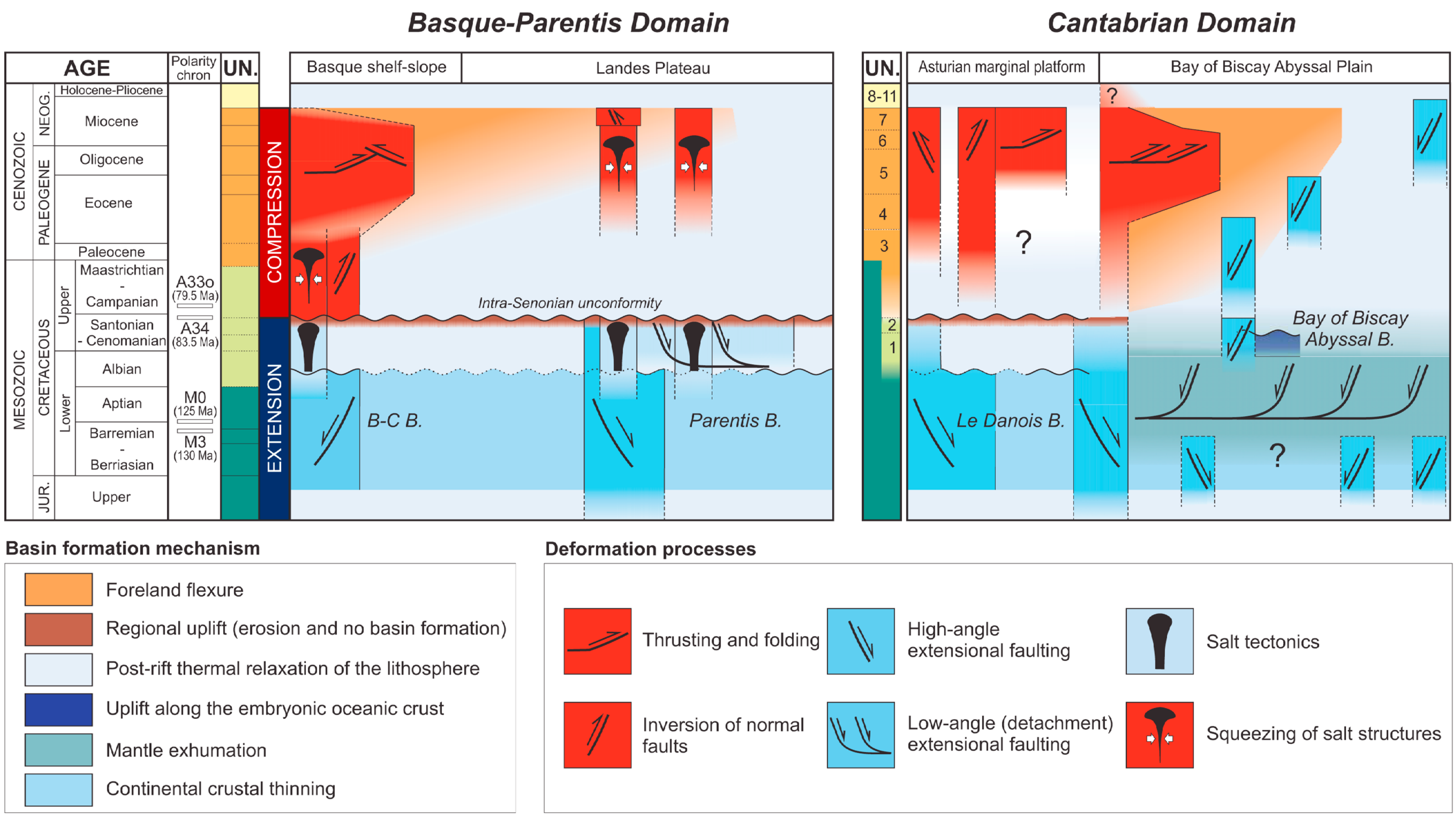

Figure 14. Timing and main features of the Alpine deformations occurred in the Basque-Parentis and Cantabrian domains of the eastern part of the Bay of Biscay. Timing and main features of the deformation along the North Iberian shelf and platform are based on *Boillot et al.* [1987], *Sánchez* [1991], *García-Mondéjar* [1996], *Gallastegui* [2000] *Gutiérrez Claverol and Gallastegui* [2002], *McDougall et al.* [2009] and *Pérez-García et al.* [2009].

latest Cretaceous continued during this stage. In the Cantabrian Domain, the foreland subsidence decreased as the width of the flexed foreland lithosphere does. As a result, during the middle Eocene to middle Miocene, the peripheral foreland bulge and the associated extensional faults migrated northward. In contrast, in the Basque-Parentis Domain, flexure subsidence and peripheral bulge location remained rather constant. Furthermore in this domain, the foreland was not affected by flexure related extensional faults but by contractional structures formed from the squeezing of Albian to Upper Cretaceous salt diapirs and, locally in the latter stages of deformation (middle Miocene), from the inversion of some extensional faults.

[58] This evolution evidences that, in the eastern part of the Bay of Biscay, (1) the onset of the Pyrenean contractional deformation corresponds to a regional event which affected the whole area and resulted into the development of the Senonian unconformity; (2) during the latest Cretaceous to early Eocene, Pyrenean contractional deformation was restricted to the inner parts of the belt, at present located along the North Iberian slope and shelves; and (3) from middle Eocene to middle Miocene, the contractional deformation progressed northward affecting the Bay of Biscay abyssal plain and the Landes Plateau.

[59] The MARCONI profiles also clearly illustrate that Pyrenean deformation ended at the middle part of the Miocene. Even though, they also show that, locally (i.e., 3270 Seamount Block crest, see profile MARCONI-1 in Figure 7), some minor contractional deformation has persisted up to nowadays.

