

Alpine Orogeny: Deformation and Structure in the Northern Iberian Margin (Pyrenees s.l.)

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Abstract

The Pyrenean Orogenic system formed by the contractional deformation of the North Iberian margin from late Santonian to Middle Miocene times as a result of the collision between the Iberian and European plates. The Iberian lithosphere subducted underneath the European one. The structural style of the Pyrenees has been mainly controlled by the reactivation of the segmented rift system that formed at Late Jurassic-Early Cretaceous times connecting the Atlantic with the Alpine Tethys along the Bay of Biscay and Pyrenean domains, as well as by the distribution of the Triassic salt. Preserved syntectonic sediments in the adjacent foreland basins and piggy-back basins combined with thermochronological data allow constraining the evolution of the orogen.

et al. 2015). The Pyrenean orogen resulted from the inversion of the rift system and related passive margin (Northern Iberian Margin) that developed during the Late Jurassic-Early Cretaceous all along the northern Iberian plate connecting the Atlantic with the Alpine Tethys realms (Stampfli and Hochard 2009; Tugend et al. 2014). Before this extensional event, Triassic rifting, Stephano-Permian late Variscan tectonics, involving strike slip and extensional deformation, and Variscan and pre-Variscan tectono-metamorphic events also affected the Pyrenean domain.

Structural style and related tectono-sedimentary evolution changes significantly along strike. These changes are mainly expressed by differences in: width, asymmetry of the double wedge, thrust kinematics, involvement of basement and topography, among others (Figs. 9.1 and 9.2). Differences in basement involvement and topography are so strong that different physiographic units formed receiving distinct names such as Cantabrian and Pyrenean Ranges. Moreover, lower relief areas in between the ranges were not initially considered parts of the orogen and have been geologically referred as basins, regardless they are part of the orogenic system. As a result, different names have been proposed to refer to the entire system or parts of it and there is a confusing nomenclature with a mixture of geographical and geological names (see Barnolas and Pujalte 2004 for a review). Herein, the term Pyrenean orogen (Pyrenees s.l.) will be used to refer to the entire system. Local names of parts of this orogen, such as the Cantabrian Mountains, will be preserved for historical reasons and avoid confusion.

The main factors controlling the Pyrenean structural style are the inversion of the inherited extensional structures and the distribution of the Triassic salt. Other factors, such as the weakness of the inherited Variscan crust and the lithospheric thermal state, have also contributed to the structural evolution (Jammes and Huisman 2012; Jammes et al. 2014; Clerc and Lagabriele 2014). The Triassic salt has resulted into decoupling of the Mesozoic cover succession and a thin-skinned style in the Pyrenean fold and thrust belt. Inversion of the Early Cretaceous extensional system and the

9.1 Introduction

The Pyrenees is the orogenic system that runs along the boundary between the Iberian and European plates (Fig. 9.1). The Pyrenean orogen can be followed beyond the present Pyrenees mountain range. It continues eastward in the Alps along Provence in southern France, and westward along the southern Bay of Biscay into the Atlantic (Fig. 9.1). The Pyrenean orogen formed as the result of the collision between the Iberian and European plates from Late Cretaceous to Miocene times (Roest and Srivastava 1991; Rosenbaum et al. 2002). It is an asymmetric doubly vergent orogenic wedge that formed above the subduction of the Iberian lithospheric mantle and lower crust under the European plate (Fig. 9.2) (Choukroune and ECORS Team 1989; Muñoz 1992; Pedreira 2003; Campanyà et al. 2012; Chevrot

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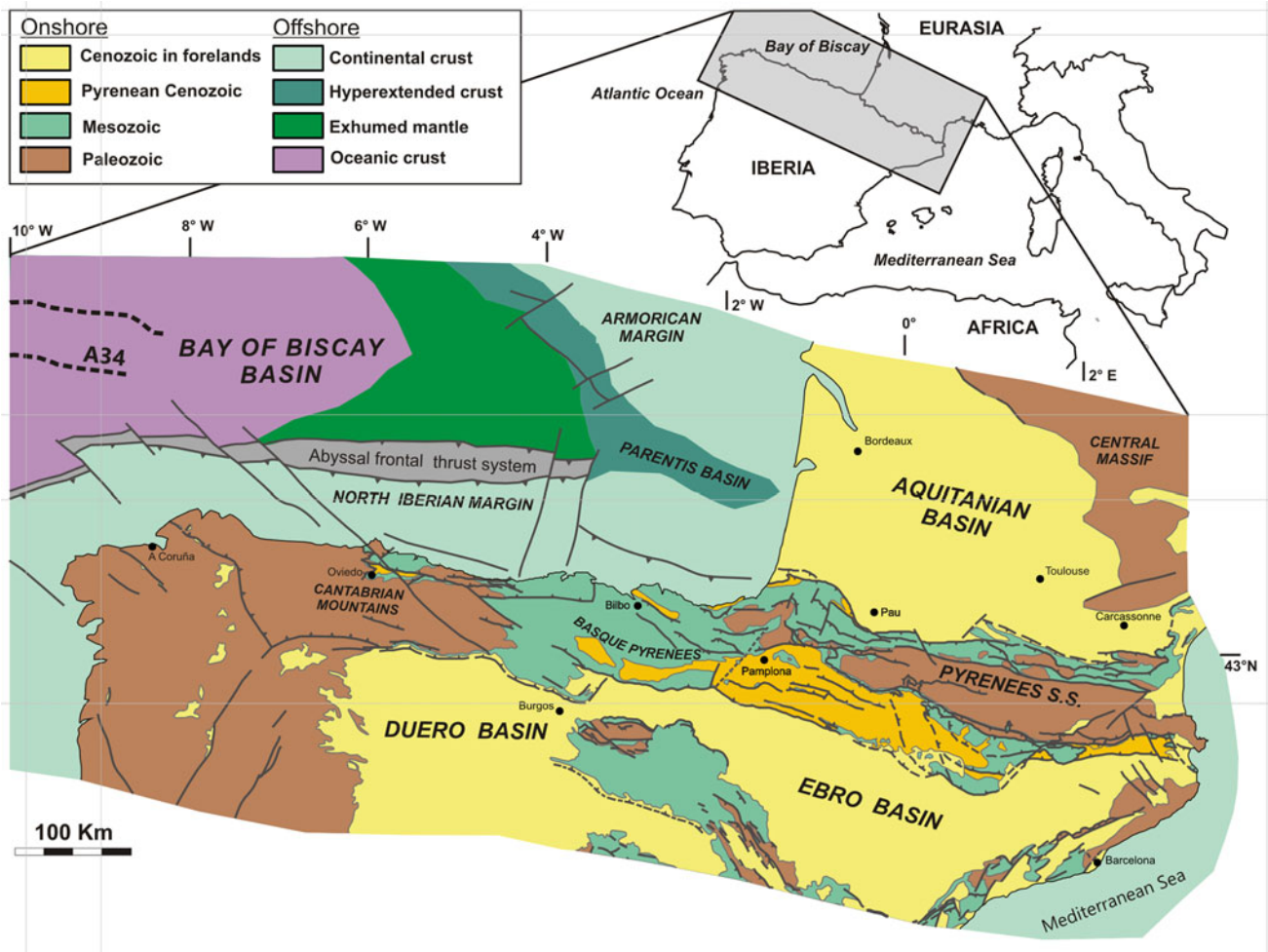


