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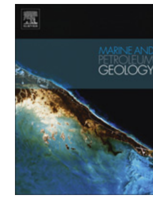

The deep seismic reflection MARCONI-3 profile: Role of extensional Mesozoic structure during the Pyrenean contractional deformation at the eastern part of the Bay of Biscay

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A B S T R A C T

The new MARCONI-3 deep seismic profile allows recognition of the upper crustal structure of the eastern part of the Bay of Biscay and the main features of its Alpine geodynamic evolution. It denotes that the easternmost part of the Bay of Biscay consists of a thick wedge of uppermost Cretaceous to Cenozoic synorogenic sediments lying unconformably on the top of a thinned continental crust with the Mesozoic Parentis Basin to the north and the coeval Landes High to the south.

The Parentis Basin appears as a major half-graben bounded southwards by a north-dipping planar fault. It is filled by a thick sequence of Jurassic–Upper Cretaceous carbonates affected by salt domes and squeezed diapirs made up of Triassic evaporites and mudstones. These salt tectonic structures also affect the overlying uppermost Cretaceous to Lower Miocene synorogenic deposits which are folded upon these structures. The Landes High includes a thin pre-Upper Cretaceous cover tilted to the south. In the Basque shelf, it is deformed by a basement-involving thrust wedge emplaced during the Late Eocene–Miocene that constitutes the North-Pyrenean contractional front.

Geometric relationships and thickness variations depict that this overall structure results from the following.

- Mesozoic extensional stage which includes a Late Jurassic (?)–Late Aptian syn-rift stage in which the Parentis Basin formed; and an Albian–early Late Cretaceous post-rift stage in which diapirs of Triassic evaporites grew close to this major fault.
- Compressive deformational stage coeval to the Pyrenean orogeny which led to (1) the development of a latest Cretaceous up to Middle Miocene foreland basin; and (2) from Late Eocene, the formation of a basement-involving thrust wedge in the innermost foreland basin, the squeezing of the diapiric structures formed previously in the Parentis Basin and, later (Middle Miocene), the partial inversion of some pre-existent faults located in the southern Parentis Basin margin.

This geodynamic evolution together with the structure of the area evidences that the extensional structure resulting from the opening of the Bay of Biscay played an important role both in the location of the North-Pyrenean front and in the North-Pyrenean foreland contractional deformation features. Specially, the lack of significant inversion structures in the Parentis Basin, despite that it belongs to a severely thinned crustal area before the Alpine compression, denotes that the Mesozoic Landes High acted as an important buffer for the propagation of the Pyrenean contractional deformation to the north.

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This deformational buffer was active until Early Miocene and probably vanished afterwards during the last stages of Pyrenean orogen development when some basement faults reactivated in the Parentis Basin.

1. Introduction

The Bay of Biscay is an E–W oriented subsidiary arm of the Atlantic Ocean located between the Iberian Peninsula and the western coast of France (Fig. 1). It belongs to a triangular shape abyssal plain floored by transitional or oceanic crust (Gallastegui, 2000; Gallastegui et al., 2002; Sibuet et al., 2004a,b; Pedreira, 2004; Ruiz, 2007) which is bounded by an extensional margin to the north (Armorican Margin) and a collisional orogen to the south (Pyrenees).

The present-day configuration of the Bay of Biscay is strongly controlled by the relative motion of the Iberian and Eurasian plates which itself is governed by the opening history of the Atlantic Ocean. This history began during the Late Permian–Triassic times with the breakup of Pangea which led to the development of an N–S trending rift along the future Atlantic Ocean, the rifting and opening of the Tethys and also the formation of a ESE–WNW trending rift that connected them along the future Iberia–Eurasia boundary (García-Mondejar, 1989; Ziegler, 1990). The ongoing extension along the Atlantic rift resulted in the opening of an oceanic realm which started first in its southern parts (Jurassic) then in the central and northern ones (Late Jurassic–Cretaceous). Late Jurassic–Late Aptian opening of the North and Central Atlantic Ocean resulted in a transtensional reactivation of the Triassic ESE–WNW rift (Montadert et al., 1979; Vergés and García-Senz, 2000) and then, during the Early Albian–Santonian, the accretion of oceanic crust in the Bay of Biscay (western part of this rift; Williams, 1975; Le Pichon and Barbier, 1987; Thinon et al., 2002; Sibuet et al., 2004a). This process led to the development of a transtensional to extensional plate boundary along the previous Pyrenean rift and as a result the individualization of the Iberian Plate which rotated 35° counterclockwise with respect to Eurasia (Srivastava et al., 1990; Olivet, 1996).

From Late Santonian a faster opening of South Atlantic Ocean produced the northward drift of Africa and as a consequence, the end of the oceanic accretion in the Bay of Biscay and the convergence and later collision between the recently individualized Iberia and Eurasia plates (Rosenbaum et al., 2002; Sibuet et al., 2004b). This drastic change in the relative motion of Iberia generated the Pyrenean orogen from the inversion of the previous extensional structures along the Iberia–Eurasia boundary (Boillot and Malod, 1988; Muñoz, 1992; Alvarez-Marrón et al., 1995; Bourrouilh et al., 1995; Vergés et al., 2002). This orogen, running along the plate boundary, developed until Middle Miocene when relative motion between Iberia and Eurasia ceased or became imperceptible (Srivastava et al., 1990; Roest and Srivastava, 1991; Rosenbaum et al., 2002).

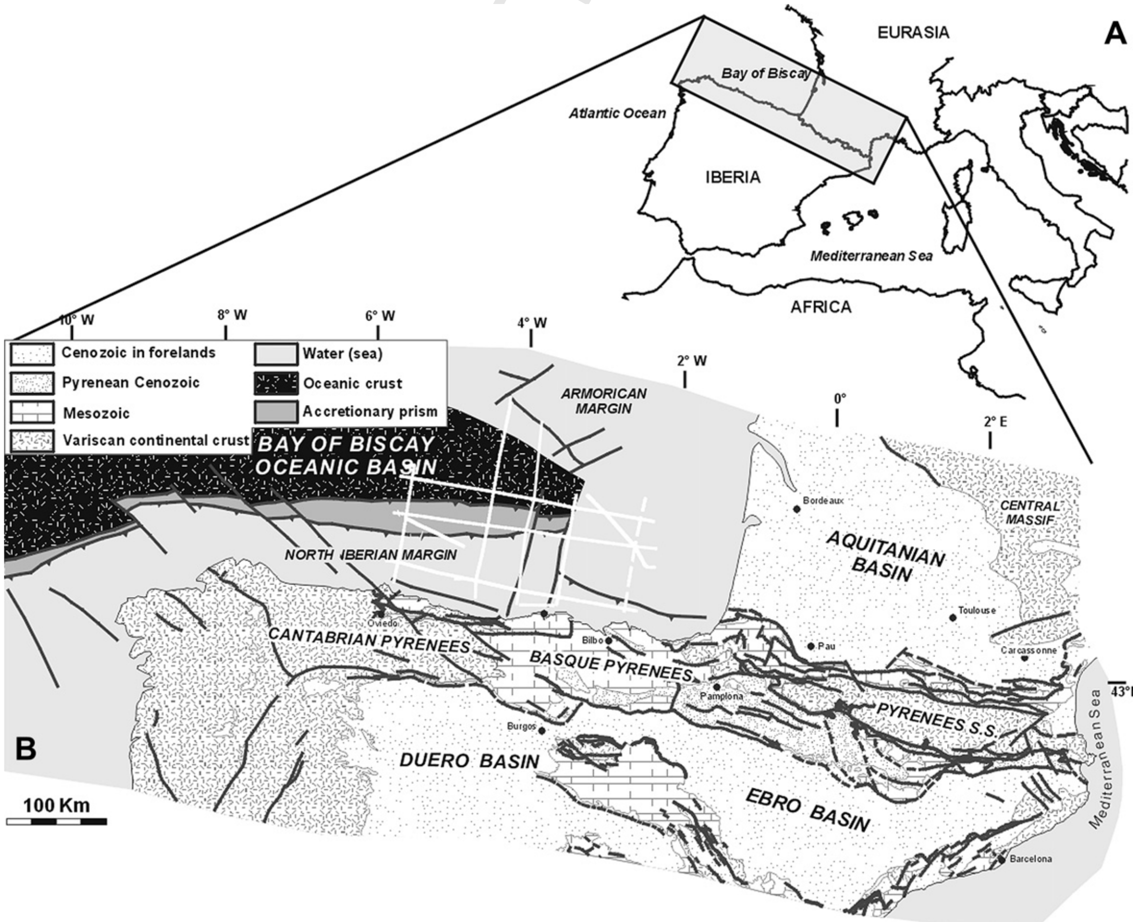

Fig. 1. (A) Location map of the Pyrenees; (B) simplified structural map of the Pyrenees and adjoining areas (modified from Muñoz, 2002).

To understand the fundamental processes that governed the evolution of the Bay of Biscay in this kinematic setting as well as to establish the lithosferic structural features of its eastern and southern parts a deep seismic survey was carried out in September 2003 (Gallart et al., 2004). This survey, called MARCONI (North-Iberian COntinental MARgin), included the acquisition of 11 multichannel deep seismic reflection profiles with a total length of 2000 km and recording signals from a network of 24 OBS/OBH instruments and 36 land stations, all of them located in the south-eastern part of the Bay of Biscay (Ruiz, 2007).

Among all these new seismic profiles our study focuses on the MARCONI-3 deep seismic profile. This profile is located in the eastern part of the Biscay Bay, where oceanic crust was not generated and therefore, in an area in which the north foreland of the Pyrenees is formed by a thin continental crust affected by syn-rift Late Jurassic–Late Aptian extensional structures. These extensional structures, mainly E–W oriented extensional faults, are parallel to the contractional structures developed in the adjoining Pyrenean orogen which developed from the tectonic inversion of the E–W trending Basque-Cantabrian Mesozoic Basin. Taking into account that in the eastern Bay of Biscay the crust was thinner than in the adjoining Basque-Cantabrian Basin (Pinet et al., 1987; Pedreira et al., 2003; García-Mondejar et al., 2004), this paper seeks to better understand why the Pyrenees do not propagate northwards in this area. For this purpose, the MARCONI-3 deep seismic profile has been analyzed in order to (1) document the main structural features of the upper crust in the eastern part of the Bay of Biscay; and (2) discern the role played by the extensional structures formed during the opening of the Bay of Biscay in the location and evolution of the Pyrenean contractional structures.

2. Geological setting

The Bay of Biscay is a deep basin with a V shape opened westwards to the Atlantic Ocean (Fig. 1). It was formed between the Late Barremian and Cenomanian times (Montadert et al., 1979; Barbier et al., 1986; García-Mondejar, 1996; Vergés and García-Senz, 2001; Sibuet et al., 2004a). The bathymetry 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; Sibuet et al., 2004a,b; Pedreira, 2004; Ruiz, 2007), and an eastern part integrated by a shallower platform (1–2 km deep) developed over a 20–25 km thick continental crust (Pinet et al., 1987; Ruiz, 2007). Both relatively deep parts are bounded north-eastwards and southwards by two well-differentiated continental margins. The north-east margin, the Armorican margin, is a Mesozoic extensional passive margin with SW-directed listric normal faults (Montadert et al., 1971; Derégnaucourt and Boillot, 1982; Le Pichon and Barbier, 1987; Thinon et al., 2003). By contrast, the southern one (the North-Iberian margin) is contractional and is part of the Pyrenean orogen that overthrusts the transitional to oceanic crust of the Bay of Biscay abyssal plain (Sibuet et al., 1971; Boillot, 1986; Alvarez-Marrón et al., 1996; Gallastegui, 2000; Gallastegui et al., 2002; Ayarza et al., 2004; Pedreira, 2004), and eastwards the thinned continental crust of the Landes Plateau and Parentis Basin (Cámara, 1997; Gómez et al., 2002).