5. The Bay of Biscay-Pyrenean Late Jurassic-Early Cretaceous Extensional Rift System

[60] Restoration of the geological cross sections, mostly taken advantage of the new MARCONI profiles, and palinspastic reconstruction of the Early Cretaceous basins give us an idea of the main structural grain related with the development of the Bay of Biscay-Pyrenean rift system. Such reconstruction can reasonably be done because of the moderate amount of convergence related with the Pyrenean orogenic event. Moreover, equivalent structures and basins are preserved in the adjacent Armorican and West Iberian margins as well as in their equivalent margins at the other side of the Atlantic. In these margins a significant amount of geophysical and well data have integrated into well constrained crustal-scale cross sections which have recently resulted into new models for the continental extension and breakup of magma-poor margins [*Manatschal*, 2004; *Lau et al.*, 2006; *Hopper et al.*, 2007; *Huismans and Beaumont*, 2007; *Péron-Pinvidic et al.*, 2007; *Tucholke et al.*, 2007; *Péron-Pinvidic and Manatschal*, 2009]. These models rely on the reconstruction of the conjugate margins at both sides of the Atlantic and the deduced evolution. However, we can take advantage of the along strike decrease in the amount of extension and plate separation in the Bay of Biscay and Pyrenean domain to decipher the evolution of lithospheric extension (from rifting to mantle exhumation and oceanization) taken as a reference the end-members observed in different transects. In other words, in the Bay of Biscay-Pyrenees we have a unique opportunity to investigate at surface an Atlantic margin with all the main structural elements well preserved, from the deepest to the upper parts, such as the exhumed mantle and basins above, the synrift basins or the postrift collapse features.

[61] The extensional rift system between Iberia and Eurasia consisted of two parallel overlapping and segmented arms oriented east-west. They overlap at the eastern Bay of Biscay where they are separated by a high that includes the Landes and Le Danois highs (Figure 15). The northern arm (the Bay of Biscay arm) includes the Bay of Biscay abyssal basin and the Parentis Basin, and the southern one (the Pyrenean arm), the Le Danois, Basque-Cantabrian and the Lacq-Mauleon basins.

[62] The architecture/structure of these basins appears strongly dependent on the crustal thickness and, in particular, on the presence or absence of ductile middle to lower continental crust beneath the basins. Thus, in the areas where crustal stretching has been minor and there is a ductile middle to lower continental crust (i.e., Le Danois Basin), the basin structure consists of a broad half-graben system of conjugate high-angle extensional faults affecting the upper crust (see Figure 7 and Figure 15, top). On the contrary, in the cases in which the lithospheric stretching is larger and the basins lie over a much thinner crust without a ductile middle to lower continental crust (i.e., Parentis, Basque-Cantabrian and the Lacq-Mauleon basins), the basin structure is more asymmetric and is controlled by the development of detachment faults that cut the entire crust. These detachments produce the exhumation of serpentinized subcontinental mantle and can carry extensional riders of upper continental crust directly above the mantle. Such structural configuration has been observed in the Newfoundland and West Iberia margins [*Péron-Pinvidic and Manatschal*, 2009] and has been interpreted to occur in the Lacq-Mauleon Basin [*Jammes et al.*, 2009]. Exhumation of the mantle during lithospheric stretching largely depends on the rheology of the crust in order to produce crustal embrittlement during extensional deformation. Such embrittlement facilitates mantle exhumation as the extensional faults cut the entire thinned continental crust and detach into the mantle. Extensional deformation and fluids along the faults produce the serpentinization of the mantle which in its turn localizes strain and detachment, processes that end up with the development of shallow detachments affecting the whole thinned lithosphere [*Pérez-Gussinyé and Reston*, 2001; *Pérez-Gussinyé et al.*, 2001; *Reston*, 2007]. This deformation mode has been documented to occur in the West Iberia Margin before the continental breakup and the accretion of normal oceanic crust. Thus, it is reasonable that it also occurred in the Bay of Biscay given the same crustal configuration of northern and western Iberia before the onset of stretching. Moreover, the wide occurrence of peridotites along major faults in the lower Cretaceous basins involved in the Pyrenean thrusts sheets [*Vielzeuf and Kornprobst*, 1984; *Mendia and Gil-Ibarguchi*, 1991; *Fabriès et al.*, 1998] and the field evidences that they were brought to the bottom of the Albian basins [*Choukroune*, 1980; *Lagabrielle and Bodinier*, 2008; *Jammes et al.*, 2009] give