Fig. 9.1 Structural map of the Pyrenean orogen. Modified from Carola et al. (2013)

weaker parts of the Variscan crust have promoted basement involvement and thick-skinned structural style. Areas with absence of Triassic salt and a relatively weak basement, such as the Cantabrian Mountains, are characterized by a thick-skinned structural style (Alonso et al. 1996), while areas with a thick Triassic salt, such as the Basque-Cantabrian Pyrenees and the central Pyrenees are characterized by thin-skinned geometries (Carola et al. 2015; Muñoz 1992). In the thin-skinned areas, basement is involved in the retro-wedge and can be involved as well in the pro-wedge underneath the detached cover, depending on the inherited rift system (Fig. 9.2). Thus, the Pyrenean fold and thrust belt exhibits both thin- and thick-tectonic style in different segments of the orogenic system as many other fold and thrust belts in the planet (Lacombe and Bellahsen 2016) and cannot be assigned to a single structural style as recently proposed (Mouthereau et al. 2013).

Constraints on the temporal and spatial evolution of the Pyrenees are provided by the preservation of syntectonic sediments since the early stages of the convergence.

Shortening estimates based on construction of crustal scale balanced cross-sections are reasonable provided the moderate amount of deformation and the absence of ductile deformation and related metamorphic processes during the convergence.

9.2 Geodynamic Setting

Structures and sedimentary record of the Mesozoic, Paleogene and Lower Neogene successions in the Pyrenees are related with the kinematics of the Iberian plate that in its turn is the result of the successive stages of the opening of the Atlantic Ocean. There is a strong debate about the evolution of Iberia during the Mesozoic and there are several published plate models (Olivet 1996; Sibuet et al. 2004; Vissers and Meijer 2012, among others). See Barnett-Moore et al. (2016) for a recent compilation and discussion. This current unresolved debate is an indication of the ambiguities associated with defining clear oceanic crust along the West Iberia and

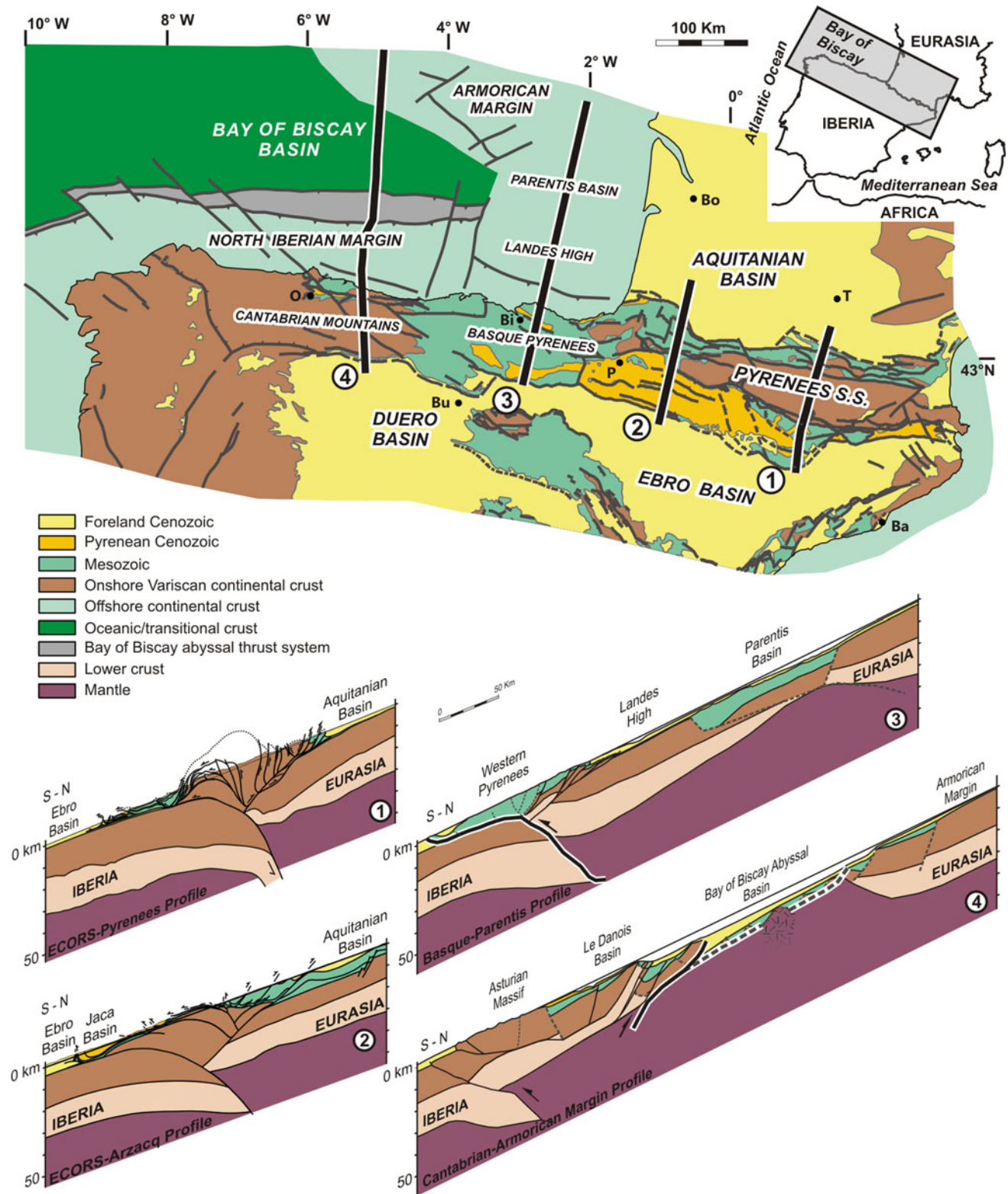


Fig. 9.2 Seriated crustal cross-sections of the Pyrenees depicting the main structural changes along the orogen. 1: ECORS–Pyrenees (from Muñoz 2002); 2: ECORS–Arzacq (modified from Teixell 1998); 3: Basque–Parentis (from Pedreira 2004 and Ferrer et al. 2008); 4: Cantabrian–Armorican Margin (modified from Pulgar et al. 1996 and Álvarez-Marrón et al. 1997)