The Pyrenees form a doubly vergent collisional mountain belt with a thrust system that displays an asymmetric V-shaped upper crustal wedge. They developed over the extensional boundary developed between Iberia and Eurasia during the Late Barremian and Cenomanian. Specifically, they formed along the axis of the intra-continental thinned lithosphere developed in the eastern part of the extensional boundary and, more to the west, over the passive continental margin fringing to the south the Bay of Biscay oceanic crust. As a result of this significant change in the pre-orogenic lithospheric configuration the Pyrenees display different characteristics along strike (Muñoz, 2002; Fig. 1). Between France and Spain they are a continental collisional orogen with a complex antiformal stack of upper crustal thrust sheets developed over a limited subduction of the continental Iberian lithospheric mantle and lower crust underneath the Eurasian plate (Muñoz, 1992). More to the west, in the Cantabrian Pyrenees, they constitute a weakly deformed upper crustal wedge of continental rocks bounded by a very limited south-directed subduction of the Bay of Biscay oceanic lithosphere (Pulgar et al., 1996; Fernández-Viejo et al., 2000) and to the south by the north-directed continental subduction of Iberia mainland.

Recording this structural change, in the Pyrenees *s.s.*, the North-Pyrenean front is made up of basement-involved north-directed thrusts developed from the inversion of the Lower Cretaceous intra-continental basins formed during the opening of the Bay of Biscay (Muñoz, 1992; Gómez et al., 2002). By contrast, in the Cantabrian Pyrenees the North-Pyrenean front belongs to an accretionary prism, located in the front of present-day Cantabrian continental slope, which deforms the sediments filling the Bay of Biscay oceanic basin (Alvarez-Marrón et al., 1995, 1996; Ayarza et al., 2004).

2.1. The eastern Bay of Biscay

In this regional setting, the eastern part of the Bay of Biscay includes the North-Pyrenean front and the adjoining North-Pyrenean foreland. Topographically, it belongs to a relative depth platform flanked to the north by the Armorican shelf and to the south by the narrow and shallower Basque shelf. Along the boundary between the platform and both shelves two major E–W oriented Cap Ferret and Cap Breton canyons are incised (Figs. 2 and 3).

MARCONI refraction and the ECORS Bay of Biscay deep reflection seismic surveys denote that depth to the Moho decreases northwards from about 30–35 km in the Basque shelf to 20–22 km beneath the Cap Ferret canyon (Pinet et al., 1987; Ruiz, 2007). Crustal thinning coincides with the location of the Mesozoic Parentis Basin, northwards the Landes structural High (Figs. 2 and 3). In this basin, the crust below the uppermost Cretaceous–Cenozoic sediments is less than 15 km (Gallart et al., 2004), being 19 km in the ECORS Bay of Biscay (Pinet et al., 1987).

The Parentis Basin, striking E–W, connects westwards with the Bay of Biscay abyssal plain (Figs. 1 and 2). It is filled by a thick (near 10 km) sequence of syn-rift Jurassic–Lower Cretaceous carbonate to terrigenous rocks that overlie a 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., 1997a). All this Mesozoic succession is affected by E-striking normal faults which compartmentalise the basin (Masse, 1997) and bound it both to the north and to the south. It is also deformed by diapirs of Upper Triassic evaporites that pierce both the basin fill (Curnelle and Marco, 1983; Mathieu, 1986; Mediavilla, 1987) and the overlying Upper Cretaceous to Cenozoic synorogenic deposits (Curnelle and Marco, 1983; Bois et al., 1997a; Masse, 1997).

The Landes High is located between the Parentis Basin and the onshore Basque-Cantabrian Basin which developed southwards and was inverted and incorporated into the Pyrenean orogen (Sanchez, 1991). It belongs to a plateau uplifted and eroded with an uppermost Cretaceous–Cenozoic thick sedimentary succession that unconformably overlies the Hercynian basement or a thin and partially eroded Triassic–Jurassic cover (Gariel et al., 1997). This unconformity together with the presence of Albian coastal terrigenous fans of northern provenance along the north Biscay coast (Robles et al., 1988) indicates that this area belongs to a plateau uplifted and eroded during the Middle Cretaceous. However, the absence of Middle Jurassic to Lower Cretaceous sedimentary record

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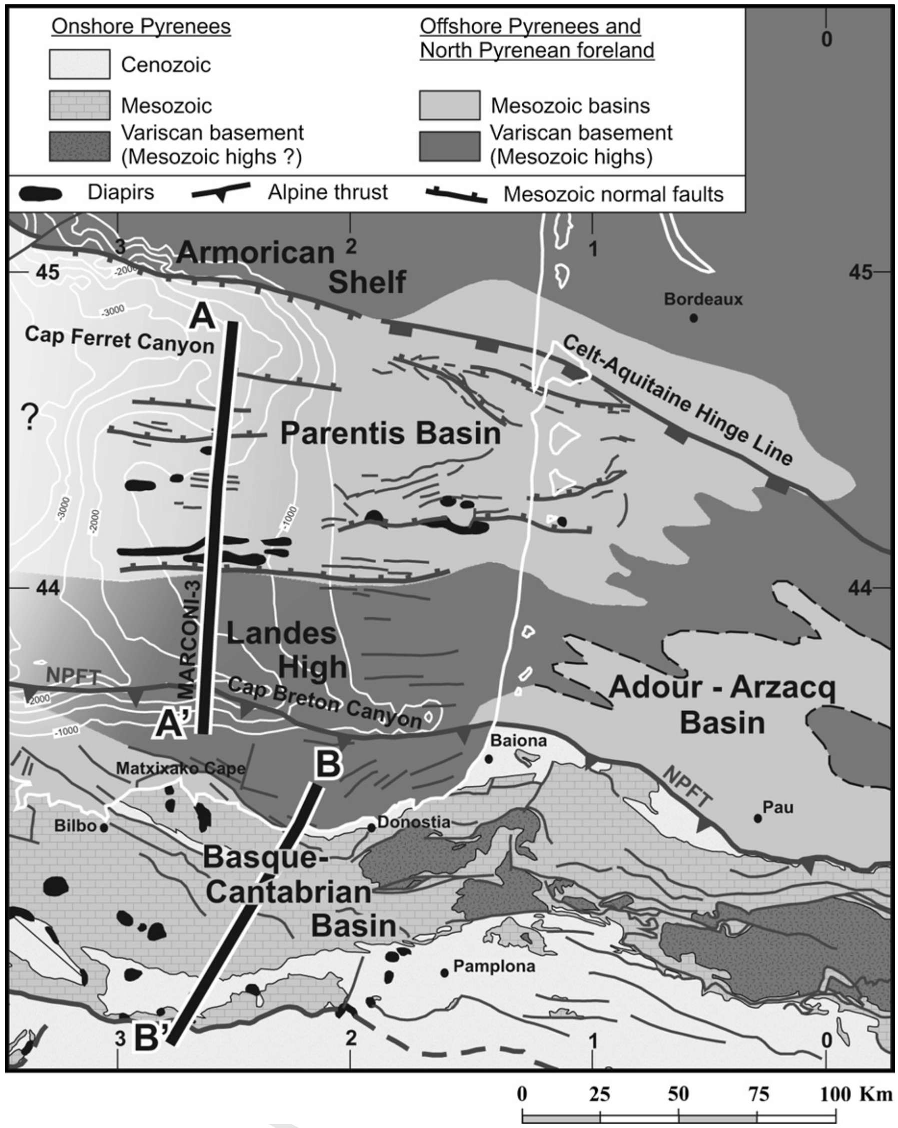

Fig. 2. Simplified structural map of the Basque Pyrenees (modified from Gómez et al., 2002) and Cenozoic subcrop map of the eastern part of the Bay of Biscay (partially compiled from Mathieu, 1986; Bourrouilh et al., 1995). This map displays the distribution of the main Jurassic–Lower Cretaceous basins and highs developed during the opening of the Bay of Biscay. Thick black lines and framed letters show location of MARCONI-3 seismic profile (A–A') and the Pedreira (2004) cross-section across the northern Basque Pyrenees (B–B'). NPFT = North-Pyrenean Frontal Thrust.

both in the Landes High and north Biscay coast prevents knowing if such plateau uplift and erosion started before, during the syn-rift event (Late Jurassic and Albian times).

Both Parentis Basin and Landes High are overlaid by a northwards-thinning thick wedge of Upper Cretaceous to Cenozoic synorogenic deposits which are affected by the North-Pyrenean frontal structures along the Basque shelf (Cámara, 1997; Gómez et al., 2002). These structures consist of north-directed basement-involved thrusts, cover thrust slices and recumbent folds whose location and geometry are largely controlled by the inversion of extensional Early Cretaceous faults bounding the Basque-Cantabrian Basin to the north (Cuevas et al., 1999; Gómez et al., 2002).

3. MARCONI-3 profile

The MARCONI-3 profile is about 122 km long and crosses the eastern Bay of Biscay from south to north. Its southern edge is located in the inner Basque shelf in front of the Matxitako Cape (north of Bilbo), crosses the Cap Breton canyon and the Landes Platform and continues northwards up to the axis of the Cap Ferret Canyon (Fig. 2). Along its trace, the profile cuts perpendicular the main structures recognized in the eastern part of the Bay of Biscay except the northern part of the Parentis Basin and the Armorican extensional margin. Therefore, it images adequately most of the structure of the eastern Bay of Biscay and the main features of its Alpine geodynamic evolution.

3.1. Seismic data acquisition and processing

The MARCONI-3 deep reflection seismic profile were part of the MARCONI survey carried out in September 2003 aboard the Spanish R.V. Hespérides in the south-easternmost part of the Bay of Biscay. In this seismic experiment, crustal reflectivity was investigated from a set of 11 N–S and E–W seismic profiles totaling more than 1800 km of multichannel reflection, and velocity–depth distribution was recorded at wide angles with a network of 24 OBS/OBH instruments and 36 land stations.

The multichannel seismic survey was acquired using a 96-channel streamer with a 2.4-km long active section towed at a nominal depth of 10 m. Hydrophone group interval was 25 m.