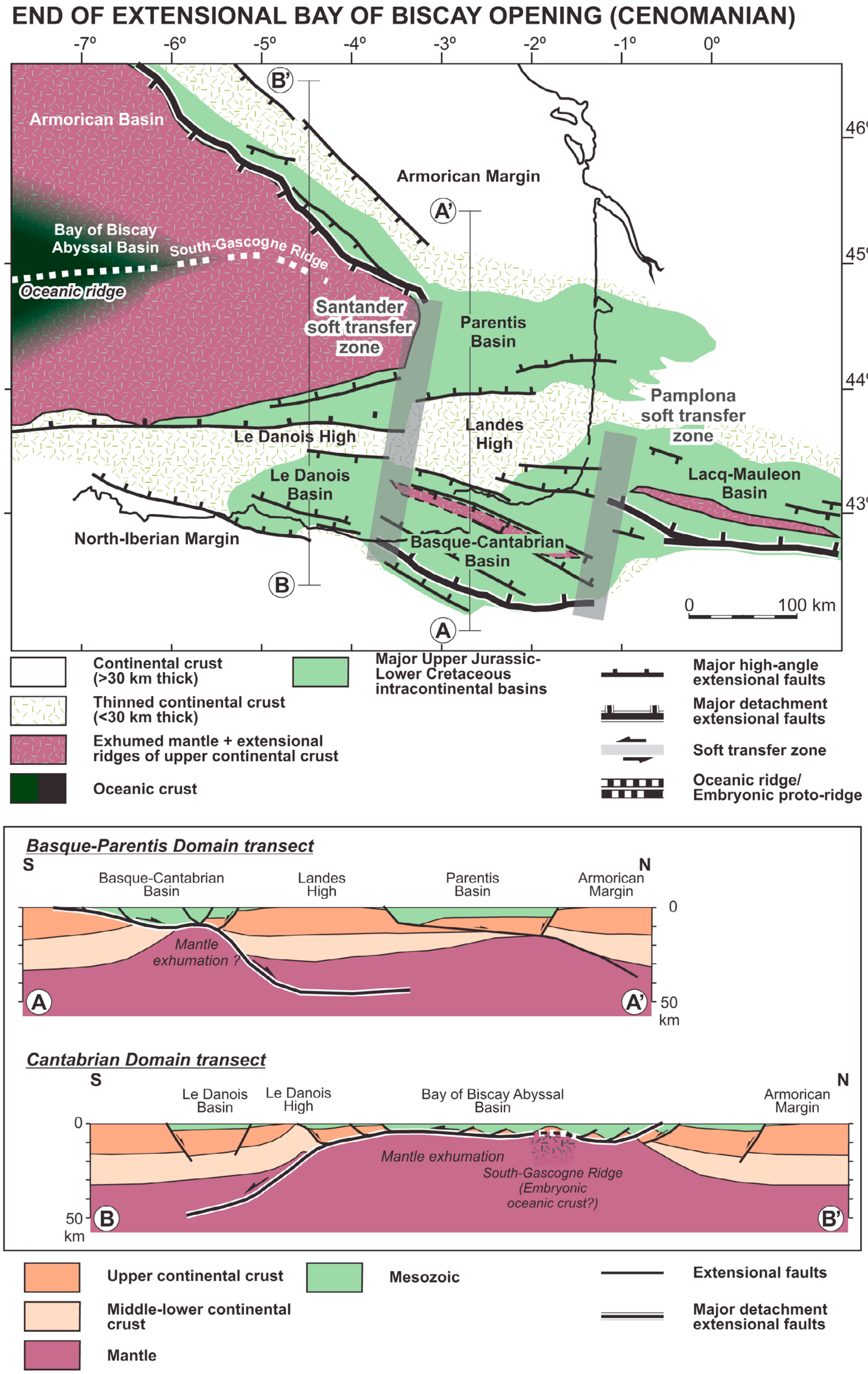
END OF EXTENSIONAL BAY OF BISCAY OPENING (CENOMANIAN)
-7º -6º -5º -4º -3º -2º -1º 0º
46º 45º 44º 43º
Armorican Basin
Armorican Margin
Bay of Biscay Abyssal Basin
South-Gascogne Ridge
Oceanic ridge
Santander soft transfer zone
Parentis Basin
Pamplona soft transfer zone
Landes High
Le Danois High
Le Danois Basin
Lacq-Mauleon Basin
North-Iberian Margin
Basque-Cantabrian Basin
B' A' B A
0 100 km
Continental crust (>30 km thick)
Thinned continental crust (<30 km thick)
Exhumed mantle + extensional ridges of upper continental crust
Oceanic crust
Major Upper Jurassic-Lower Cretaceous intracontinental basins
Major high-angle extensional faults
Major detachment extensional faults
Soft transfer zone
Oceanic ridge/ Embryonic proto-ridge
Basque-Parentis Domain transect
S N
Basque-Cantabrian Basin
Landes High
Parentis Basin
Armorican Margin
Mantle exhumation ?
0
50 km
Cantabrian Domain transect
Le Danois Basin
Le Danois High
Bay of Biscay Abyssal Basin
Armorican Margin
Mantle exhumation
South-Gascogne Ridge (Embryonic oceanic crust?)
Upper continental crust
Middle-lower continental crust
Mantle
Mesozoic
Extensional faults
Major detachment extensional faults

Figure 15

further support to the idea that mantle exhumation occurred in the Bay of Biscay-Pyrenean rift system before and during the normal oceanic accretion in the western Bay of Biscay (Aptian-Albian to Santonian in age [*Sibuet et al.*, 2004b]).

[63] In relation to this oceanic accretion, the geometrical arrangement of the mapped magnetic lineations in the Bay of Biscay denote a progressive decrease of the amount of accreted oceanic crust toward the east up to its disappearance close to the 6°W meridian [*Sibuet et al.*, 2004a, 2004b] (Figure 17). Eastward, the Bay of Biscay abyssal plain appears only floored by transitional crust with a major ridge striking along its axis (South Gascogne Ridge; Figure 7) that vanishes toward the east. This ridge, formed after the low-angle extensional faulting that triggered the exhumation of the continental mantle, is interpreted (see section 3.2.2.3) as being formed by an embryonic oceanic crust that warped up during the last stages of Bay of Biscay oceanic accretion (Cenomanian to Santonian times). The presence of this embryonic oceanic crust in the eastern Bay of Biscay abyssal plain would imply that the oceanic accretion in the Bay of Biscay propagated eastward at the same time that its extension increased along the Bay of Biscay-Pyrenean rift system.

[64] From a more regional point of view, the two arms of the Bay of Biscay-Pyrenean rift system appear formed by segmented extensional fault systems that are linked by north to NNE trending accommodation zones. Among these accommodation zones are outlined (Figure 15) (1) the Pamplona transfer zone linking the Basque-Cantabrian and Lacq-Mauleon basins (Pamplona Fault [*Schoeffler*, 1982; *Turner*, 1996; *Larrasoaña et al.*, 2003]) and (2) the Santander soft transfer zone that, separating the Basque-Cantabrian from Le Danois basins, extends northward up to mark the boundary between the Bay of Biscay abyssal and Parentis basins.

[65] This last transfer zone is especially significant since it bounds the transitional crust of the Bay of Biscay abyssal basin toward the east and displaces the locus of the major lithospheric thinning from the Bay of Biscay arm to the Pyrenean arm where continental mantle was exhumed along the Basque-Cantabrian and Lacq-Mauleon basins (Figure 15). This exhumed continental mantle is not present in the Parentis Basin (western prolongation of the Bay of Biscay abyssal basin) where the Upper Jurassic-Cretaceous basin fill overlies older Mesozoic or crustal rocks. The greater extension along the Pyrenean arm eastward of the Santander transfer zone is also evidenced by the much thicker thickness of Cretaceous successions [*Brinkmann and Lögters*, 1968; *García-Mondéjar et al.*, 2004; *García-Senz*, 2004] as well as the widespread Cretaceous alkaline magmatism [*Azambre and Rossy*, 1976; *Castañares et al.*, 2001; *Castañares and Robles*, 2004] and coeval thermal metamorphism [*Albarède and Michard-Vitrac*, 1978; *Montigny et al.*, 1986; *Golberg and Leyreloup*, 1990; *Cuevas and Tubía*, 1999] which have not been described in the Parentis Basin.