Newfoundland margins as well as knowing the nature and origin of the magnetic anomalies used for the reconstruction of the Iberian plate motion (Bronner et al. 2011; Nirrengarten et al. 2016).

Opening of the Central Atlantic progressed from south to north. South of the Newfoundland-Azores-Gibraltar Fracture Zone (NAGFZ) oceanic crust formed at Middle Jurassic times connecting the Atlantic with the Ligurian-Tethys along the Gulf of Cádiz and Betics (Stampfli and Hochard 2009). North of the NAGFZ, plate separation between Iberia and Newfoundland and crustal thinning to exhumate the mantle occurred at latest Jurassic to Early Cretaceous times. Magnetic anomalies older than M3 (ca. 127 Ma) that were considered before as recording oceanic crust (M3 to M20, 127-147 Ma) have been reinterpreted as exhumed continental mantle (Sibuet et al. 2007). Oceanic crust formed during Aptian times (between 125 Ma and 112 Ma) west of Iberia from south to north (Sibuet et al. 2007; Bronner et al. 2011; Tucholke and Sibuet 2012; Nirrengarten et al. 2016) and, accordingly, it progressed into the western part of the Bay of Biscay most probably during Albian times. Anticlockwise rotation of the Iberian plate has been related with the opening of the Bay of Biscay and dated as Aptian in age from paleomagnetic data in Cretaceous Pyrenean basins (Dinarès-Turell and García-Senz 2000; Gong et al. 2008). Thus, this vertical axis rotation could be related with mantle exhumation during hyperextension between Iberia and Europe more than oceanic accretion. Nevertheless, the reliability of the available paleomagnetic data has been questioned (Neres et al. 2013).

Following reconstruction of different magnetic anomalies, such as the J anomaly, two end-member plate models have been proposed: (i) reconstructions with an Early Cretaceous ocean all along the Pyrenean domain and a separation between Iberia and Europe of up to 600 km (Srivastava et al. 2000; Sibuet et al. 2004; Vissers and Meijer 2012); (ii) reconstructions with a separation between Iberia and Europe of no more of 200 km along a N-S direction and with oceanic crust restricted to the western Bay of Biscay (Klitgord and Schouten 1986; Olivet 1996; Jammes et al. 2009, among others). The other difference between the proposed plate models for Iberia concerns about the magnitude and timing for the eastward movement of Iberia with respect to Europe. All the proposed models agree that Iberia experienced a left-lateral strike-slip motion along the Northern Iberian Margin, as the Northern Atlantic oceanic crust propagated northwards, although, the proposed amount of left-lateral strike-slip varies from 100 km to several hundreds. However, there is not agreement about the timing. In the seventies and until recently, it was believed that the left-lateral displacement of Iberia along the North Pyrenean Fault mainly occurred during the Albian and Cenomanian and subsequently this fault acted as the main plate boundary

during the Pyrenean orogeny (Le Pichon and Sibuet 1971; Choukroune and Mattauer 1978; Olivet 1996). In that case strike-slip motion would have been synchronous (transtensional motion) to younger than the main extensional event in the Northern Iberian Margin and the anticlockwise rotation of Iberia. On the contrary, recent reconstructions propose that the transtensional motion occurred until the Aptian and then it was followed by orthogonal extension between Iberia and Eurasia (Jammes et al. 2009).

The existence of an Early Cretaceous oceanic crust all along the Pyrenean domain, subsequently subducted during the Aptian and Albian (Vissers and Meijer 2012; Vissers et al. 2016) contradicts with all the Pyrenean geological data and recently acquired high-resolution tomography models (Chevrot et al. 2014, 2015; Barnett-Moore et al. 2016).

Discrepancies in the plate models are mainly related with the Jurassic-Cretaceous evolution, but it is widely accepted that by late Santonian times (ca. 83 Ma) a major shift in plate boundary forces between Africa, Iberia and Eurasia triggered a compressive regime in the Northern Iberian Margin, lasting until the Miocene and forming the present-day Pyrenees (Rosenbaum et al. 2002).

9.3 Geophysical Data and Lithospheric Structure

The crustal and lithospheric structure of the Pyrenees has been constrained by different geophysical techniques (deep reflection and refraction seismic profiles, gravity, magnetotellurics, magnetic anomalies, tomography, heat flow). There is not any other orogen with such an amount and quality of geophysical data as it is available in the Pyrenees. Moreover, recent progresses in deep imaging, such as new seismic tomographic techniques and magnetotelluric modelling have considerably improved our understanding of the lithospheric structure of the Pyrenees (Campanyà et al. 2012; Chevrot et al. 2015; Wang et al. 2016).

The available geophysical data provide clear and definite evidence for the subduction of the Iberian continental lithosphere underneath the European plate (Fig. 9.2). This was already shown by the ECORS profile in the Central Pyrenees (Choukroune and ECORS Team 1989; Muñoz 1992) and has been confirmed also to exist in the Basque-Cantabrian Pyrenees by the ESCIN-2 reflection profile (Pulgar et al. 1996), the wide-angle refraction surveys (Fernández-Viejo et al. 2000) and the seismic data acquired during the MARCONI experiment in the Bay of Biscay (Roca et al. 2011; Ruiz et al. 2017). In the easternmost Pyrenees the subducted Iberian crust is not observed (Gallart et al. 2001; Campanyà et al. 2017), as the crustal thinning related with the opening of the Western Mediterranean has been superimposed on the previously thickened lithosphere.