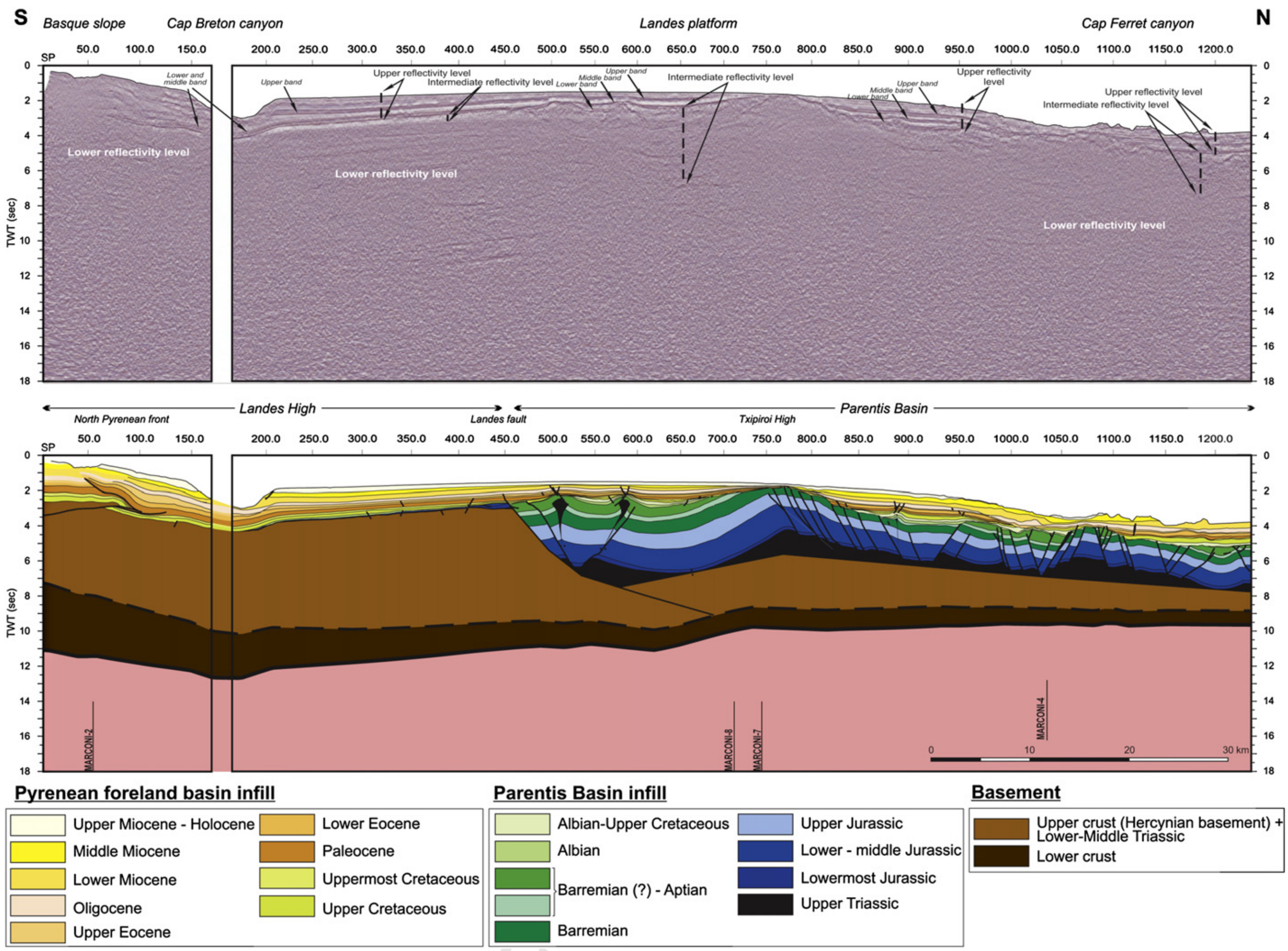

Fig. 3. MARCONI-3 deep seismic profile showing the main stratigraphic units characterized from reflectivity patterns (upper) and interpreted two-way travel time line-drawing of the MARCONI-3 deep seismic reflection profile showing the main geological features of the upper crust in the easternmost part of the Bay of Biscay (lower). Moho and lower/upper crust boundaries are compiled from Muñoz (2002), Pedreira (2004), Ruiz (2007).

Data were recorded at a sampling rate of 4 ms in 18 s long records in order to reach the deep structure. Seismic source was an air-gun array towed at a nominal depth of 8 m and fired every 40 s. This corresponds to a shot-to-shot distance within a range of 100–120 m depending on the vessel speed. This relatively large shot rate was chosen in order to acquire signals from deep levels. As a consequence, spatial aliasing may distort signals in dipping reflections.

MARCONI-3 seismic data processing consisted of conventional steps such as trace editing, deconvolution, filtering, time-variant filtering, normal moveout (NMO) with stacking velocity, stacking and migration and multiple removal process. The latter was focused on reducing the spatial aliasing effects using F-X trace interpolation to increase trace number for each CDP. Before applying NMO correction with stacking velocity, CDP traces were NMO corrected with an adequate velocity function to separate primary from multiple reflections in the F-K domain. F-K filter helped in removing multiple reflections energy before stacking.

3.2. Reflectivity patterns of MARCONI-3 seismic profile

The lateral and vertical variation of the reflection patterns throughout the MARCONI-3 seismic profile allows identification of three superposed reflectivity packages (Fig. 3).

The *upper one* extends beneath the sea floor with a thickness of 1.5–2 s TWT that decreases up to 0.2 s TWT in the central portion of the profile (SP 750). It is highly reflective with closely spaced and generally continuous reflectors which are parallel (Fig. 3) except near the Cap Ferret and Cap Breton canyons where they are shorter and showing frequent cross-cutting relationships. Within this upper level, three bands constituted by reflectors of higher amplitude outline along the profile being the two lower ones assembled in a single thicker band southwards of SP 500 (Figs. 3 and 4).

In the upper Basque continental slope (SP 0–60) the upper reflective level does not appear clearly differentiated. There the reflectivity is very poor and only some near horizontal reflectors at a depth of 2.5 s TWT can be traced southwards to the edge of the profile.

The *intermediate reflectivity level* appears north of Cap Breton canyon just beneath the middle-lower band of high amplitude reflectors described above (Fig. 3). This is also a reflective level but with reflectors showing less amplitude, frequency and lateral continuity. The thickness and reflective pattern of this intermediate level are clearly different north and south of SP 450. Thus, south of SP 450, it is thin (<0.3 s TWT), rather transparent and constituted by a band of discontinuous and low amplitude reflectors. In contrast, north of SP 450, it is much thicker (>4 s TWT), and formed by more continuous and high amplitude reflectors that, locally, appear disrupted by minor transparent and chaotic seismic facies bodies showing inverted-teardrop geometries. The reflectivity along this northern sector decreases gradually in depth and/or vanishes laterally beneath some segments (i.e., between SP 700 and 850).

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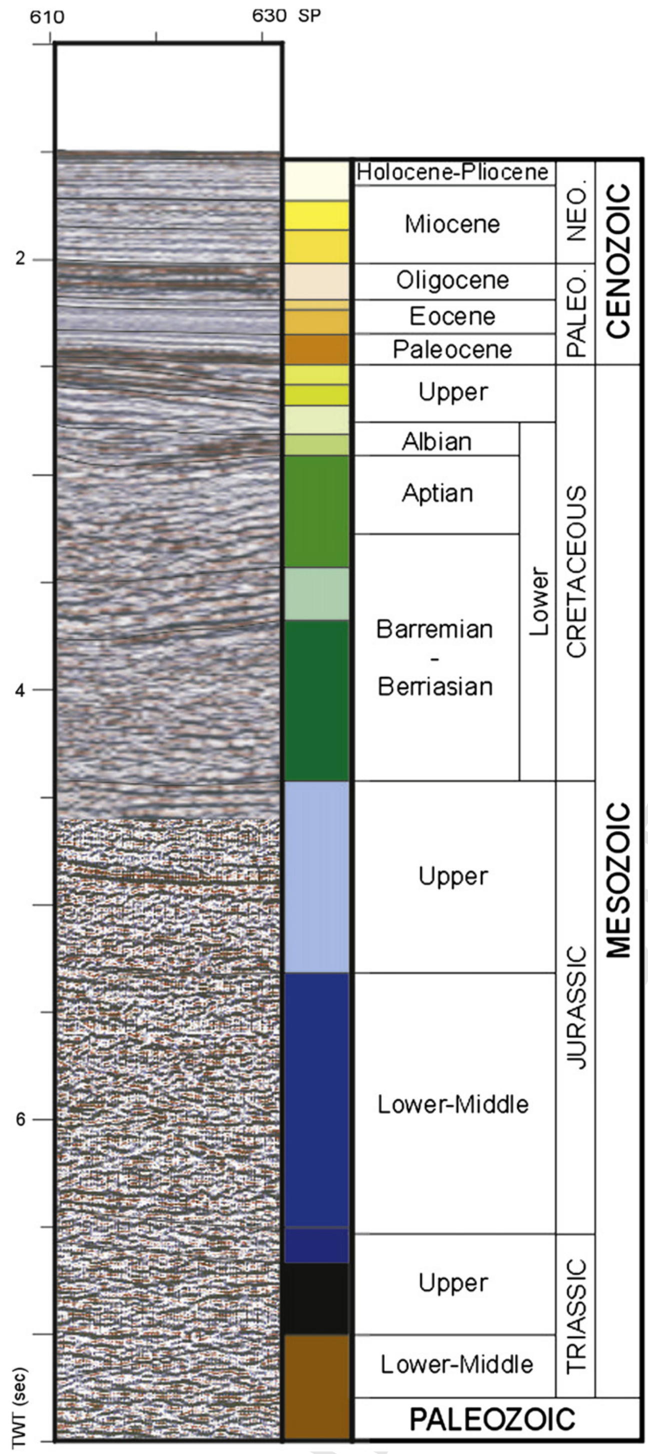

Fig. 4. Stratigraphic units differentiated in the MARCONI-3 seismic profile with their characterized reflectivity pattern.

Finally, the *lower reflectivity level* belongs to the acoustic basement. It is integrated by a chaotic seismic facies which, south of SP 450, includes continuous reflectors that coincide with multiples both of the sea floor and of the reflectors of the upper reflective package. In this lower level, any reflectivity associated with the Moho or lower reflective crust has been identified. The depth of the Moho as well as the thickness of the lower crust has been traced on the basis of the MARCONI refraction data (Ruiz, 2007) and projection from the ECORS Bay of Biscay reflection profile (Pinet et al., 1987; Pedreira, 2004).

The absence of wells along and near the MARCONI-3 profile, and through the whole Landes Platform prevents a precise datation of these three levels and the reflectors depicted in this profile. Nevertheless, from the reflector correlation along conventional seismic surveys intersecting the MARCONI-3 seismic profile, which have been dated using time–depth curves recorded in the wells drilled eastwards in the Parentis Basin, we can decipher that the (1) upper reflective level belongs to the uppermost Cretaceous–Cenozoic sediments; (2) intermediate less reflective level correlates with the Jurassic and middle Upper Cretaceous basin fill of the Parentis Basin; (3) acoustic basement (lower level) is made up of the Hercynian basement and probably also by Permian to Middle Triassic rocks (Fig. 4). The age of the reflectors inside the three reflective levels is rather uncertain in the intermediate reflective level (Mesozoic) where reflectors are less continuous and disrupted laterally chaotic seismic facies. Comparison between Parentis Basin information (wells and profiles) and MARCONI-3 profile also allows the correlation of the transparent to chaotic bodies included in the intermediate level with diapirs formed by Upper Triassic mudstones and evaporites.

3.3. *Profile description*

From the spatial distribution and thickness variation of the reflective levels as well as from the reflectors geometry, two sectors can be differentiated along the profile: the Basque slope–Landes High sector and the western part of the Parentis Basin (Fig. 3).

3.3.1. *Basque slope–Landes High (SP 0–450)*

This sector is characterized by the presence of a thick uppermost Cretaceous–Cenozoic succession unconformably overlying the Hercynian basement or a thin older Mesozoic cover. It is deeply incised by the Cap Breton submarine canyon (Fig. 3).

Some reflectors, mostly parallel to the overlying sediments, have been recognized between the upper reflective level and the acoustic basement north of Cap Breton canyon. They have been attributed to the pre-uppermost Cretaceous Mesozoic cover. These reflectors draw a gentle syncline cut by the overlying sediments (uppermost Cretaceous) north of SP 420, dip southwards and are truncated by horizons of the same uppermost Cretaceous unit beneath the Cap Breton canyon (SP 190) (Fig. 5). The geometric features of this last truncation point out the absence of pre-uppermost Cretaceous Mesozoic cover southwards and therefore the uppermost Cretaceous rests directly upon the Hercynian basement.

Over this Hercynian basement and the thin pre-uppermost Cretaceous Mesozoic cover, the uppermost Cretaceous–Cenozoic package depicts a wedge-shaped geometry thickening southwards. It is made up of several sedimentary units bounded by unconformities as demonstrated by onlap surfaces that, locally, truncate underlying reflectors (Figs. 3 and 5). North of Cap Breton canyon, these units are integrated by continuous reflectors that dip gently southwards. The dip of reflectors is fairly constant along each unit but progressively increases downwards in the older units. However, beneath the Cap Breton canyon and southwards up to SP 110, the unconformities bounding the middle and upper sedimentary units (Eocene–Holocene) show a nested furrow geometry (Fig. 5). The reflectors of the overlying units onlap over the walls of these furrows whereas those of the underlying units are often truncated in the same areas. Also, the reflectors show flat with a raised side or with horseshoe structure morphologies.