[66] Also, the polarity of the extensional crustal-scale detachments is as well opposite on both sides of this soft transfer zone. Thus, westward, in the Bay of Biscay abyssal basin, the predominance of extensional riders of "continental crust" bounded by low-angle south dipping faults over the exhumed mantle (Figures 7 and 15) enable us to decipher that the main detachment dips southward. On the contrary, east of the transfer zone, the geometry of the Parentis, Basque-Cantabrian and Lacq-Mauleon Cretaceous basin infills, as well as the cross-fault relationships with the underlying rocks, denote that the main detachments dip northward.

6. The Significance of the Late Aptian to Middle Albian "Breakup" Unconformity

[67] Since the arrival of plate tectonics, the end of extensional processes on passive margins and the onset of oceanic accretion have been correlated with the development of a broad well-developed unconformity (breakup unconformity) on the continental margins separating the synrift to the postrift sediments [*Falvey*, 1974]. However, in the last years, such correlation has been questioned in different rift systems as in the Newfoundland-Iberian Rift where the "breakup" unconformity appears coeval to mantle exhumation and older to the onset of oceanic crust accretion [*Péron-Pinvidic et al.*, 2007; *Sibuet et al.*, 2007b; *Tucholke et al.*, 2007].

[68] In the Bay of Biscay-Pyrenean rift system, the age of the "breakup" unconformity that separates the synrift and postrift sediments on the Bay of Biscay continental margins also does not coincide with the onset of mantle exhumation or oceanic accretion defined with the magnetic anomalies. The "breakup" unconformity is middle to late Aptian in age in the continental margins fringing the Bay of Biscay oceanic crust (Ortegal Spur and Armorican Margin [*Montadert et al.*, 1979a; *Derégnaucourt and Boillot*, 1982; *Boillot et al.*, 1987; *Thinon et al.*, 2002; *Wallrabe-Adams et al.*, 2005]) and younger eastward (Parentis, Basque-Cantabrian, Mauleon-Arzacq and Organyà-Turbón basins) where it has been dated as middle Albian [*Le Vot et al.*, 1996; *Bois et al.*, 1997; *Gräfe*, 1999; *Vergés and García-Senz*, 2001; *García-Mondéjar et al.*, 2004; *García-Senz*, 2004; *Biteau et al.*, 2006; *Jammes et al.*, 2009]. Instead, the onset of the oceanic accretion has been traditionally postulated to be Barremian-early Aptian in age based on the presence of the M3 and M0 magnetic lineations along the northern margin of Bay of Biscay abyssal plain [*Olivet*, 1996; *Sibuet et al.*, 2004a] as well as the seismic signature of the uppermost crust [*Thinon et al.*, 2003]. This mismatch between the "breakup" unconformity and the onset of the oceanic accretion ages could be solved in the West Bay of Biscay con-

Figure 15. The structure of the eastern Bay of Biscay and surrounding areas at the end of the Bay of Biscay opening (Cenomanian). (top) Palinspastic reconstruction; (bottom) schematic crustal-scale transects. Plate reconstruction based on *Sibuet et al.* [2004b].

sidering that the M0 and M3 anomalies were not caused by the magnetization of the volcanic rocks formed by seafloor spreading but primarily by the serpentinization of the exhuming mantle rocks as described by *Sibuet et al.* [2007b] in the Newfoundland-Iberia Rift. In agreement with this interpretation, refraction and multichannel seismic studies carried out across the M0-M3 magnetic sequence in the northern Bay of Biscay abyssal plain [*Thinon et al.*, 2003] document a smoothed reflection Moho and the presence of 5 km thick crust (velocity 4.8–7.0 km/s) overlying a 3–4 km thick "abnormal" mantle with velocities of 7.4–7.5 km/s. These basement features are similar to those described in the transitional crust of the Newfoundland-Iberia Rift where there is a general absence of a Moho reflection and the basement is integrated by a thin low-velocity (4.5–7 km/s) basement overlying a approximately 5 km thick serpentinized continental mantle with <7.2–7.9 km/s velocities [*Chian et al.*, 1999; *Lau et al.*, 2006; *Tucholke et al.*, 2007].

[69] Although a correlation between the breakup unconformity and the onset of the oceanic crust could be therefore postulated in the West Bay of Biscay, in the East Bay of Biscay and Pyrenees, where oceanic crust is not present, there is no doubt that the "breakup" unconformity did not form at the end of the extensional processes on the continental lithosphere. Here, the "breakup" unconformity, middle Albian in age, developed before the development of the embryonic oceanic crust in the South Gascogne Ridge (Cenomanian to Santonian in age) and during the mantle exhumation process which was active at least up to the end of Albian times [*Henry et al.*, 1998; *Jammes et al.*, 2009]. Consequently, in these areas, this is not a true breakup unconformity but an unconformity associated to a change in the mode of the lithospheric extensional deformation or in the rate of extension. The origin of this "breakup" unconformity cannot be associated with the final extension of continental crust and initial exhumation of lithospheric mantle since this process began before: during the Berriasian (West Bay of Biscay [*Fügenschuh et al.*, 2003]) to latest Aptian (Pyrenees [*Henry et al.*, 1998]) times. Therefore, it can be neither correlated with the change of crustal thinning mechanism that occurred before when the development of the extensional detachment faults began. Consequently, the "breakup" unconformity seems more related to a change in the extension rate. In this way, the extension rates calculated in the West Bay of Biscay depict a significant drop of the extension rates after the formation of the middle Aptian to middle Albian "breakup" unconformity. Concretely, using the *Gradstein et al.* [2004] time scale, the extension rate calculated between the nearly parallel Barremian M0 and M3 magnetic lineations (10–12, 3 mm/yr) appears higher than the one calculated for the Santonian-Campanian interval comprised between the A34 magnetic lineation and the Bay of Biscay ocean spreading axis (greater than 6.7–7.1 mm/yr). This inferred drop of the extensional rates occurred during or after the counter-clockwise rotation of Iberia [*Gong et al.*, 2008] between the formation in the West Bay of Biscay of the V shaped M0 and the A33 and A34 east trending magnetic lineations (Figure 17).