Westward, in the western Cantabrian Mountains an abrupt transition from the Pyrenean thickened crust to the Variscan crust with a 30–32 km thickness of the Iberian Massif is observed (Pérez-Estaún et al. 1994; Fernández-Viejo et al. 2000). This transition trends NW-SE and coincides with oblique faults at surface, such as the Ventaniella Fault (Muñoz 2002). The subducted Iberian crust has reached a depth of up to 60–70 km in most of the studied Pyrenean transects. The subducted Iberian slab mostly corresponds with the lower part of the crust with a thickness varying between 15 km in the western Pyrenees to 25 km in the central Pyrenees (Fig. 9.2, Muñoz 1992; Gallastegui et al. 2002; Chevrot et al. 2015; Pedreira et al. 2015; Wang et al. 2016). A deeper depth for the subducted Iberian lower crust as suggested by crustal balanced cross-sections (Beaumont et al. 2000) would be also compatible with geophysical data (Pedreira et al. 2015). The subducted lower crust could have been experienced either partial melting to account for the conductivity anomalies (Glover et al. 2000) or an eclogitic metamorphism with water release to better fit with the proposed 3D density models of the Pyrenean deep structure (Vacher and Souriau 2001).

In recent years there is growing evidence that the Bay of Biscay-Pyrenean domains experienced hyperextension during the Early Cretaceous rifting that occurred between the Iberian and European plates (Jammes et al. 2009; Lagabrielle et al. 2010; Roca et al. 2011; Clerc et al. 2012; Masini et al. 2014; Tugend et al. 2014). In this rift system the oceanic domain was restricted to the western Bay of Biscay, westward of 6–8 W (Fig. 9.1) as evidenced by the ocean floor magnetic anomalies recorded symmetrically at both sides of the spreading centre (Srivastava et al. 1990; Sibuet et al. 2004). There the youngest identified anomaly A33o dates the end of seafloor spreading at ~80 Ma, which coincides with the onset of contractional deformation in the Pyrenees. East of this spreading centre, the abyssal plain of the Bay of Biscay is floored by a basement characterized by a high-velocity lower crust (~7.20–7.30 km s⁻¹) on top of a low-velocity upper mantle (~7.7–7.9 km s⁻¹) (Fernández-Viejo et al. 1998; Gallart et al. 1997; Ruiz et al. 2017). These anomalous velocities can be explained either by upper mantle hydration/serpentinization (Roca et al. 2011; Pedreira et al. 2015) and/or by transitional continental crust, highly stretched and thinned, lying on top of serpentinized mantle and intruded by gabbroic igneous rocks (Ruiz et al. 2017). Exhumed mantle would have also rimmed the oceanic crust in the western Bay of Biscay although it would have been subducted in the Cantabrian margin (Tugend et al. 2014). Eastwards, exhumed mantle along the Pyrenean rift axis continued as far as the eastern Pyrenees (Clerc and Lagabrielle 2014).

Apart from the geophysical characteristics of the Bay of Biscay lithosphere and the outcropping bodies of lherzolites

as well as sediments sourced by exhumed mantle along the Pyrenean chain, other evidences of mantle exhumation come from the presence of strong positive Bouguer anomalies along the core of the chain (Bilbao, Mauléon and Sant Gaudens anomalies, Casas et al. 1997). The strong positive Bouguer anomaly of the Mauléon basin has been recently interpreted as a continuous mantle wedge lying on top of the subducted Iberian plate reaching the bottom of the basin during the pre-compressional hyperextension (Wang et al. 2016). This mantle wedge would explain why the Pyrenees appear isostatically over-compensated, and why the deep crustal roots are shifted 50 km northward with respect to the topographic highs (Wang et al. 2016).

Several different interpretations of the crustal structure of the Pyrenees have been given on the basis of the combined geological and geophysical data (Roure et al. 1989; Muñoz 1992; Teixell 1998; Pedreira et al. 2015; Wang et al. 2016). Recent geophysical data and modelling rule out the interpretation of a Pyrenean crustal structure driven by the North-Pyrenean Fault (Choukroune and Mattauer 1978; Mattauer 1990). The crustal structure that best integrates all the available data in the Pyrenees ss. and the Basque Pyrenees is that of an asymmetric orogenic double-wedge involving only upper crustal rocks of the Northern Iberian margin south of the hyperextended rift domain (southern wedge). The extensional detachment system that thinned out the crust and exhumed the mantle would have been reactivated and the transition to the narrow northern wedge is located in its hangingwall (Roca et al. 2011). Apparently, the Iberian crust was decoupled and the lower crust, below the upper crustal double-wedge, was subducted (together with the lithospheric mantle) into the mantle (Beaumont et al. 2000).

Thus, the Pyrenean lithospheric structure is strongly controlled by the inversion of the previous segmented hyperextended rift system (Roca et al. 2011; Tugend et al. 2014). For a given realm of the Pyrenean system, previous rift domains controlled the localization of the deformation, the geometry and structural style of the orogenic double-wedge (Tugend et al. 2014), whereas along strike changes have been determined by the transfer zones that segmented the rift system (Roca et al. 2011).

A debate still exists about the crustal structure of the westernmost part of the Pyrenean orogenic system, the realm facing the oceanic crust of the Bay of Biscay (Fig. 9.1). Boillot and Capdevila (1977) proposed the subduction of the oceanic crust southwards underneath the Iberian plate at the Cantabrian margin, as also suggested by the asymmetry of the magnetic anomalies recorded in the western Bay of Biscay (Sibuet et al. 2004). The hypothesis of the southward subduction of the Bay of Biscay oceanic crust would imply a change in the polarity of subduction with respect to the central Pyrenees that was interpreted to occur across a flip

zone coinciding with the Pamplona transfer zone (Engeser and Shwentke 1986). Nevertheless, several seismic profiles of the ESCIN and MARCONI programmes and related projects (wide-angle reflection seismic data and wide-angle phases recorded by the OBS and land seismic stations deployed at the MARCONI N-S profiles) showed that the Iberian continental crust is subducting to the north along the Cantabrian margin. Such configuration has been interpreted as a result of the southward displacement and underthrusting of the Bay of Biscay lower crust, indenting into the Iberian crust (Pulgar et al. 1996; Fernández-Viejo et al. 1998, 2000; Gallastegui et al. 2002; Pedreira et al. 2015). These authors emphasize that there are not any geophysical evidences for an oceanic slab below the Cantabrian margin and consider that the Bay of Biscay oceanic crust has not been subducted southwards. Instead, it has been underthrust underneath the Cantabrian accretionary prism (Ruiz et al. 2017). Regardless the terminology used, the oceanic crust of the western Bay of Biscay together with the exhumed mantle or the extremely thinned continental crust that rimmed the oceanic one have been subducted/underthrust underneath the former distal Cantabrian rift domain (Fig. 9.1, Tugend et al. 2014). A change of the polarity of the extensional system that exhumed the mantle across the Santander transfer zone would explain the differences of the crustal structure once the Pyrenean rift system was inverted (Roca et al. 2011). There was not any change in the polarity either of the extensional system or the subduction of the Iberian crust across the Pamplona transfer zone as previously suggested.