From a structural point of view, the uppermost Cretaceous–Cenozoic sedimentary wedge and the underlying rocks only appear affected by some minor normal faults recognized north of SP 300 and by a major but localized thrust wedge with a backthrust in the Basque slope (Figs. 3 and 5). This thrust wedge is located south of the Cap Breton canyon between SP

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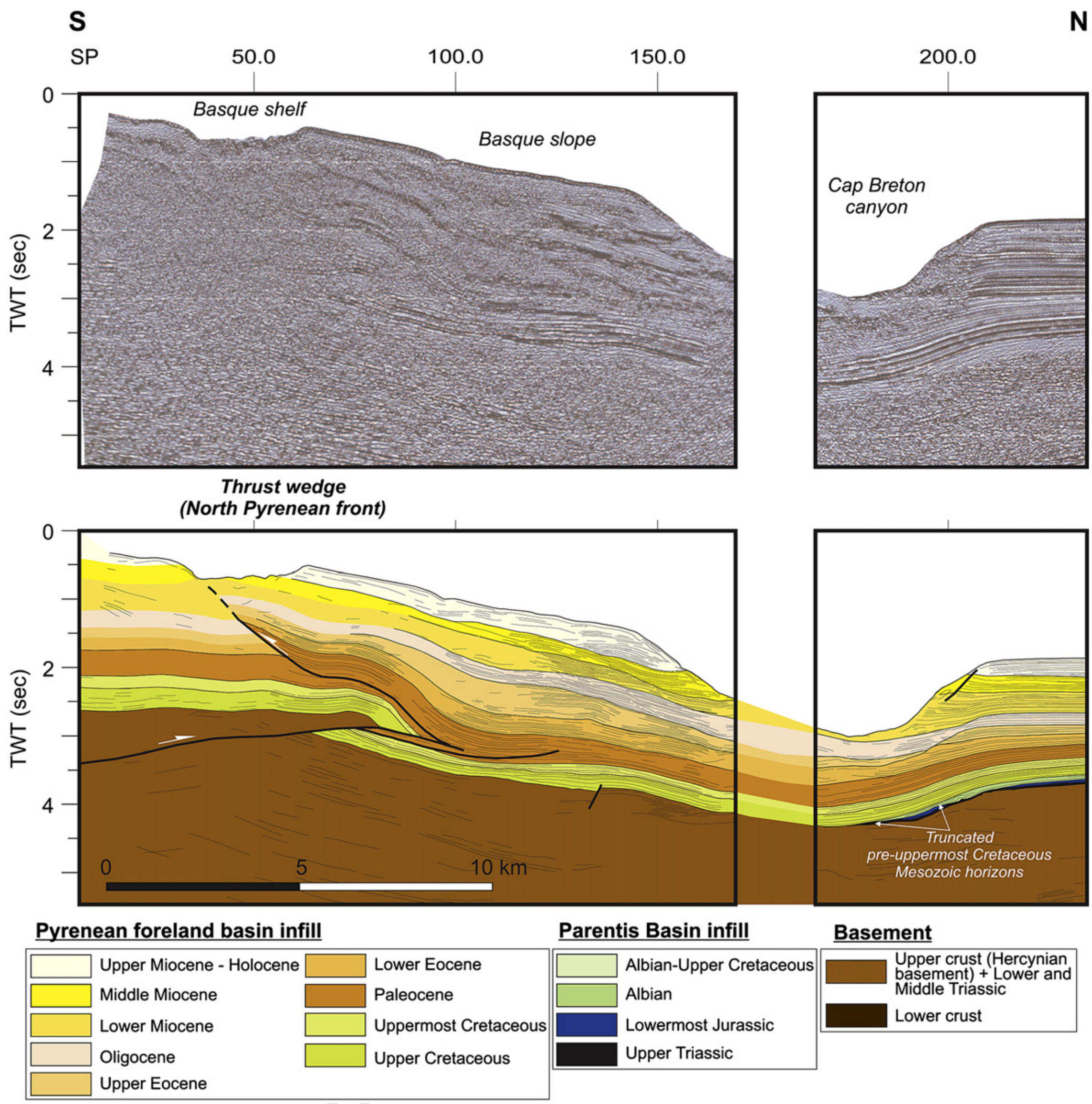

Fig. 5. Southernmost part of the unmigrated MARCONI-3 seismic profile with line-drawing interpretation illustrating the structure of the uppermost Cretaceous–Cenozoic sediments in the Basque shelf and Cap Breton canyon.

50 and 100 (Figs. 3 and 5). It is depicted by north-dipping reflectors on top of a north-verging anticline developed on the hanging wall of a north-directed thrust. The displacement of this thrust has been evaluated to be about 2 km according to the cut-offs of some distinctive reflectors (uppermost Cretaceous–Paleocene). The existence of the upper south-directed backthrust detached into the Paleocene level is inferred from the lack of deformation northwards of SP 110 as well as the ramp geometry observed beneath SP 50 and 65 (Fig. 5). This ramp geometry implies the presence of a south-directed thrust with a displacement greater than 1.5 km that connects with the observed north-directed thrust at SP 110 (tip of the thrust wedge).

3.3.2. Parentis Basin (SP 450–1226)

Unlike the previous sector, north of SP 450, the MARCONI-3 profile depicts a thick Triassic to Upper Cretaceous basin (3–5 s TWT thick) which corresponds to the westwards prolongation of the Parentis Basin (Fig. 2). This basin is overlain by an uppermost Cretaceous–Cenozoic package with frequent thickness and bedding dip variations.

The Mesozoic basin is imaged by a southern segment constituted by a 4 s TWT thick and southward-dipping reflective package and a northern segment where the Jurassic–Upper Cretaceous is thinner (2–2.5 s TWT) and near horizontal. The boundary between these two segments coincides with a major structural high (henceforward called Txipiroi High) that form the pre-uppermost Cretaceous rocks at SP 750.

Along the southern south-dipping segment, the reflective package of Jurassic to Upper Cretaceous deposits shows that the basin fill is structured in two synclines separated, at the upper part, by two anticlines cored by minor transparent and chaotic bodies (Fig. 6). These bodies cut most of the folded Albian and Lower Cretaceous reflectors but not the uppermost Albian and Upper Cretaceous ones. Their shape is not well defined although it seems that they show a broad inverted-teardrop geometry. In relation to this internal segment structure most of the Jurassic to Cretaceous basin fill shows a rather constant thickness and parallel reflectors without any signature of growth sequences. Only the uppermost sequences of the basin fill (Albian–Late Cretaceous) show significant thickness variations and fan shape geometries. In these sequences, thickness is maximum at the

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syncline hinges and decreases both towards the anticlines cored by chaotic bodies and towards the Txipiroi High hinge (Fig. 6). Changes of thickness take place along several unconformities involving truncation of reflectors and onlap above in the uplifted areas (Figs. 7 and 8).

Unlike the previous segment, in the northern segment the Jurassic to Lower Cretaceous basin fill is seismically imaged poorly (Fig. 3) and consequently the location of the bottom of the intermediate level as well as the geometry in some profile portions is not well defined. However, from the reflective segments and crossings with the MARCONI-4 and 8 profiles it is possible to recognize that the Jurassic–Albian successions are cut by both north- and south-dipping extensional faults with major displacements located in two faults that dip towards the south. In this sense, the structure of the Jurassic–Albian package could be envisaged as three major blocks tilted to the north that are bounded by these south-dipping faults (Fig. 3). This extensional fault system does not affect the overlying Upper Albian–Upper Cretaceous successions which unconformably overlie the Jurassic–Albian successions with an irregular and erosive unconformity (Fig. 3). Greater erosions recorded by this unconformity are located in the crest of the tilted extensional blocks where the unconformity surface depicts a convex geometry and is onlapped by the Upper Albian–Eocene sediments (Fig. 9).

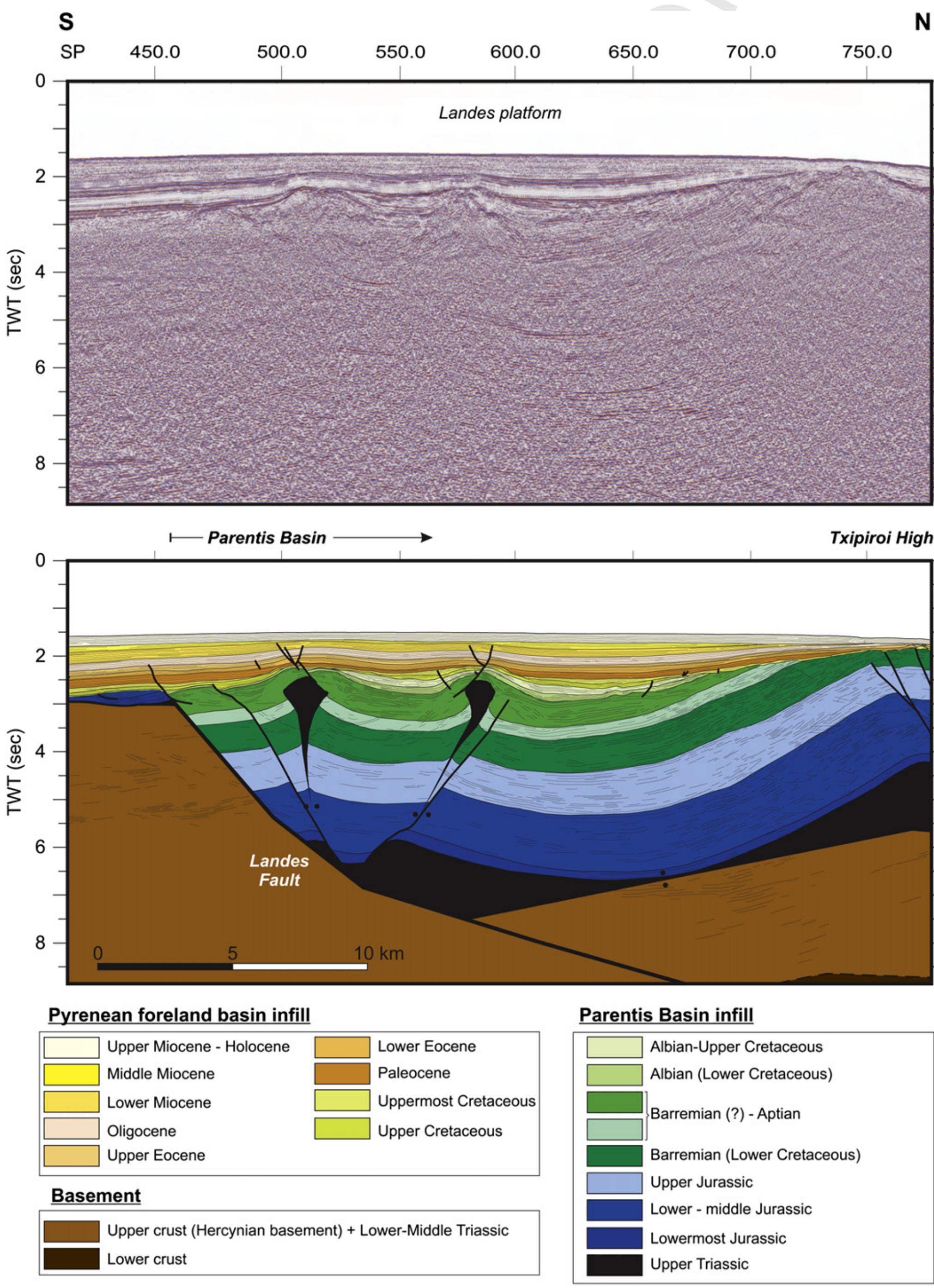

Fig. 6. Details of the unmigrated MARCONI-3 seismic profile in the segment comprised of SP 410 and 775. It shows the structure of the southern part of the Parentis Basin as well as inferred diapir and fault geometries.