[70] Thus, we suggest that the widespread late Aptian to middle Albian unconformity present in the entire Bay of Biscay-Pyrenean rift system does not record a continental breakup but is related to a dramatic decrease of the extension rate which occurred after the Iberian clockwise rotation. It does not record the end of the extension since after the Albian and up to Santonian, there is some lithospheric mantle exhumation in the Pyrenees, the development of the South Gascogne embryonic oceanic crust in the Cantabrian Domain and the accretion of oceanic crust in the West Bay of Biscay.

7. Role of the Preexisting Bay of Biscay Mesozoic Structure in the Development of the Pyrenean Deformation

[71] The interpretation of the MARCONI profiles clearly shows that the contractional structure of the eastern part of the Bay of Biscay is strongly controlled by the previous Mesozoic extensional structural grain. Both in the Pyrenees and in its northern foreland, main Pyrenean related contractional deformation is located along the extensionally faulted margins of the major Early Cretaceous basins generated during the Bay of Biscay opening (Figure 16). In the Pyrenees, the features of the contractional structures appear strongly dependent on the geometry of the previous extensional Mesozoic basin bounding faults. Thus, in the basin margins bounded by upper crustal high-angle faults, contractional deformation is distributed along a broad region and includes basement thrusts, cover folds and thrust slices that depict typical geometric features of an inverted fault (i.e., northern margin of the Basque-Cantabrian Basin [*Gómez et al.*, 2002]) or basement thrusts that cut the preexistent faults (i.e, Cantabrian continental slope [*Gallastegui*, 2000]). In contrast, in the basin margins bounded by low-angle detachment faults (i.e., southern margin of the Basque-Cantabrian Basin), contractional deformation is much more localized and mainly consists on the reactivation of these preexistent planes.

[72] These preexistent crustal-scale extensional detachments play a fundamental role in the development of the Basque and Cantabrian Pyrenees not only because most of Alpine shortening is taken up in the contractional reactivation of these structures but also because they control the orogen architecture (Figures 15 and 16). Thus, in the rift segments where they did not develop during the opening of the Bay of Biscay (Eastern and Central Pyrenees), the orogen mainly consist on a wedge made up by an antiformal stack of upper crustal thrust sheets (Figure 1). By contrast, in the rift segments where large crustal-scale extensional detachments were present (Western, Basque and Cantabrian Pyrenees), the orogen architecture is simpler and consists on a slightly deformed wedge of upper crustal rocks displaced above the reactivated detachment (Figures 1 and 16). These architectural differences could be related to the disappearance of the ductile middle to lower crust in the areas affected by the crustal extensional detachments [see *Péron-Pinvidic and Manatschal*, 2009] which would prevent the later

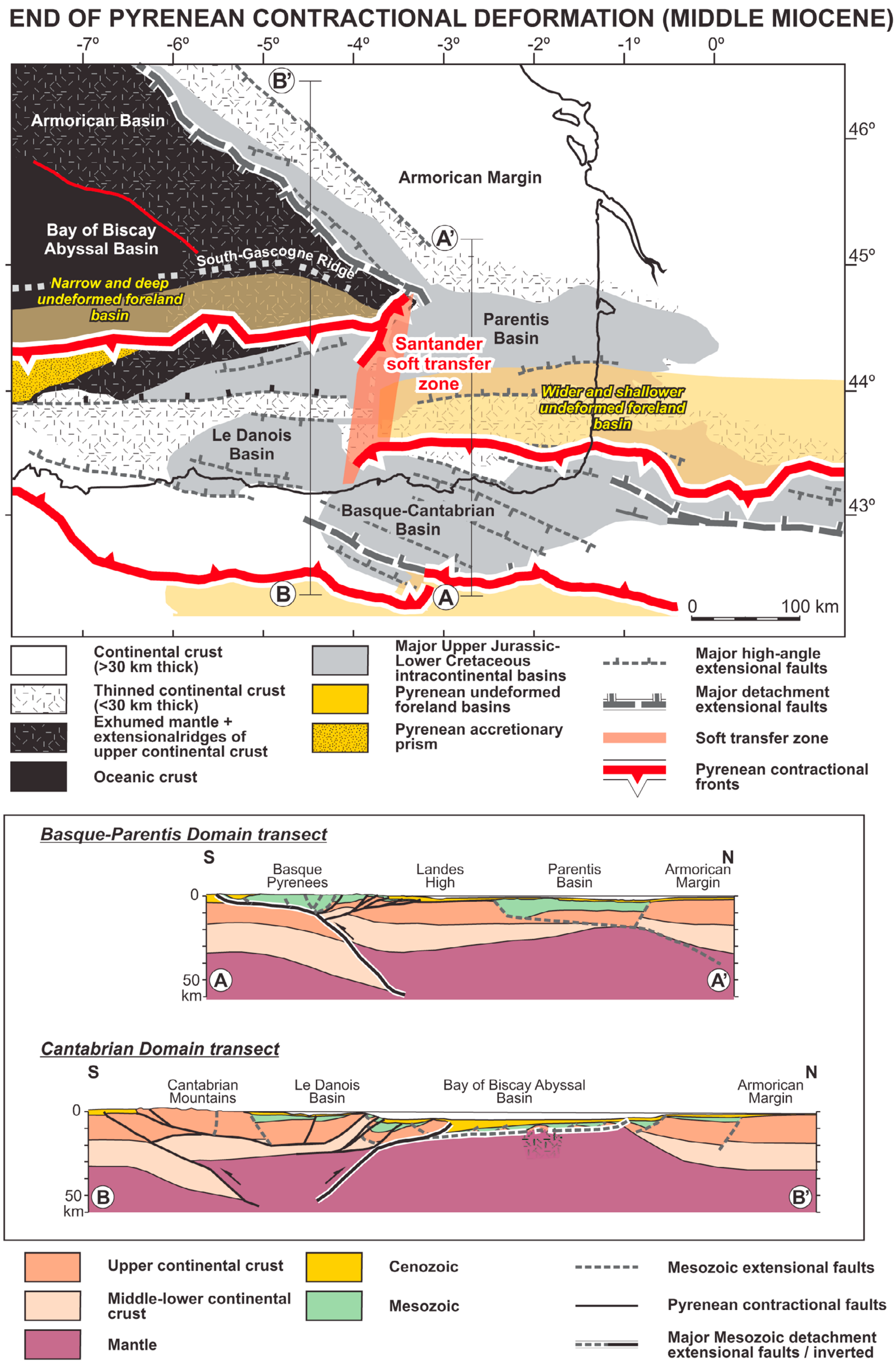
END OF PYRENEAN CONTRACTIONAL DEFORMATION (MIDDLE MIOCENE)
-7º
-6º
-5º
-4º
-3º
-2º
-1º
0º
46º
45º
44º
43º
Armorican Basin
Bay of Biscay
Abyssal Basin
South-Gascogne Ridge
Armorican Margin
Narrow and deep
undeformed foreland
basin
Parentis
Basin
Santander
soft transfer
zone
Wider and shallower
undeformed foreland
basin
Le Danois
Basin
Basque-Cantabrian
Basin
B'
A'
B
A
0
100 km
Continental crust
(>30 km thick)
Thinned continental crust
(<30 km thick)
Exhumed mantle +
extensionalridges of
upper continental crust
Oceanic crust
Major Upper Jurassic-
Lower Cretaceous
intracontinental basins
Pyrenean undeformed
foreland basins
Pyrenean accretionary
prism
Major high-angle
extensional faults
Major detachment
extensional faults
Soft transfer zone
Pyrenean contractional
fronts
Basque-Parentis Domain transect
S
N
Basque
Pyrenees
Landes
High
Parentis
Basin
Armorican
Margin
0
50
km
Cantabrian Domain transect
Cantabrian
Mountains
Le Danois
Basin
Bay of Biscay Abyssal
Basin
Upper continental crust
Middle-lower continental
crust
Mantle
Cenozoic
Mesozoic
Mesozoic extensional faults
Pyrenean contractional faults
Major Mesozoic detachment
extensional faults / inverted