Factors controlling the deformation style of orogens involving crustal subduction, and particularly for the case of the Pyrenees, have been analysed by numerical modelling (Beaumont and Quinlan 1994; Beaumont et al. 2000; Jammes and Huisman 2012; Jammes et al. 2014). These authors concluded that simple models involving lower crustal subduction together with the mantle best explain the observed features. The initial simple models have refined to include the geodynamic evolution as deduced from the geological record (Beaumont et al. 2000) as well as the inversion of the rift system (Jammes and Huisman 2012) and the inherited Variscan crustal features (Jammes et al. 2014). The lateral variation in crustal strength attributed to inherited Variscan crustal composition accentuated during Mesozoic rifting explains the variation in structural style observed during Pyrenean mountain building (Jammes et al. 2014). Accordion models illustrate the localization of the contractional deformation in the centre of the rift system where the mantle was exhumed, the subsequent reactivation of the extensional system and the propagation of the deformation into the proximal rift domains or areas non previously stretched when the European crust was indented into the delaminated Iberian crust (Jammes and Huisman 2012; Jammes et al. 2014).

9.4 Main Structural Features

Understanding the main structural features of the Pyrenees requires not only the knowledge of the structural evolution along well-constrained cross-sections but also knowing the along strike changes and the factors controlling them.

9.4.1 The ECORS Cross-Section of the Central Pyrenees

The most constrained section from geological and geophysical data is the central Pyrenees cross-section along the ECORS seismic profile (Choukroune et al. 1989; Muñoz 1992; Fitzgerald et al. 1999; Beaumont et al. 2000; Campanyà et al. 2012; Mouthereau et al. 2014; Chevrot et al. 2015).

The ECORS cross-section shows the geometry of the Pyrenean asymmetric orogenic double-wedge dominated by a central antiformal stack of basement-involved thrust sheets, classically referred as the Axial Zone (Figs. 9.3, 9.4 and 9.5, Muñoz 1992, 2002). These thrust sheets have been transported to the south and deformed progressively by underthrusting of the lower and younger units. They are mainly three thrust sheets, which from top to bottom are: Noguères, Orri and Rialp (Fig. 9.5, Muñoz 1992). The northern limb of the antiformal stack occupies most the Axial Zone and is characterized by a moderately-dipping to steeply-dipping Variscan main foliation which was deformed by back-tilting. The southern limb is relatively narrow and is characterized by downward facing folds involving upper Paleozoic rocks and Permian to Lower Triassic red beds of the Noguères Zone (Fig. 9.4, Dalloni 1913; Séguret 1972).

The basement has been decoupled from the cover South Pyrenean thrust sheets along the Triassic evaporites. These thrust sheets are from north to south: Bóixols, Montsec and Serres Marginals (Figs. 9.3, 9.4 and 9.5). They have involved different parts of Mesozoic and Paleogene basins with distinct significances and they have initially developed following a forward propagating thrust sequence from Late Cretaceous to Oligocene.

The Bóixols thrust sheet resulted from the inversion of the Lower Cretaceous basins at the southern margin of the Pyrenean rift system (Berastegui et al. 1990). Contractional deformation started at late Santonian and continued during the Late Cretaceous (MacClay et al. 2004; López-Mir et al. 2014).

The Montsec thrust sheet involved the northern part of the Upper Cretaceous foreland basin characterised by a strong subsident turbiditic trough at the footwall of the Bóixols thrust. These turbidites grade southward into a carbonatic platform that constitutes the backbone of the Montsec Range. The Montsec thrust sheet developed from the

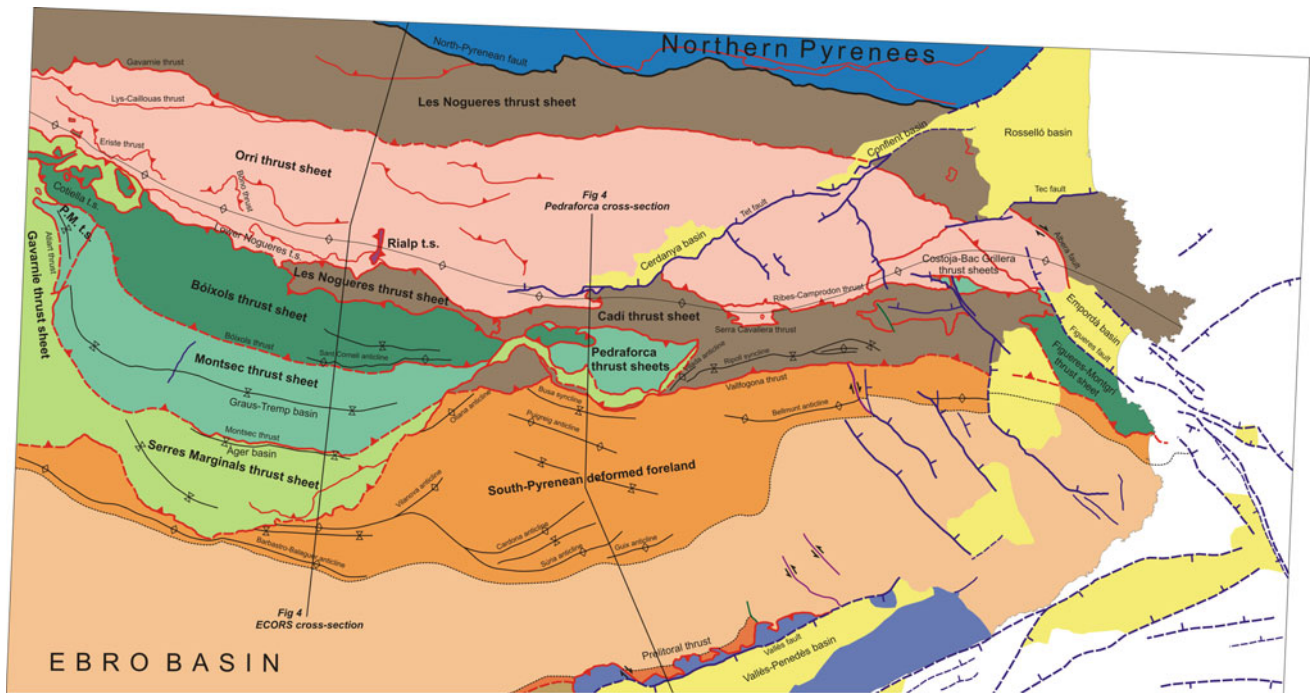


Fig. 9.3 Structural map of the central and eastern Pyrenees showing the main structural units. See location in Fig. 9.1