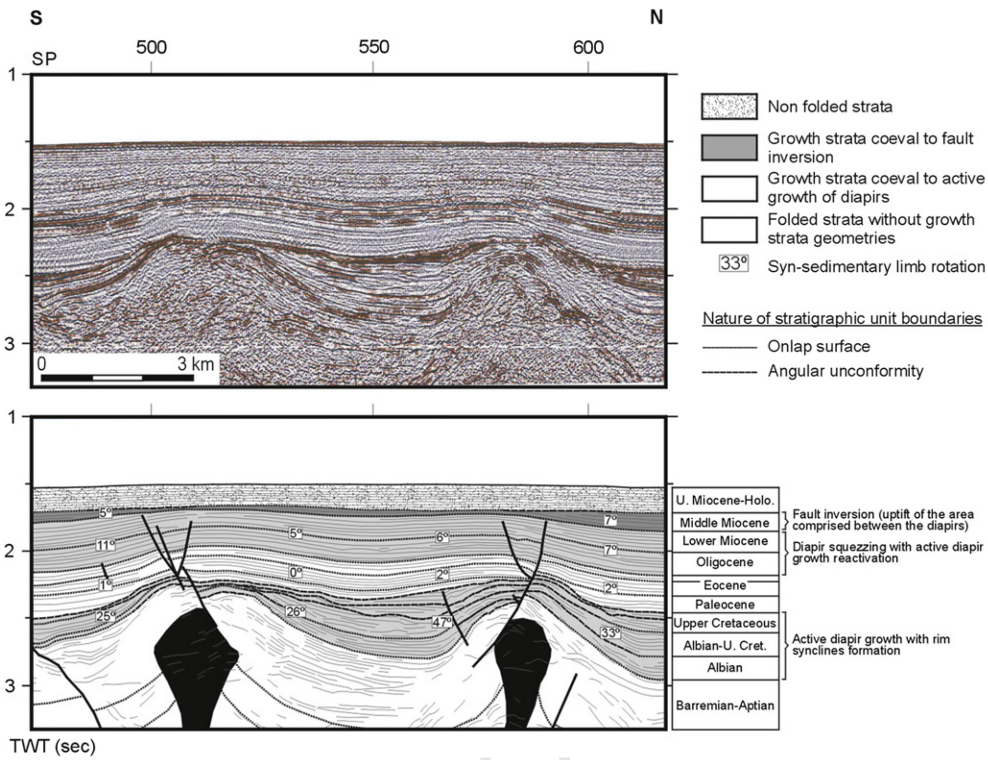

Fig. 7. Geometry of the Cretaceous and Cenozoic sedimentary units in the two diapir-related anticlines observed in the southern part of the Parentis Basin. Reflector rotations as well as thickness variations in their limbs allow deciphering that both diapirs grew actively during the Albian–Upper Cretaceous, were later squeezed during the Oligocene–Early Miocene and were uplifted during the Middle Miocene by the contractional reactivation of the conjugate faults that controlled the Jurassic-Early Cretaceous reactive growth of the diapirs. Note that rotation angles are not real, since the profile has not been depth converted.

As far as the uppermost Cretaceous–Cenozoic package is concerned, it rests unconformable over the Parentis Basin fill with a basal unconformity which is more noticeable on the top of the extensional tilted blocks and above all, on the Txipiroi High, where a significant amount of Mesozoic sediments have been truncated and eroded below (Fig. 6 and 8). It should be noted that this basal surface shows also a relatively higher position over the space comprised between the two anticlines observed close to the southern boundary of the Parentis Basin (Figs. 6 and 7).

Over this basal unconformity, the uppermost Cretaceous–Cenozoic package depicts significant differences north and south of the Txipiroi High, where the package is thinnest being only 0.2 s TWT thick (Fig. 6). South of the Txipiroi High, it is formed by a continuous set of reflectors forming a northwards-thinning wedge together with the coeval reflectors present in the Landes High. By contrast, north of the high it is thicker without any clear wedge geometry and integrated by sub-horizontal, but less continuous, reflectors that are disrupted by numerous internal unconformities (Fig. 9). Emphasizing this thickness, erosive irregular geometry of the sea floor in the Cap Ferret canyon area accounts for a significant removal of the upper part of the Cenozoic package (Fig. 3).

Beneath Cap Ferret canyon (SP 950–1200) the unit bounding unconformities are more irregular and draw a general furrow geometry truncating the underlying reflectors. The truncation and incision of these surfaces is minor in the units attributed to the Upper Cretaceous but becomes major in the ones located at the bottom of the Lower Eocene and younger units (Fig. 9).

Unlike the older rocks, the uppermost Cretaceous–Cenozoic is only affected by few minor normal faults and locally by the small wavelength anticlines which, cored by chaotic bodies, are present at SP 500, 580 and 890 (Fig. 3). On top of these anticlines, as well as on top of the Txipiroi High, the uppermost Cretaceous–Cenozoic package thins. Regardless, it unconformably rests upon the truncated anticlines (Figs. 6 and 7).

In relation to these local thickness variations, the uppermost Cretaceous–Cenozoic package can be divided into (Figs. 7 and 8):

- an upper unit (upper Miocene–Holocene) of relatively constant thickness, made up of flat and sub-horizontal reflectors. It unconformably overlies the underlying successions on the limbs of the anticlines and the southern limb of the Txipiroi High.
- a middle unit (Middle Miocene) that is significantly thinner between the two anticlines present, close to the southern margin of the Parentis Basin and that is erosively truncated in the southern limb of the Txipiroi High. This unit is slightly tilted along this last limb and also on the margins of the area comprised between these two anticlines. In these areas the reflectors show a progressive decrease of their dip upwards denoting a syn-sedimentary limb rotation.
- a lower package (uppermost Cretaceous–Lower Miocene) integrated by several units bounded by onlap surfaces that are: folded upon the anticlines; tilted over the Txipiroi High; and affected by normal faults over the crest of the anticlines. Tilting of the reflectors is constant up to uppermost Eocene but decreases progressively upwards in the younger units. The thickness of this lower package also decreases towards the crest of anticlines and the Txipiroi High. Such thickness variations are more noticeable at Txipiroi High and in the lower units (latest Cretaceous–Paleocene age) that onlap and overlap the irregular basal unconformity of the package (Figs. 7 and 8).

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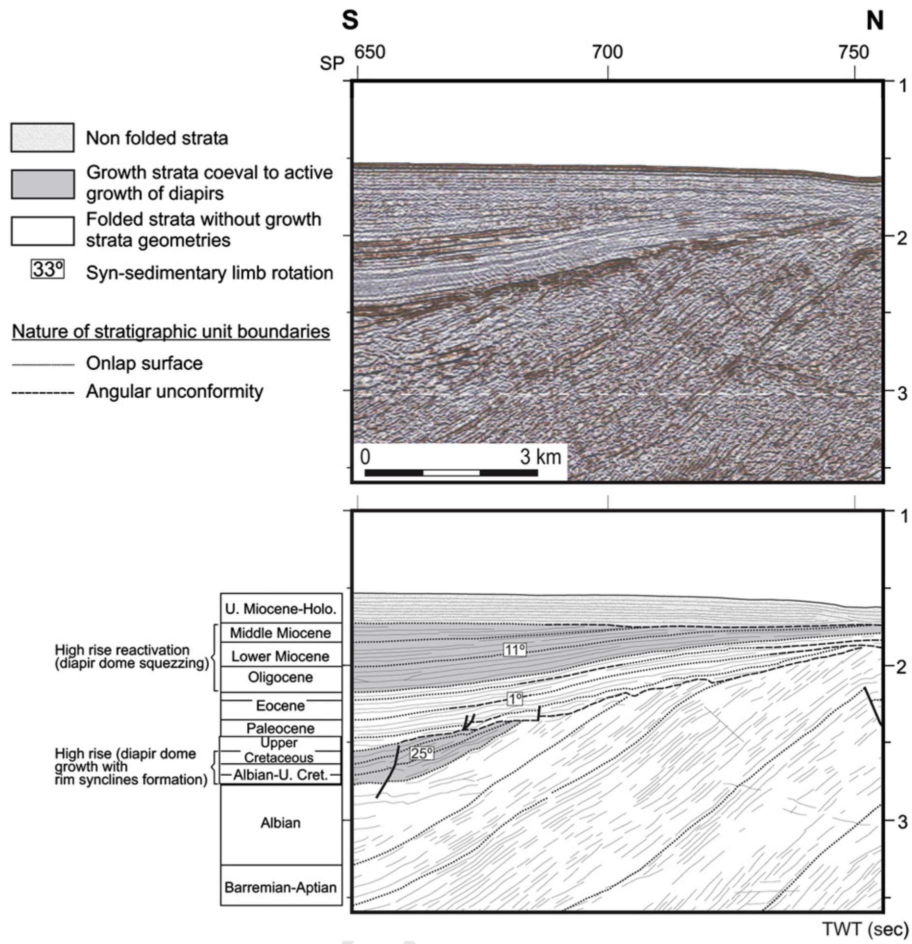

Fig. 8. Geometry of the Late Jurassic and Cenozoic sedimentary units in the southern limb of the Txipiroi High. As in the two diapir-related anticlines shown in Fig. 7, reflector rotations as well as thickness variations in this limb denote that the Txipiroi High grew actively during the Albian–Upper Cretaceous and reactivated later by squeezing during the Oligocene–Middle Miocene. Note that rotation angles are also not real, since the profile has not been depth converted.

4. Geological interpretation of the MARCONI-3 profile

4.1. Salt tectonics

According to their seismic signature and geometric features, the small wavelength anticlines located at SP 500, 580 and 890 (Fig. 3), have been interpreted as related to the rise of evaporite diapirs (Fig. 6). The transparent and chaotic bodies coring the anticlines show seismic facies as the diapirs observed eastwards in the Parentis Basin (Gailhard et al., 1971; Curnelle and Marco, 1983; Mathieu, 1986; Mediavilla, 1987; Bourrouilh et al., 1995; Biteau et al., 2006) as well as southwards in the Pyrenees (Brinkmann et al., 1967; Serrano and Martínez del Olmo, 1990). As them, the MARCONI-3 interpreted diapirs would be formed by Upper Triassic mudstones and evaporites.

The presence of these diapirs implies that, beneath the reflective Jurassic–Cretaceous package, the basin has to include an autochthonous Upper Triassic evaporitic layer. Its geometry is highly speculative and only seems constrained between the SP 630 and 700. Here, beneath the bottom of the reflective Jurassic a non-reflective unit appears, bounded underneath by a trend of less tilted reflectors which joints the bottom of the reflective Jurassic just below the syncline hinge (Fig. 6). These lower and less tilted reflectors could belong to the Lower–Middle Triassic rocks that underlie the Upper Triassic evaporites and, consequently, the non-reflective unit would represent the Upper Triassic evaporite layer. The lack of this unit beneath the Jurassic syncline hinge would indicate, in such a case, the presence of a primary salt weld at this point. Outside of this "constrained" area the bottom of the Upper Triassic has been drawn extending the top of the inferred Lower–Middle Triassic reflectors with the same dip (Figs. 3 and 6).

The divergent patterns of the Albian–Upper Cretaceous reflectors towards the hinges of the synclines fringing the diapir-related anticlines as well as the fact that these reflectors are truncated by the diapirs demonstrate that diapir emplacement took place during the Albian–Late Cretaceous. Accordingly, the synclines have been interpreted as rim synclines associated with the active growth of these structures.

On the other hand, the inverted-teardrop bulb shape of the diapirs above a secondary near-vertical weld denotes that these diapirs were squeezed later on. The age of this squeezing is recorded by the reflector geometry of the sediments affected by the anticline developed upon the diapirs from the upward expelling of the viscous material. This geometry (see Section 3.3.2) clearly shows that the squeezing of the diapirs took place from Oligocene to Early Miocene (diapirs located at SP 500 and 580) and during the Eocene–Oligocene (diapir located at SP 890).