Figure 16

development of midcrustal detachment levels and, therefore, the formation of imbricate and duplex thrust systems at the upper crust during the Pyrenean compression. As a result, in these areas, most or a very significant part of Alpine shortening is taken up in the contractional reactivation of the preexistent crustal low-angle extensional detachments. This is well illustrated in the cross sections of Figure 16 which show that (1) the Basque Pyrenees are an intracontinental orogen developed from the subduction of the Iberian plate beneath the Eurasian plate along the preexistent north dipping extensional detachment that bounded the large Cretaceous Basque-Cantabrian Basin and (2) the Cantabrian Pyrenees is a crustal wedge bounded northward by a south directed subduction developed from the contractional reactivation of the extensional detachment flooring the Cretaceous Bay of Biscay abyssal basin. In relation of this last subduction, the present-day absence of transitional crust south of the Bay of Biscay oceanic crust as well as of the M0 and M3 magnetic lineations (Figure 17) suggest that the amount of subducted crust beneath the Cantabrian continental slope could be much greater than proposed in the previous works [*Fernández-Viejo et al.*, 1998; *Gallastegui*, 2000]. In fact, if we take into account that the width of the magnetic lineations on both sides of the Bay of Biscay oceanic ridge is similar (Figure 17a), and that this proportionality is maintained between the M0 and M3 magnetic lineations, we can infer that about 100 km of the transitional and/or oceanic crust flooring the Cretaceous Bay of Biscay abyssal basin has been consumed beneath such subduction during the Pyrenean orogen development.

[73] North of the North Pyrenean front, in the foreland, the Pyrenean contractional deformation also appears strongly controlled by the preexistent Mesozoic structural configuration. It mainly developed over the foreland areas floored by continental crust (Basque-Parentis Domain) and is represented by gentle folds resulting, first, from the squeezing of the stems of Albian to Upper Cretaceous diapirs and, then, from the limited inversion of the related extensional faults [*Ferrer et al.*, 2008]. Westward, in the *Cantabrian Domain*, there is not significant Pyrenean contractional deformation in the foreland but the spatial coincidence of the North Pyrenean forebulge with the preexistent South Gascogne Ridge (Figures 7 and 11) also denotes a strong interaction between the inherited lithospheric Mesozoic grain and Pyrenean deformation. Specifically, in this case, the location of the Pyrenean forebulge appears determined by the asthenosphere uplift that developed beneath the South Gascogne embryonic oceanic crust during the last stages of the Bay of Biscay opening.

[74] On the other hand, it should be noted that the MARCONI profiles not only show that the location and features of the Pyrenean contractional structures are controlled by the preexistent Mesozoic faults and diapirs, they also reveal that the different characteristics along strike of the Pyrenean orogen are strongly conditioned by the segmentation and geometry of the rift system developed between Iberia and Eurasia from the Late Jurassic up to Late Cretaceous. Indeed, the comparison between the structural maps depicted in Figures 15 and 16 evidences an almost perfect correlation between the extensional Late Cretaceous rift segmentation and the segments of the Pyrenees with a well differentiated contractional structure (Cantabrian Pyrenees, Basque Pyrenees and Pyrenees s.s). Also, these maps show that the Basque Pyrenees and the Pyrenees s.s. did not grow from the tectonic inversion of the Mesozoic basins located in the eastern prolongation of the Bay of Biscay abyssal basin (Parentis Basin), but from the tectonic inversion of a system of Mesozoic basins that, running parallel in the south, included the Basque-Cantabrian, Organyà and Laq-Mauleon basins (Figure 16) [*García-Senz*, 2002]. The origin of this jump could be influenced by the change in the dip sense of the low-angle crustal detachment in the Bay of Biscay rift arm (from south to north) but seems mainly controlled by the major lithospheric and crustal thinning that would present the Pyrenean arm of the Bay of Biscay-Pyrenean rift eastward of the Santander soft transfer zone at the early Late Cretaceous (Figure 15). This major thinning of the Pyrenean arm would result in a weaker lithosphere at the beginning of the Pyrenean compression and, therefore, in a more suitable area to concentrate the Pyrenean contractional deformation.

[75] Therefore, as a conclusion, the MARCONI survey allows to recognize that the rift system developed between Iberia and Eurasia during the Late Jurassic and Cretaceous appears as a major factor controlling not only the location and features of most of the Pyrenean thrust sheets but also the overall structure of the chain.