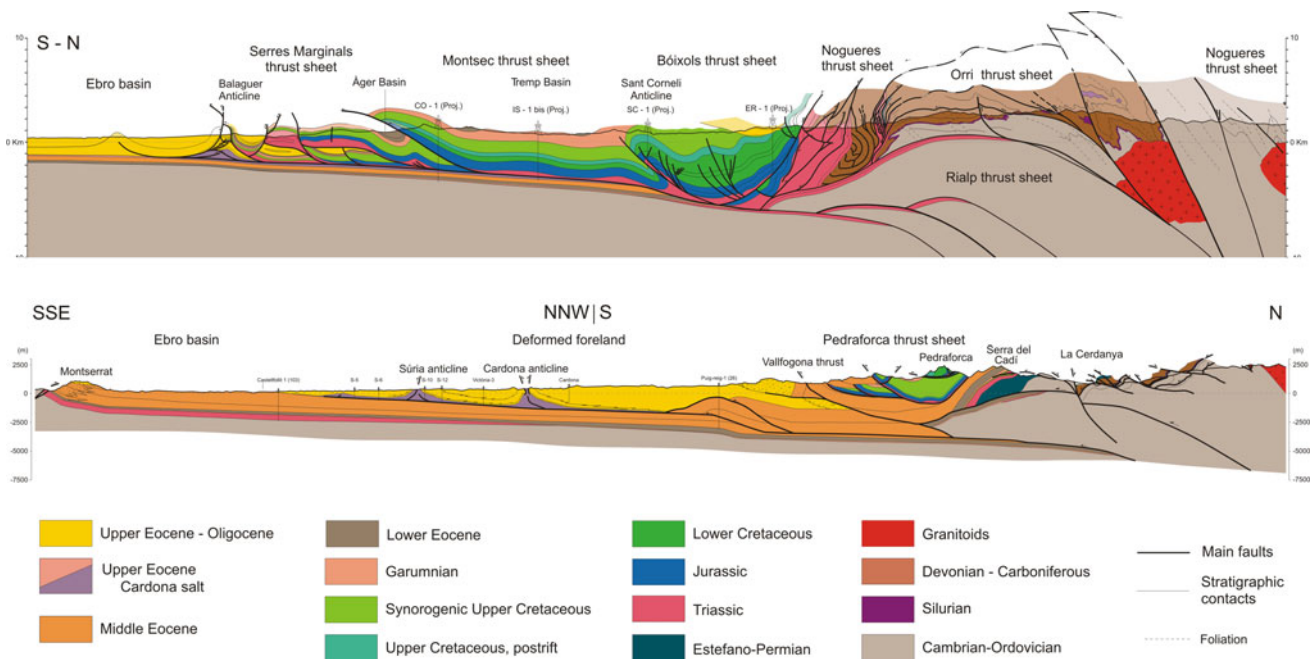


Fig. 9.4 Cross-sections across the central and eastern Pyrenees. The upper cross-section follows the ECORS seismic profile and has been modified from Berástegui et al. (1993). The lower section across the

Pedraforca thrust sheets of the eastern Pyrenees has been modified from Vergés (1993) and Casas et al. (2015). See location in Fig. 9.3

Paleocene to the Early Eocene as recorded by continental to shallow marine sediments deposited in its footwall (Ager basin) as well as in the Tremp-Graus piggy-back basin (Fig. 9.4).

The Serres Marginals thrust sheet is characterised by an incomplete and thin Mesozoic-Paleocene succession, which progressively reduces southwards. The synorogenic Upper Cretaceous carbonates are represented in all the Serres

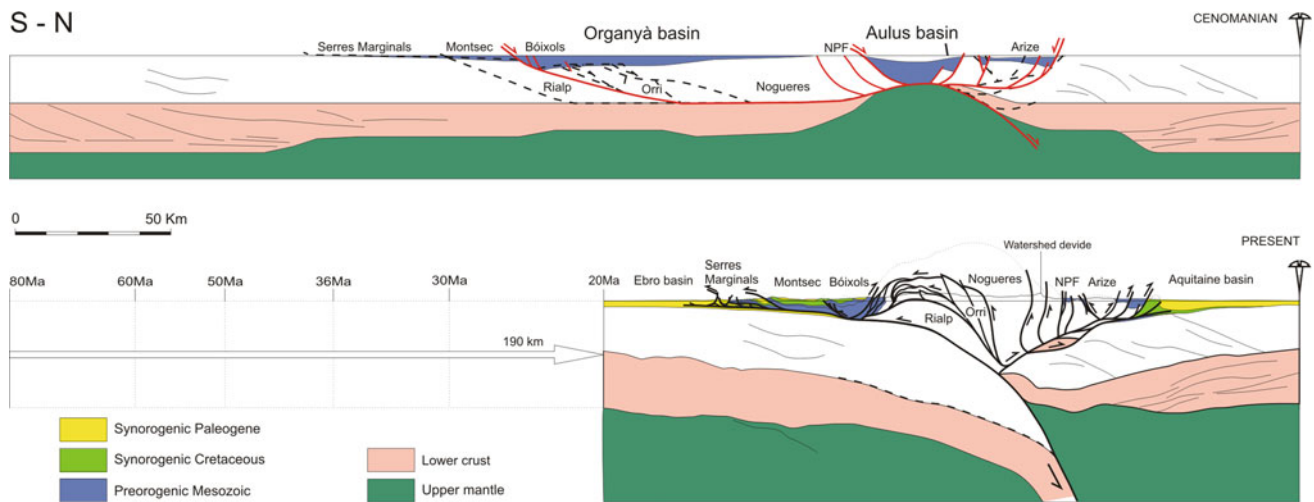


Fig. 9.5 ECORS balanced and restored cross-sections. The restored cross-section has been modified from Beaumont et al. (2000) and Muñoz (2002) to include the exhumation of the lithospheric mantle in

the central part of the Pyrenean rift system at the end of the Early Cretaceous rifting. See location in Fig. 9.3