4.1.1. Structural origin of Txipiroi High

The high that appears between the SP 640 and 850 is one of the most outstanding structural elements depicted by the MARCONI-3 profile as already discussed (Fig. 3). The seismic signature of the MARCONI-3 profile, however, only allows observing the structure (geometry) of this high in the uppermost Cretaceous–Cenozoic

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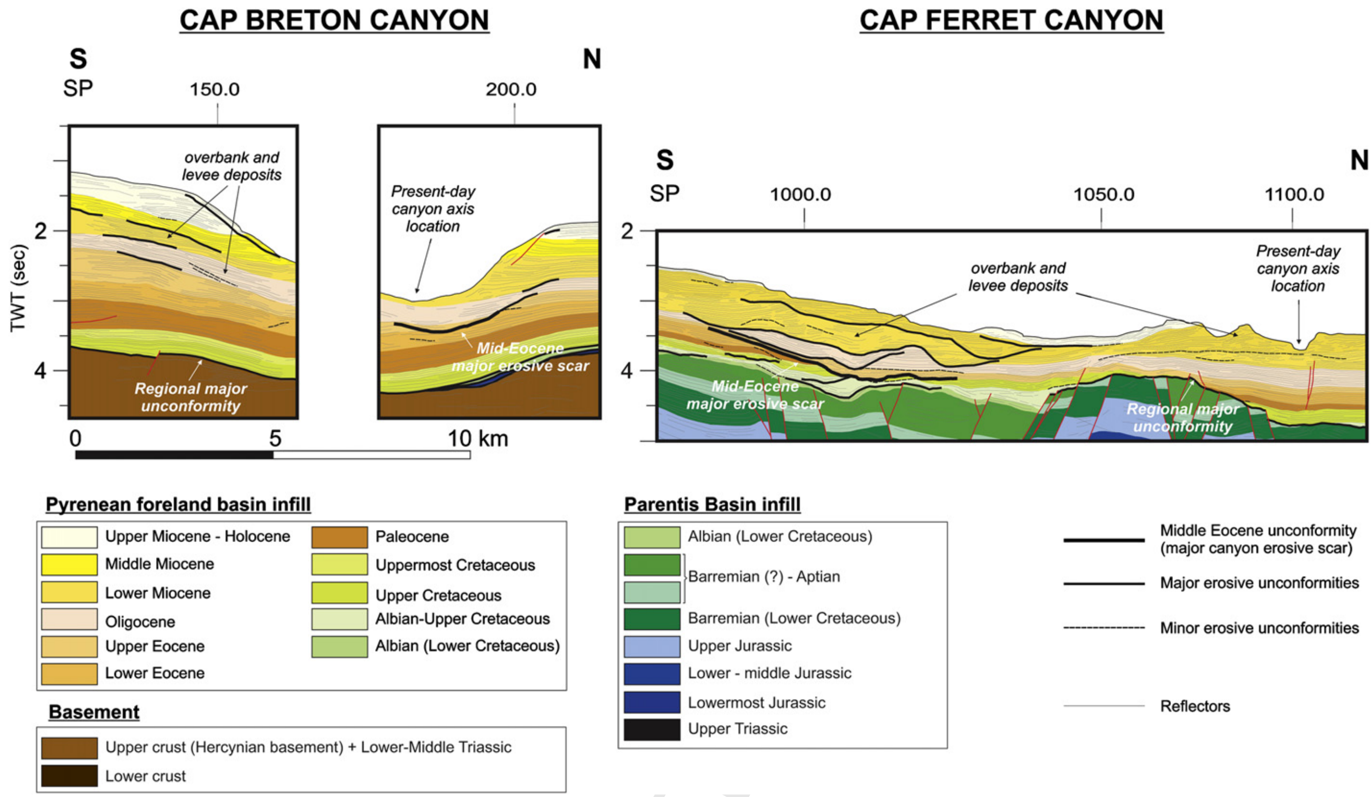

Fig. 9. Seismic signature of the Cap Breton and Cap Ferret canyons along the MARCONI-3 profile. Comparison between the geometry of Cretaceous–Cenozoic reflectors beneath the canyons allows recognizing that although it formed before (Upper Cretaceous), Cap Ferret canyon mainly developed from Late Eocene when Cap Breton canyon was also created.

package and beneath, in Jurassic–Upper Cretaceous successions cropping out in its southern flank. Therefore, it does not show the structural features at deeper levels and specifically the geometry of the top of the basement beneath the high.

In the absence of other data the interpretation of the deep structure of the Txipiroi High, in consequence, must be done indirectly from the information provided by the MARCONI-3 profile in its more superficial part (Figs. 3 and 8). This information (reflectivity pattern and reflectors geometry) shows that

- the uppermost Cretaceous–Cenozoic sediments thin towards the crest on both sides of the High with a horizontal arrangement that indicates progressive rotation of both high flanks during the deposition of the Oligocene–Lower Miocene successions;
- on the southern flank of the High the south-dipping uppermost Cretaceous and Paleocene reflectors lie unconformably over a truncated Lower and Upper Cretaceous succession significantly more tilted to the south;
- the Albian and Upper Cretaceous, observed at the toe of its southern flank, also thins towards the crest of the High with reflectors showing a fan shape that again denotes rotation of this flank during their sedimentation;
- finally, beneath all these successions the High is made up of the Jurassic–Aptian basin fill of the Parentis Basin that draws an antiform with a southern limb formed by horizons dipping to the south and a northern limb cut by several north-dipping normal faults.

This data denote that the High grew with progressive rotation of its flanks during two well-differentiated episodes: an Albian–Late Cretaceous episode and a younger one (Oligocene–Early Miocene in age) during the foreland basin development (Fig. 8). These episodes were coeval with the two stages of diapir activity deduced more to the south in the profile (Fig. 7). Also, it indicates that it was a positive relief submitted to erosion during the Albian–Late Cretaceous; a period in which the whole area of the Bay of Biscay was not submitted to compression (Mathieu, 1986; Masse, 1997; Muñoz, 2002; Biteau et al., 2006). Finally, it reveals that the formation of this Albian to Upper Cretaceous positive relief was accompanied by the development of normal faults on its northern flank that cut the Upper Cretaceous and not the overlying uppermost Cretaceous–Cenozoic successions.

Taking into account all these considerations, we interpret that the Txipiroi High corresponds to a saline dome of Upper Triassic diapiric rocks which was formed during the Albian–Late Cretaceous. It was subsequently squeezed as the diapirs during the Oligocene–Early Miocene, controlling the location of the foreland fore-bulge. This saline dome would be developed in the crest of the roll-over generated by the normal displacement of the Landes Fault that bounds southwards the Parentis Basin, as it is attested by the coincident position of the high with the passive axial surface that presents this fault in its hanging wall (see cross-section in Fig. 3).

In relation to this interpreted origin for the Txipiroi High, the geometric features of major highs observed between SP 950 and 990 and between SP 1060 and 1100 (positive relieves formed with progressive limb rotation during Albian–Upper Cretaceous and submitted to erosion during the lattermost Cretaceous) also point to they rose from the accumulation of Upper Triassic diapiric rocks beneath the Jurassic-Lower Cretaceous overburden. Concretely, fault geometry and distribution point to that these two highs grew from the accumulation of these diapiric rocks under the cutoff of its top in the footwall block of the main south-dipping extensional faults (Fig. 3).

4.2. The extensional structure: geometry and evolution of the Parentis Basin

In relation to the structure of the Parentis Basin, the description and geometric analysis of the MARCONI-3 profile presented in the previous section bring up to two structural questions which cannot be directly solved from the observed reflectivity: the geometry of the main fault bounding the Parentis Basin to the south as well as the extensional fault system and the significance of the Txipiroi High during the extensional event. Both items are discussed below, proposing for each one a solution that tries to fit with the available data.

4.2.1. Geometry of the extensional fault system of the Parentis Basin

The sudden change between the thick (>5 s TWT) Jurassic to Aptian fill of the Parentis Basin and the extremely thin (<0.3 s TWT) time equivalent successions on the Landes High demonstrates that the southern boundary of the Parentis Basin is a major fault (Landes Fault; Fig. 6). However, the MARCONI-3 does not depict any significant fault. The southernmost imaged Mesozoic reflectors of the Parentis Basin are affected by the diapir-related anticlines. Taking into account that diapir development precise changes in the overburden thickness but not erosive surfaces and overburden thickness variations are observed in this profile segment, we interpret that these diapirs formed from extensional faults. Location of these faults (not observed in the profile, although they have been seen in the parallel V80 conventional seismic profiles) has been done considering that (1) the fault trace should not cut the observed continuous reflectors; and (2) in a extensional faulting setting diapirs rise in the areas where the overburden thickness is minor (cutoff of the top of the evaporitic source layer in the footwall block of the extensional faults). According to this hypothesis, the location of southernmost diapir implies the presence of a deep Upper Triassic evaporite layer in the footwall of the related extensional fault and, therefore, that the major bounding fault of the Parentis Basin has to be located more to the south (Fig. 6). The southernmost location of this fault is constrained by the reflectors attributed to the Jurassic at the northern end of the Landes High (Fig. 6). Considering the location of the reflectors attributed to the bottom of Jurassic on both sides of the fault, the dip of the fault is higher than 40° and the fault displacement generates a fault dip separation of the base of the Jurassic along the MARCONI-3 profile on the order of 15 km (Fig. 3).

There are not conclusive arguments with respect to the geometry of the Landes Fault and its related extensional fault system. The main thickness variations occur at expenses of the Upper Jurassic–Albian successions. However, no fan attitude or growing geometry has been observed. The thickness of Jurassic to Aptian successions is rather constant (5 s TWT) south of the Txipiroi High. Northwards, it continues to be constant but much thinner (3 s TWT).

The Txipiroi High crest also separates two panels in which the Lower–Middle Triassic and the concordant top of the Hercynian basement depict a distinct inclination (Fig. 3). South of the Txipiroi High crest this is beneath the major syncline, they are planar and inclined towards the south. By contrast, north of the crest the roughly near horizontal arrangement of the top of the Upper Cretaceous together with the lack of significant Cretaceous–Jurassic thickness variations indicates that they are near horizontal.

These data denote that the fault bounding southwards the Parentis Basin is not listric but composed of three planar segments separated by flattening bends. The dip (α) of the lower segment could be approximately evaluated from the fault displacement and the thickness of the basin fill north of the Txipiroi High applying the Pythagoras theorem: arcsin $\alpha = 5$ km (basin thickness)/15 km (displacement) $= 20°$. On the contrary, the dip of the upper segment is more uncertain: the location of the reflectors attributed to the bottom of Jurassic on both sides of the fault only allow to precise that it must higher than 40°.

Applying the Coulomb shear model developed by Xiao and Suppe (1992), we can also infer that the (1) fault bend has to be located southwards of the top of the basement cutoff in the hanging wall, probably close the plane which defined by the antithetic fault; and (2) Txipiroi High coincides with the position of the fold hinge that marks the northern boundary of the hanging wall rocks sheared along the fold bend and translated away from the bend by slip on the lower fault segment (inactive axial surface). According to the Xiao and Suppe (1992) model, this axial surface must dip southwards along the pre-extensional rocks with an angle equal to the Coulomb shear angle that have these rocks in extension and be more vertical or even dip towards the opposite dip direction along the syn-extensional rocks. However, the geometry of the reflectors along the MARCONI-3 profile as well as the lateral extension of the panels in which the thickness of Jurassic–Upper Cretaceous keeps constant only allow to deduce that this axial surface is rather vertical along these syn-extensional rocks. This steep dip indicates that during fault activity deposition rates were high relative to the subsidence rates generated by the fault slip (Xiao and Suppe, 1992; Shaw et al., 1997).

The rest of the extensional faults are also planar, both non-rotational and rotational as deduced from the geometry of the syn-rift sequences and their relationships with the post-rift.

With the deduced geometry of the Parentis extensional fault system a problem lies on the configuration of the basement rocks, as they do not simply balance at both sides of the main bounding fault. One possible solution would be to invoke a pure shear thinning of the crustal rocks below the Parentis Basin as observed in the ECORS Bay of Biscay profile (Pinet et al., 1987; Masse, 1997) or an intracrustal shear zone detachment model (Lister and Davis, 1989). A question also arises about the possible obliquity or strike slip component of the main southern fault.