8. Conclusions

[76] Profile interpretation of MARCONI deep reflection seismic profiles enables recognition of the present upper crustal structure of the eastern part of the Bay of Biscay and the main features of its Alpine geodynamic evolution. Thus, geometric analysis of the seismic signature of the profile, combined with available oil-well and refraction data denote that two domains with a different Pyrenean and north foreland structural evolution exist in the Bay of Biscay: the eastern, Basque-Parentis Domain and the western, Cantabrian Domain.

[77] In the eastern sector (Basque-Parentis Domain), the North Pyrenean front is located close to the coast. It corresponds to a system of east trending basement-involved thrusts and cover thrust sheets controlled by the inversion of the Early Cretaceous crustal low-angle extensional detachment that, dipping to the south, bounded the Early Cretaceous Basque-Cantabrian Basin to the north. North of this

Figure 16. The structure of the eastern Bay of Biscay and surrounding areas at the end of Pyrenean contractional deformation (middle Miocene). (top) Palinspastic reconstruction; (bottom) schematic crustal-scale transects (onshore portions based on *Pulgar et al.* [1999], *Gallastegui* [2000], *Gómez et al.* [2002], *Pedreira et al.* [2003, 2007], and *Ferrer et al.* [2008]). Note the role played by the major Cretaceous extensional detachments.

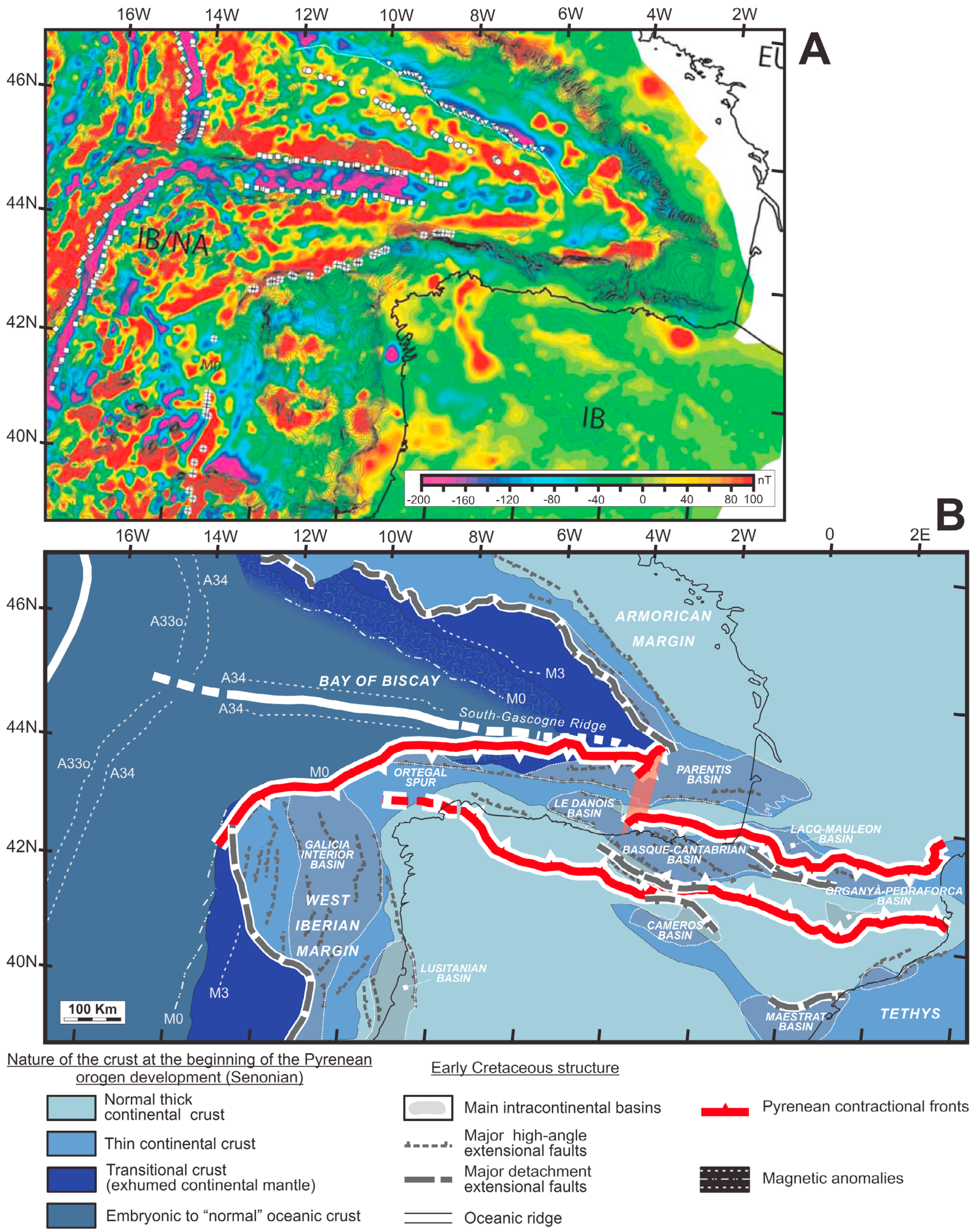
A
B
16W
14W
12W
10W
8W
6W
4W
2W
0
2E
46N
44N
42N
40N
EU
IB/NA
IB
M0
nT
-200
-160
-120
-80
-40
0
40
80
100
A34
A33o
BAY OF BISCAY
South-Gascogne Ridge
M3
M0
ARMORICAN
MARGIN
PARENTIS
BASIN
ORTEGAL
SPUR
LE DANOIS
BASIN
LACQ-MAULEON
BASIN
BASQUE-CANTABRIAN
BASIN
ORGANYÀ-PEDRAFORCA
BASIN
CAMEROS
BASIN
GALICIA
INTERIOR
BASIN
WEST
IBERIAN
MARGIN
LUSITANIAN
BASIN
MAESTRAT
BASIN
TETHYS
100 Km
Nature of the crust at the beginning of the Pyrenean orogen development (Senonian)
Normal thick continental crust
Thin continental crust
Transitional crust (exhumed continental mantle)
Embryonic to "normal" oceanic crust
Early Cretaceous structure
Main intracontinental basins
Major high-angle extensional faults
Major detachment extensional faults
Oceanic ridge
Pyrenean contractional fronts
Magnetic anomalies

Figure 17

contractional front, there is an undeformed latest Cretaceous to early Miocene North Pyrenean foreland basin lying on top of a thinned continental crust. This thinned crust includes a structural high (Landes Plateau) and a deep Late Jurassic-Early Cretaceous extensional basin (the Parentis Basin) bounded southward by a north dipping low-angle extensional detachment that truncated the whole crust and thinned it down to 7 km.