Marginals imbricates with a thickness varying from few hundred meters in the northern ones to few tens of meters in the most frontal imbricates (Fig. 9.4). These Upper Cretaceous sediments unconformably overlie the Jurassic carbonates that were tilted northwards and removed by erosion at the forebulge of the Upper Cretaceous foreland basin. The southern edge of the Triassic evaporites preserved by erosion at the northern limb of the forebulge controlled the location of the thrust front. Triassic evaporites also controlled the internal structure of the Serres Marginals thrust sheets. These evaporites were inflated by evacuation from the bottom of the Montsec thrust sheet and together with the thin Mesozoic succession favoured the development of detachment anticlines and diapirs (Santolaria 2015; Santolaria et al. 2015).

The Serres Marginals and Montsec thrust sheets are at present above and autochthonous succession of Paleocene and Eocene sediments in continuation with the Ebro foreland basin as evidenced by well and seismic data (Fig. 9.4). The detachment is located in the Upper Eocene salts giving a minimum displacement for the South Pyrenean thrust sheets during the Late Eocene and Oligocene of more than 40 km. The northern cut-off of these autochthonous sediments, in the footwall of the sole thrust, is not constrained, but for the Upper Eocene sediments should be located close to the northern edge of the Montsec thrust sheet with an approximate E-W trend as clearly evidenced by seismic data tight to the Comiols and Isona wells, and contrary to recent interpretations which tend to minimize thrust displacement (Mouthereau et al. 2014). The South-Pyrenean thrust sheets were reactivated by a break back thrust sequence during the sedimentation of the Upper Eocene-Oligocene conglomerates once the Ebro basin became endorheic and filled by a thick succession of syntectonic sediments (Vergés and

Muñoz 1990; Burbank et al. 1992a; Coney et al. 1996; Meigs and Burbank 1997; Muñoz et al. 1997; Fillon et al. 2013). The break-back reactivation of the thrust system previously developed by a piggy-back sequence was coeval with both the displacement of all the cover thrust sheets above the Upper Eocene salt and the significant increase of the uplift rate and denudation of the basement thrust sheets of the Axial Zone (Fitzgerald et al. 1999; Beamud et al. 2011).

The Northern Pyrenees (North Pyrenean Zone, Choukroune 1976) are significantly narrower (25–35 km) than the Southern Pyrenees and constitute the retrowedge of the Pyrenean double-wedge (Fig. 9.5). The North Pyrenean thrust system inverted the Lower Cretaceous basins that developed in the hyperextended rift domain in the hangingwall (northwards) of the sole extensional detachment, which at present corresponds to the North Pyrenean Fault (Fig. 9.5, Mouthereau et al. 2014). During the Early Cretaceous continental stretching crustal boudins surrounding the exhumed mantle rocks and the Lower Cretaceous basins developed in a hot-regime (Clerc and Lagabrielle 2014). These crustal blocks form at present the North-Pyrenean massifs. Synrift Lower Cretaceous deep-water sediments and prerift Jurassic carbonates were affected by HT-LP metamorphism between 110 Ma (Albian) and 85 Ma (Santonian). Metamorphism occurred synchronously and afterwards the mantle exhumation but spatially and temporally related to the crustal attenuation predating the contractional deformation (Albarède and Michard-Vitrac 1978; Golberg et al. 1986; Montigny et al. 1986). The North Pyrenean frontal thrust resulted from the inversion of the extensional faults at the northern margin of the rift system (Baby et al. 1988; Déramond et al. 1990). In its footwall the

retroforeland basin is filled up by a thick (5 km) mainly clastic succession consisting in two main subsiding shallowing-up cycles: a first deep-water late Santonian-Paleocene succession and a second shallow marine to continental Paleocene-Oligocene one (Ford et al. 2016).

9.4.2 The Eastern Pyrenees

The main structural units above explained for the Central Pyrenees continue along strike into the eastern Pyrenees, east of the Segre valley (Fig. 9.3). However, there are significant structural differences between the two realms of the orogen, mainly in the southern Pyrenees. The most obvious one is the thrust re-entrant geometry depicted by the main thrusts that connect with the central Pyrenees along an oblique thrust system (Fig. 9.3). The Bóixols, Montsec and Serres Marginals thrust sheets are piled one on top of the other and constitute the Pedraforca thrust sheets (Figs. 9.3 and 9.4, Solé-Sugrañes 1971; Séguret 1972; Vergés 1993; Vergés et al. 1995). As a result, the Eastern Pyrenees shows the narrowest areal extension of the entire system. The Pedraforca thrust sheets lie on top of the Eocene syntectonic sediments of the Cadí thrust sheet and altogether form a synform in the hangingwall of the Vallfogona frontal thrust, southwards of the Axial Zone antiformal stack (Fig. 9.4, Muñoz et al. 1986; Vergés 1993). The Cadí thrust sheet involves Upper Paleozoic rocks underneath a thick succession, up to 6 km thick, of Paleocene and Eocene sediments. Mesozoic rocks are only restricted to a thin succession of Triassic red beds and carbonates and Upper Cretaceous carbonates and sandstones only represented in the western part of the Cadí thrust sheet. The Cadí thrust sheet is the structural equivalent of the Noguères thrust sheet in the Central Pyrenees (Muñoz et al. 1986). It supports the Ripoll piggy-back basin characterized by a Lower Eocene deep-water sedimentary trough that was deposited synchronously to the emplacement of the lower Pedraforca thrust sheets and a Middle Eocene-Oligocene shallow water to continental succession coeval with the emergence of the Vallfogona frontal thrust and the underthrusting of basement-involved thrust sheets (Figs. 9.3 and 9.4, Muñoz et al. 1986; Puigdefàbregas et al. 1986; Burbank et al. 1992b). Above the turbiditic trough a salt unit was deposited (Beuda Fm). It has acted as a main detachment in the eastern Pyrenees and controlled the late stages of the structural evolution.