4.3. The uppermost Cretaceous–Cenozoic foreland basin (the upper reflective level)

The eastern part of the Bay of Biscay shows the development of the North-Pyrenean foreland basin since the later most Cretaceous, which formed on a crust affected by a previous extensional thinning. The present-day geometry of the foreland basin, as well as its evolution, has been strongly influenced by the location of the extensional basins, the subsequent evolution and the related crustal structure at depth. As a result, the North-Pyrenean foreland basin shows two different parts along the MARCONI-3 profile. In the southern part, south of Txipiroi High, the uppermost Cretaceous–Cenozoic sediments constitute a southwards-thickening wedge with a depocentre controlled by the location of the tip of the Pyrenean frontal thrust wedge. Taking into account that sediment transport direction was practically perpendicular to the profile (Gaudin et al., 2003, 2006; Bourillet et al., 2006), this denotes that the deposition of the uppermost Cretaceous–Cenozoic sediments was synchronous with a progressive flexure of the crust towards the south, coeval with basin inversion and northwards thrusting of the Basque-Cantabrian Basin (Fig. 5 and 11). In the northern part, north of Txipiroi High, uppermost Cretaceous–Cenozoic sediments do not follow the pattern of a flexured basin. Subsidence is higher than in the southern part and several depocentres are distributed along the basin. They are controlled by the previous extensional geometry and the erosional features related with the Cap Ferret canyon evolution. This part of the basin developed on top of the Parentis Basin, far away from the Pyrenean orogen and over a lithosphere significantly thinned a short time before (10 Ma) during the Late Jurassic–Early Cretaceous (Sibuet et al., 2004b). In this context the observed high subsidence could be related to the

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thermal cooling of the recently thinned lithosphere, an interpretation that fits with the subsidence and thermo-mechanical analyses done in the eastern part of the Parentis Basin (Brunet, 1986, 1991; Desegaulx et al., 1991). These analyses show a significant and asymptotic decay of the tectonic subsidence during the lattermost Cretaceous–Cenozoic which, distorted by other processes (Eocene compression, plate rigidity and/or thermal regime variations), is related to the thermal re-equilibration of the lithosphere thinned during the Late Jurassic–Early Cretaceous rifting (Brunet, 1991). Thus, a question arises if the northern part of the uppermost Cretaceous–Cenozoic basin observed in the MARCONI-3 profile has to be considered a part of the foreland basin or not, as no evidence of subsidence by thrust loading is recorded.

The boundary between these two distinct parts of the uppermost Cretaceous–Cenozoic basin is surprisingly located in the middle of the Parentis Basin, in the Txipiroi High. This High could be considered as the fore-bulge of the North-Pyrenean foreland basin and also the hinge line for the more subsiding northern part of the basin. It represents the boundary between the southern part of the Parentis Basin which has been slightly deformed and tilted during the Pyrenean convergence and its northern part which coevally subsided from thermal re-equilibration of the lithosphere thinned during Late Jurassic–Early Cretaceous rifting.

The presence of a widespread unconformity at the bottom of the uppermost Cretaceous unit together with the limited thickness of uppermost Cretaceous (Fig. 7) denotes that the southern parts of the profile (Landes High and southern Parentis Basin) were uplifted during the uppermost Cretaceous. However, main Pyrenean-related contractional deformation along the profile is recorded by the Eocene–Middle Miocene sediments. First contractional deformation is recorded in the North-Pyrenean frontal thrust wedge located beneath the Basque slope where thickness and reflector dip variations in the foreland syncline related to this wedge denote that it developed between the Eocene and Early Miocene. More to the north contractional deformation began at the Oligocene with the squeezing of Txipiroi saline dome and the two diapirs located in the southern boundary of the Parentis Basin. Anticline growing related to this diapir and salt dome squeezing persisted up to the Early Miocene. Subsequently, at middle to early Late Miocene times the whole area between the two diapir-related anticlines was uplifted because of the inversion of the two conjugate normal faults that controlled initially the location of the diapirs. Deformation ceased at Late Miocene times as revealed by the continuous thickness of the upper Miocene–Holocene sediments and the unconformity at their bottom (Figs. 7 and 9).

4.3.1. *Age of Cap Breton and Cap Ferret canyons*

Formation and evolution of the Cap Breton and Cap Ferret canyons is rather well constrained by the seismic signature of the MARCONI-3 profile (Fig. 9). Reflectors' geometry denotes that Cap Breton began to be formed during the middle part of the Eocene when the older erosive scar observed beneath the present-day location of the canyon developed. This erosive scar truncates the Lower Eocene packages and is partially covered by onlapping Upper Eocene reflectors. Before this scar development, there is no evidence of the existence of a canyon in this area: the Upper Cretaceous–lowermost Eocene successions do not show thickness changes and the reflectors are continuous, parallel and with a constant and gentle south-directed dip (Fig. 9).

Regarding the latter Cap Breton canyon evolution, the profile shows that until the Middle Miocene the depositional processes in the valley floor prevailed (sedimentation of overbank, levee and mass flow deposits) leading to a progressive smoothening of the channel incision. Later on, at an imprecise age, the canyon valley became erosional until now. During this last period, predominance of erosive processes on the canyon results in the lack of post-Lower Miocene deposits in the valley floor and the development of mass-gravity structures on their walls.

In relation to the Cap Ferret canyon, reflector geometry shows that its development did not took place clearly until the formation of a major scar during the Middle Eocene. This wide and incised scar is totally onlaped by Upper Eocene reflectors and erodes partially the uppermost Cretaceous and all Paleocene and Lower Eocene packages between SP 980 and 1090 (Fig. 9). Beneath this major scar some minor Late Cretaceous erosive scars also appear in the most depressed parts of a major Jurassic–Albian half-graben (Fig. 9) that are onlapped by Upper Cretaceous and truncate as older than the Albian reflectors. However, these scars are smaller and much less incised than the Middle Eocene and younger ones suggesting that they probably are not related to the Cap Ferret canyon development but to a transport network formed previously during the thermal subsidence of the Parentis Basin.

From this Middle Eocene scar development, the MARCONI-3 profile shows that Cap Ferret canyon migrated northwards and that, although new erosive scars formed, depositional processes have prevailed in the valley floor leading to a progressive smoothening of the channel incision.

As a conclusion, the MARCONI-3 profile denotes that both Cap Breton and Cap Ferret canyons started to develop during the Middle Eocene, although a precursor of the Cap Ferret canyon already existed. These canyon ages are much older than that proposed previously by Dauvillier (1961) and are similar to those proposed by other authors who indicate that the age for the Cap Breton canyon can be Late Eocene (Cirac et al., 2001) or even older (Schoeffler, 1965).

The fixed location of the Cap Breton canyon probably resulted from the relatively fixed position of the frontal North-Pyrenean thrust wedge tip, whereas the migration of the Cap Ferret results from the non-localized thermal subsidence of the northern part of the uppermost Cretaceous–Cenozoic basin.

5. Evolution of deformation along MARCONI-3 profile

The description and geometric analysis of the MARCONI-3 profile show that the present-day structure of the eastern part of the Bay of Biscay results from the succession of two well-differentiated deformational stages (Fig. 10): an initial one which was extensional and a younger compressive one.

The first stage is coeval to the opening of the North Atlantic Ocean and the Bay of Biscay. It includes a syn-rift stage in which the Parentis Basin formed from the normal motion of a major north-dipping fault located along its southern boundary and a post-rift stage in which diapirs of Triassic evaporites grew close to this major fault. These diapirs led to the development of rym-synclines at their flanks and their development was accompanied by the formation of a major saline dome over the crest of the roll-over generated by the normal displacement of the Landes Fault.

The MARCONI-3 profile together with available well data dates the post-rift stage as Late Albian–early Late Cretaceous and show that the syn-rift stage ended during the Middle Albian and was active, at least, since the Late Jurassic. The poor reflectivity shown by the seismic profiles beneath the Jurassic and the scarcity of wells cutting older Jurassic rocks prevents knowing if the syn-rift activity took place also before, and therefore, dating the age in which this activity started. However, data compiled in the other Pyrenean Mesozoic basins (Organyà, Mauléon-Lacq and Basque-Cantabrian basins) denote that rifting activity did not begin before the Late Jurassic (Curnelle and Dubois, 1986; Berástegui et al., 1990; Hernández et al., 1999; Vergés and García-Senz, 2001; García-Senz, 2002). Consequently we assume the same age for the beginning of rifting activity in the Parentis Basin. Before this rifting stage related to the opening of Bay of Biscay regional data and quantitative

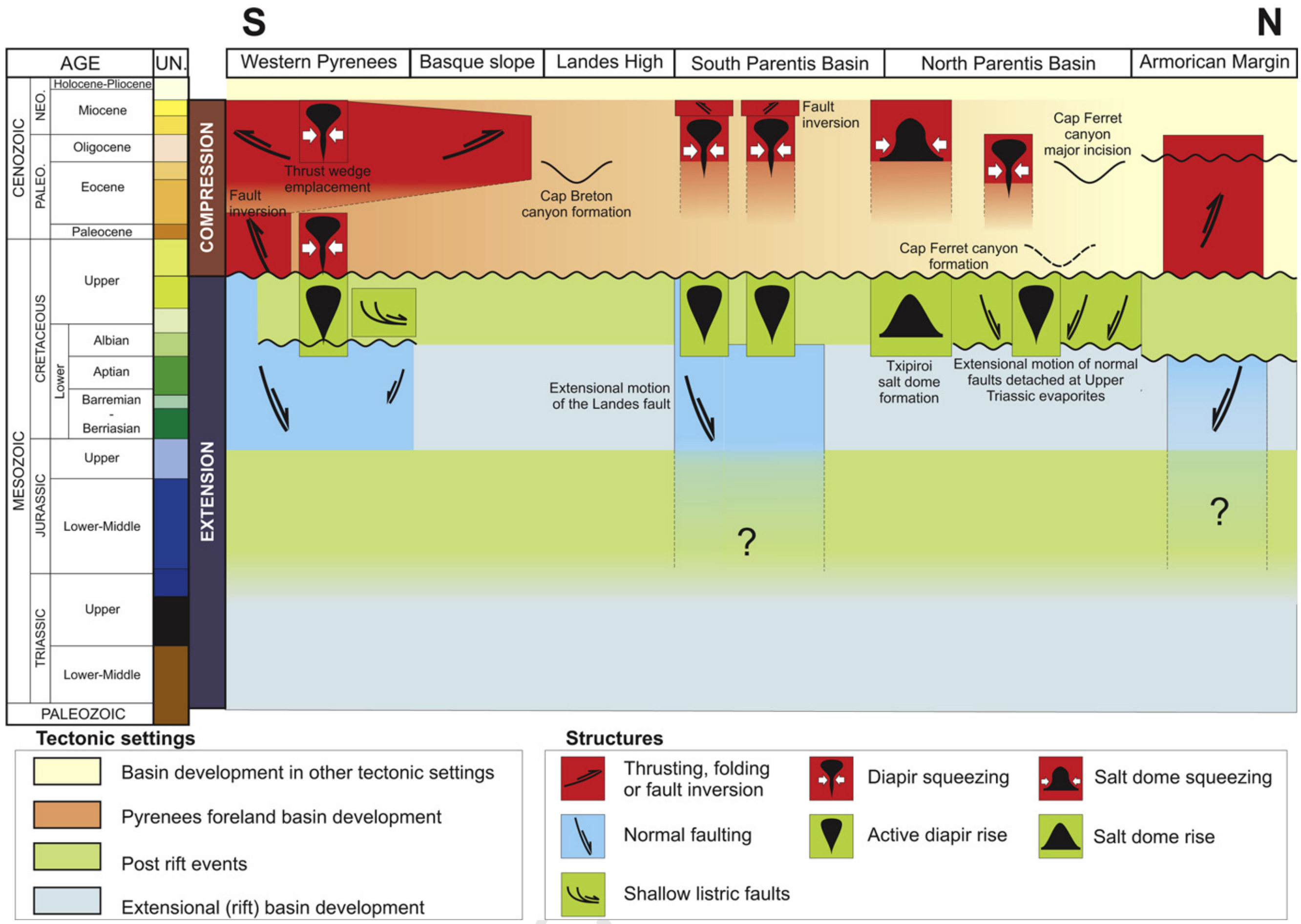

Fig. 10. Timing and main features of the Alpine deformations occurred along the MARCONI-3 and adjacent areas (Pyrenees and Armorican Margin). Pyrenees and Armorican Margin timing and main features of Alpine deformation are based on García-Mondejar (1996), Muñoz (2002), Thinon et al. (2002).

subsidence analyses indicate that the Parentis Basin was affected by two older rifting phases, namely during the Late Permian and during the latest Triassic–Early Liassic (Desegaulx and Brunet, 1990; Ziegler, 1990; Brunet, 1997, ~~1994~~). Q2

The second compressive deformational stage is coeval with the building of the Pyrenean orogen which led to the development of a foreland basin in the eastern Bay of Biscay at the southern portion of the MARCONI-3 profile (Fig. 3). This foreland basin developed from latest Cretaceous up to Middle Miocene as shown both by the (a) southward thickening of the uppermost Cretaceous–Middle Miocene successions; and (b) southwards dip of the same age horizons which progressively increases in depth. Although foreland basin development indicates that the building of the Pyrenean orogen began southwards during the latest Cretaceous, significant contractional deformation did not affect the area imaged by the MARCONI-3 profile up to Late Eocene. Before, older uppermost Cretaceous–Middle Eocene horizons only record a progressive flexure of the lithosphere towards the south and the formation of some minor reverse and normal faults in the outer foreland basin. The Upper Eocene–Middle Miocene horizon geometry, by contrast, depicts the development of major contractional structures affecting the foreland sediments and underlying rocks along the whole profile. From the geometric relationships between reflectors and these structures we can identify an evolutionary trend of the deformation in which two stages can be distinguished.