[78] In the western, Cantabrian Domain, the North Pyrenean front is shifted to the north and its geometry clearly differs from the eastern sector. It is a north directed blind thrust imbricated stack detached at the bottom of the continental crust. This thrust system cut Early Cretaceous crustal low-angle extensional faults dipping to the north and would passively transport Le Danois Basin and the coeval extensional riders of upper continental crustal rocks located at the toe of the continental slope (i.e., 3270 Seamount Block). North of the North Pyrenean front, there is a latest Cretaceous-middle Miocene foreland basin overlying a precontractional Albian to Campanian succession that rests upon a very thin (5–8 km) transitional crust. This transitional crust formed during the opening of the Bay of Biscay and is interpreted as the result of the motion of a south dipping detachment fault that, cutting the entire crust, produced the exhumation of serpentinized subcontinental mantle. This exhumation was probably accompanied by the emplacement of extensional riders of upper continental crust on its top and, in the study area, was followed by the development of a Cenomanian to Santonian embryonic oceanic crust along the Bay of Biscay abyssal plain (the South Gascogne Ridge).

[79] The transition between these two well differentiated sectors of the North Pyrenean structure corresponds to a soft transfer zone trending NNE-SSW that is located north of Santander. North of this soft transfer zone, at the NW corner of the Landes Plateau, the boundary between the Basque-Parentis and the Cantabrian domains corresponds to a northeast trending thrust system that places the continental basement of the Landes Plateau over the transitional crust of the Bay of Biscay abyssal plain.

[80] Comparison between this inferred present-day structure of the eastern Bay of Biscay and regional data, enables the interpretation of the extensional rift system developed between Iberia and Eurasia during the Late Jurassic-Cretaceous as consisting of two parallel overlapping arms oriented east-west. The northern arm (the Bay of Biscay arm) included the Bay of Biscay abyssal basin and the Parentis Basin, and the southern one (the Pyrenean arm), the Le Danois, Basque-Cantabrian and the Lacq-Mauleon basins. Both arms appear formed by segmented extensional fault systems that are linked by north to NNE trending accommodation zones. Among these accommodation zones the Santander soft transfer zone that bounded toward the east the transitional crust of the Bay of Biscay abyssal basin and displaced the locus of the major lithospheric thinning from the Bay of Biscay arm to the Pyrenean arm is outlined.

[81] On the other hand, the architecture of the Late Jurassic-Cretaceous basins appears strongly dependent on the presence or absence of ductile middle to lower continental crust beneath the basins. Thus, in the areas where crustal stretching has been minor and there is a ductile middle to lower continental crust (i.e., Le Danois Basin), the basin structure consists of a broad half-graben system of conjugate high-angle extensional faults affecting the upper crust. On the contrary, in the cases in which the lithospheric stretching is larger and the basins lie over a much thinner crust without a ductile middle to lower continental crust (i.e., Parentis, Basque-Cantabrian and the Lacq-Mauleon basins), the basin structure is more asymmetric and is controlled by the development of detachment faults that cut the entire crust. These detachments produce the exhumation of serpentinized subcontinental mantle and can carry extensional riders of upper continental crust directly above the mantle (i.e., Bay of Biscay abyssal basin).

[82] Finally, the interpretation of the MARCONI profiles clearly shows that the Pyrenean contractional structure is strongly controlled by the previous Mesozoic extensional structural grain. Main Pyrenean related contractional deformation is located along the extensionally faulted margins of the preexistent latest Jurassic-Early Cretaceous basins, and the architecture of the orogen appears strongly dependent on the geometry of the previous basin-bounding faults and the related absence or presence of a ductile middle to lower crust. Thus, in the rift segments where this ductile crust was present and the extensional basins were bounded by upper crustal high-angle faults (central and eastern Pyrenees), the Pyrenean orogen mainly consist on a wedge made up by an antiformal stack of upper crustal thrust sheets. Instead, in the segments in which large crustal-scale extensional detachments were present and there was not a ductile crust at the onset of the compression (Western, Basque and Cantabrian Pyrenees), the orogen architecture is simpler and consists on a slightly deformed wedge of upper crustal rocks displaced above the reactivated detachment. On a large and regional scale, it is evident that the extensional Late Cretaceous rift segmentation and changes in the sense of the dip of the extensional Cretaceous detachments control the different characteristics along strike of the Pyrenean orogen in such a way that there is an almost perfect correlation between the extensional Late Cretaceous rift segmentation and the segments of the Pyrenees with a well differentiated contractional structure.

Figure 17. (a) Magnetic anomaly map of the Bay of Biscay [*Sibuet et al.*, 2004b]. (b) Map of the Bay of Biscay, Pyrenees, and surrounding areas illustrating the relationships between the contractional structure of the Pyrenees and the rift system developed between Iberia and Eurasia during the Early Cretaceous. Magnetic picks are derived from *Russell and Whitmarsh* [2003] and *Sibuet et al.* [2004b]. Mesozoic oceanic to transitional and transitional to continental crust boundaries in the Atlantic and northwestern Bay of Biscay areas are based on *Thinon* [1999], *Thinon et al.* [2003], *Tucholke et al.* [2007] and *Jammes et al.* [2009].

[83] **Acknowledgments.** This work was carried out under the financial support of the MARCONI and MODES4D projects (REN2001-1734-C03-02/MAR; CGL2007-66431-C02-02/BTE), as well as the Grup de Recerca de Geodinàmica i Anàlisi de Conques (2001SRG-000074). Research by O. Ferrer was funded by a predoctoral grant from the Ministerio de Educación, Cultura y Deporte (AP2002-0988). The authors also acknowledge support of this research by Landmark Graphics Corporation via the Landmark University Grant Program. The interpretation of the processed seismic data was performed using Kingdom Company software, generously provided as part of an educational grant, by Seismic Micro-Technology.

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