The first, Late Eocene–Early Miocene in age is characterized by the formation of a basement-involved thrust wedge in the innermost parts of the foreland basin and the squeezing of the diapiric structures formed previously in the Parentis Basin (diapirs and Txipiroi High saline dome). This squeezing generates the reactivation of the Txipiroi High saline dome and the expulsion, mainly upwards, of the viscous material intruded along the diapirs stems. This resulted in the second episode of active growth of the pre-existing diapirs with formation of secondary near-vertical welds that isolate diapir bulbs from their source layer. The formation of broad dynamic bulges above the diapirs generated the development of (1) narrow anticlines affecting the pre-Middle Miocene sediments; (2) growth strata with high onlap angles in the Upper Eocene–Lower Miocene deposits sedimented on its flanks; and (3) crestal normal faults cutting the same horizons.

The second, Middle Miocene in age, records the final stages of contractional deformation in the profile. It is a stage characterized by: the continuation of the Txipiroi High saline dome growth and the formation of a pop-up in the area comprised between the southern diapirs. This pop-up developed when contractional reactivation of these diapirs ceased (probably due to the closing of the diapir stems) and resulted from the inversion of the conjugate normal faults that generate these two diapirs.

6. Discussion Q3

6.1. Role of extensional mesozoic structure in the development of the Northern Deformational Front Of The Basque Pyrenees

The construction of an upper crustal transect joining the MARCONI-3 profile with the cross-section made by Pedreira (2004)

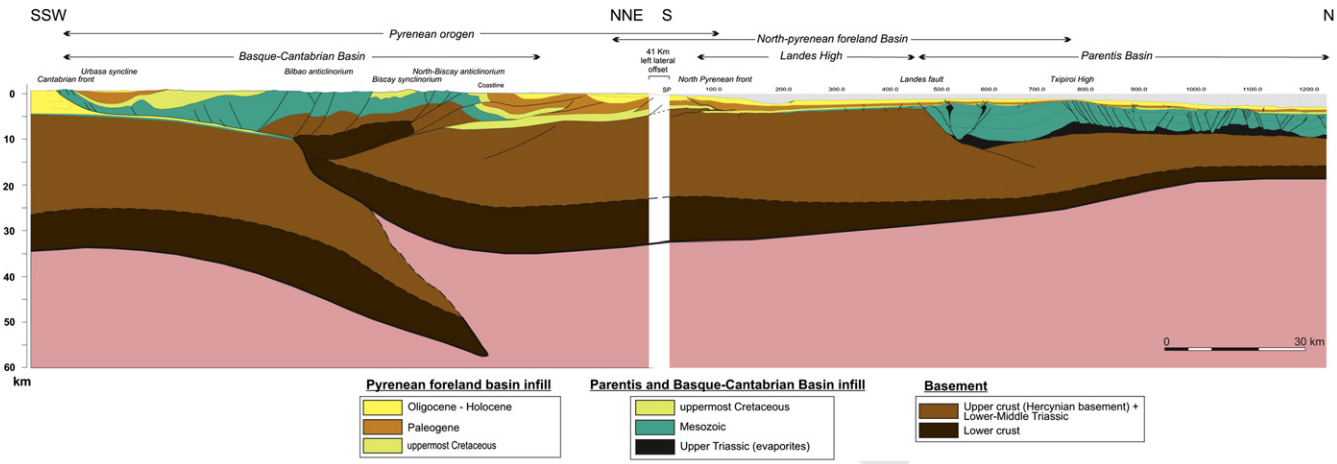

Fig. 11. Upper crustal transect through the eastern Bay of Biscay and adjoining northern part of the Basque Pyrenees based on MARCONI-3 profile interpretation and the cross-section made by Pedreira (2004) immediately southwards. See locations in Fig. 2.

immediately southwards (Fig. 11) confirms that the contractional structure of the Basque Pyrenees is strongly controlled by the previous Mesozoic extensional structure. The Basque Pyrenees results from the inversion of the Mesozoic Basque-Cantabrian Basin that led to north-directed thrusting of basement units and to the formation of thrust slices or inverted folds in the cover along its northern margin (Soler et al., 1981; Sanchez, 1991; Bois and Gariel, 1994; Cuevas et al., 1999). These structures resulted from the inversion of the south-dipping normal faults and the squeezing of the related salt diapirs which had also developed during the extensional period (Mathieu, 1986; García-Mondejar, 1987; Cuevas et al., 1999; Gómez et al., 2002). North of the basement thrusts developed close to these south-dipping normal faults in their footwall (basement short-cuts), Pyrenean contractional deformation diminishes drastically. It is only represented in the Landes High by some minor reverse faults, probably reactivated older normal faults, and more to the north in the Parentis Basin by gentle folds resulting, first from the squeezing of older diapir stems and then from the limited inversion of the normal faults bounding this basin to the south.

The lack of significant inversion structures in the Parentis Basin, despite that it belongs to a severely thinned crustal area before the Alpine compression, denotes that the Mesozoic Landes High acted as an important buffer for the propagation of the Pyrenean contractional deformation to the north. This deformational buffer was active from Late Cretaceous and especially during the period comprised between the Eocene and Early Miocene. During this time the role of the Landes High as a buffer is evidenced by the fact that the basement-involving Pyrenean contractional deformation was only present south of the Landes High and mainly concentrated along the northern margin of the Basque-Cantabrian Basin. North of this margin, in the Landes High and Parentis Basin, coeval contractional deformation was clearly minor and apparently restricted to the Mesozoic cover where the pre-existent diapirs were squeezed. The similar crustal signature of Basque-Cantabrian and Parentis Basins denotes that compressive stresses diminished abruptly northwards of the southern boundary of the Landes High which can be considered a buffer. From Early Miocene, during the last stages of orogen development, it seems that the Landes High leaved to act as a buffer since it was a build-up of intraplate compressive stresses northwards that produced the reactivation of some basement faults in the Parentis Basin.

The origin of the Landes High deformational buffer is probably related to the different crustal signature that showed this domain in relation to the adjoining Basque-Cantabrian and Parentis basins at the beginning of the compression. Strongly affected by the Early Cretaceous extension in the two basins the crust was thin, warm and mainly integrated by Mesozoic sediments. By contrast in the Landes High, where the Mesozoic extension was much lesser, the crust was thicker, colder and mainly made up of basement rocks. This resulted in a crust which was significantly stronger in the Landes High, and consequently, a differentiated crust which would be able to dissipate part of the compressive stresses generated southwards along the Iberian-Eurasia collision boundary.

This major strength of the Landes High in respect to the areas located immediately southwards (Basque-Cantabrian Basin) is expected that would disappear progressively with the growth of the Pyrenean orogen which would produce the thickening and cooling of the crust in these last areas and therefore an increase of their strength. This would explain why at the end of the Pyrenean building (Middle Miocene) the Landes High leaved to act as a buffer allowing the build-up of intraplate compressive stresses in the Parentis Basin area.

7. Conclusions

The new MARCONI-3 deep seismic profile allows recognizing the present upper crustal structure of the eastern part of the Bay of Biscay and the main features of its Alpine geodynamic evolution. Thus, geometry analysis of the seismic signature of the profile combined with available oil-well and refraction data denotes that the structure of the easternmost part of the Bay of Biscay consists of two distinct parts: a major Mesozoic basin to the north, the Parentis Basin, and a coeval structural high to the south, the Landes High.

The Parentis Basin, striking E–W, appears bounded southwards by a major north-dipping planar fault with a normal displacement of near 15 km. It is filled up by a thick (up to 10 km) sequence of Jurassic–Upper Cretaceous carbonates affected by salt domes and squeezed diapirs made up by Triassic evaporites and mudstones. These salt tectonic structures also affect the overlying uppermost Cretaceous–Lower Miocene synorogenic deposits which are folded upon these structures. Close to the southern basin boundary the profile also shows the development of a pop-up affecting the pre-Upper Miocene deposits which resulted from the reactivation of the major faults bounding the basin to the south.

The Landes High, located between the Parentis Basin and the Basque-Cantabrian Basin, belongs to a plateau uplifted and eroded during Early Cretaceous. It includes a thin pre-Upper Cretaceous cover tilted to the south unconformably overlaid by a thick wedge of uppermost Cretaceous to Cenozoic deposits filling the North-Pyrenean foreland basin. North of the Basque-Cantabrian Basin, in

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the Basque shelf, the Landes High and most of the overlying foreland basin fill are deformed by a thrust wedge developed during the Late Eocene–Miocene. This wedge constitutes the North-Pyrenean thrust front which developed along the northern margin of the Early Cretaceous Basque-Cantabrian Basin.

From geometric relationships and thickness variations it has been inferred that the overall structure of the eastern part of the Bay of Biscay results from the succession of two well-differentiated deformational stages: an initial one, coeval to the opening of the North Atlantic Ocean and the Bay of Biscay, which was extensional; and a younger compressive one, coeval by the building of the Pyrenean orogen.

The first stage includes a Late Jurassic (?)–Early Albian syn-rift stage in which the Parentis Basin formed from the normal motion of a major north-dipping fault located along its southern boundary and a Late Albian–early Late Cretaceous post-rift stage in which diapirs of Triassic evaporites grew close to this major fault.

The second compressive deformational stage led to the development of the latest Cretaceous up to Middle Miocene foreland basin in the southern portion of the MARCONI-3 profile. Although foreland basin development indicates that the building of the Pyrenean orogen began southwards during the latest Cretaceous, significant contractional deformation did not affect the present-day North-Pyrenean thrust front and the foreland until the Late Eocene. From the geometric relationships between dated reflectors and the contractional structures we can identify an evolutionary trend of the deformation in which two phases can be distinguished.

- The first, Late Eocene–Early Miocene in age, coincides with the formation of a basement-involving thrust wedge in the innermost parts of the foreland basin and the squeezing of the diapiric structures formed previously in the Parentis Basin.
- The second, Middle Miocene in age, is characterized by the continuation of the SP 750 saline dome growth and the formation of a pop-up close to the Landes Fault from the inversion of older conjugate normal faults.

This geodynamic evolution together with the structure of the area evidences that the extensional structure resulting from the opening of the Bay of Biscay played an important role both in the location of the North-Pyrenean front and in the features of the Pyrenean contractional deformation that took place in the northern foreland. Specially, the lack of significant inversion structures in the Parentis Basin, despite that it belongs to a severely thinned crustal area before the Alpine compression, denotes that the Mesozoic Landes High acted as an important buffer for the propagation of the Pyrenean contractional deformation to the north. This deformational buffer was active up to Early Miocene and probably vanished afterwards during the last stages of Pyrenean orogen development when some basement faults reactivated in the Parentis Basin.

Uncited references

Q4 Bois et al., 1997b; Cannerot, 1989; Derégnaucourt, 1981; Malod and Mauffred, 1990; Montadert and Winnock, 1971; Olivet et al., 1984.

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