There is a regional westward plunge of the Pyrenean structures. As a result, the deepest structural levels are cropping out in the Eastern Pyrenees and allow us to complete the structural picture observed in the Central Pyrenees. The deep structural exposures show the antiformal stack geometry of the basement-involved thrust sheets, as revealed

by vertically stacked tectonic windows (Muñoz et al. 1986). The structural culmination of the southern Pyrenees is located in the easternmost Pyrenees. From there, an eastward structural plunge towards the Mediterranean coast allows us to observe upper thrust sheets involving Mesozoic sediments detached on Triassic evaporites (Fig. 9.3, Pujadas et al. 1989; Tassone et al. 1994). The uppermost of these thrust sheets, the Figueres thrust sheet, involves a complete Mesozoic sequence, including synrift Lower Cretaceous carbonates (Fig. 9.3). It developed out-of-sequence on top of Middle Eocene continental sediments and the amount of thrust displacement as well as its restored position is still unknown (Pujadas et al. 1989).

Another distinct feature of the Eastern Pyrenees with respect the Central Pyrenees is the width of the deformed Ebro foreland basin (Fig. 9.3). South of the Vallfogona frontal thrust a wide area is characterized by different trends of detachment anticlines which have been detached above three distinct salt horizons, which from north to south are the Beuda, Cardona and Barbastro evaporitic units (Figs. 9.3 and 9.4, Vergés et al. 1992).

9.4.3 The Central Pyrenees and the Transition to the Western Pyrenees

The Bóixols, Montsec and Serres Marginals thrust sheets form the most prominent thrust salient in the Pyrenees (Figs. 9.1, 9.3 and 9.6). It was named South Pyrenean Central Unit and it was considered as a primary curvature inherited from the geometry of Mesozoic basins (Séguret 1972; Soto et al. 2002). However, it has been demonstrated by the integration of structural, stratigraphic and paleomagnetic data that this thrust salient developed by a progressive curvature with a divergent thrust transport direction from Middle Eocene to Oligocene during the formation of the Serres Marginals thrust sheets and its western continuation into the Gavarnie thrust sheet (Sussman et al. 2004; Muñoz et al. 2013). The amount of clockwise vertical rotation, up to 70° in the Ainsa basin and related N-S trending folds, and the areal extension of the rotated area resulted from a thrust displacement gradient of about 50 km that relates with the uneven distribution of the Upper Triassic salt (Muñoz et al. 2013). The amount of the displacement gradient is in agreement with the existence of an autochthonous Paleogene succession underneath the Serres Marginals and Montsec thrust sheets.

A major structural change does occur between the central and the west central southern Pyrenees (Fig. 9.2). In the west central Pyrenees the Upper Cretaceous carbonates unconformably overlie the Paleozoic rocks of the Axial Zone. The absence of Triassic evaporites has resulted in the coupling of the Mesozoic and Paleogene successions to the basement.

Basement thrust sheets are mostly imbricated (Cámara and Klimowitz 1985; Teixell 1996, 1998; Labaume et al. 2016) instead of piled one on top of the other as observed in the antiformal stack of the central Pyrenees (Muñoz 1992). The change in thrust geometry in turn leads to a more distributed structural relief in the west central Pyrenees (Fig. 9.2). As basement thrust sheets are located further south in the west central than in the central Pyrenees, the cover thrust sheets are widely exhumed in the west. In addition, the floor thrusts of the Cotiella and Peña Montañesa thrust sheets climb up section westwards (Muñoz et al. 2013). Thus, the western lateral equivalents of the Peña Montañesa-Montsec and Cotiella-Bóixols thrusts sheets have been uplifted and eroded, whereas the Gavarnie thrust sheet occupies a wide area (Fig. 9.6). This change of the structural style was accompanied by an increase of the western plunge of the Axial Zone antiformal stack, the decrease of the structural relief in the central part of the orogen and the western termination of basement rocks at outcrop. All of them had a strong influence in the routing systems of the synorogenic sediments.

The Gavarnie thrust sheet is characterized by a wide, up to 200 km long synclinorium, which supports the Jaca Basin filled up by Eocene and Oligocene synorogenic sediments (Figs. 9.6 and 9.7). The southern limb of the synclinorium consists of several imbricates of Mesozoic and Paleogene rocks in the hangingwall of the emergent south Pyrenean frontal thrust. These imbricates have been detached on the Triassic evaporites and are referred as Sierras Exteriores (Almela and Ríos 1951; Puigdefàbregas 1975; Millán et al. 1995, 2000). Similarly to the frontal imbricates of the Serres Marginals thrust sheet, the Sierras Exteriores imbricates involve a relatively thin (up to few hundred metres thick) Mesozoic succession of synorogenic Upper Cretaceous shallow marine limestones, unconformably overlying Triassic evaporites, and a thicker Paleogene sequence characterized by the Middle Eocene limestones of the Guara Fm (Puigdefàbregas 1975; Millán et al. 1994). One of the peculiarities of the External Sierras is the presence of a set of irregularly spaced transverse NW-SE to N-S anticlines in continuation with the N-S trending anticlines of the Ainsa Basin at the eastern edge of the Gavarnie thrust sheet (Fig. 9.6). These N-S anticlines become younger westward and their growth was synchronous with the deposition of the Middle-Upper Eocene sediments and the development of the Gavarnie thrust sheet (Fig. 9.7, Millán et al. 1994; Poblet and Hardy 1995). The oldest of this transverse system of folds are the Mediano and Añisclo anticlines at the eastern edge of the Ainsa Basin, which started to develop at early Lutetian times (Poblet et al. 1998; Fernández et al. 2012; Muñoz et al. 2013). These folds experienced a vertical axial clockwise rotation as they progressively developed (Pueyo et al. 2002; Soto et al. 2006; Muñoz et al. 2013). Vertical axis rotations have been recorded all along the Sierras

Exteriores structures (Oliva-Urcia et al. 2012; Ramón et al. 2012; Pueyo-Anchuela et al. 2012). Further west, in the western Pyrenees, no significant vertical axis rotations have been recorded by paleomagnetic data (Larrasoña et al. 2003), defining a western limit for vertical axis rotation at the termination of the Santo Domingo anticline (Pueyo et al. 2002). The amount of vertical axial rotation decreases westward as the age of the onset of folding does.

The northern limb of the Jaca Basin synclinorium (also referred as Guarga synclinorium) consists of a fold and thrust belt that affects the Upper Cretaceous to Paleogene succession (Fig. 9.7). They are located above a tilted panel of basement that in detail constitutes an imbricated system of basement-involved thrust sheets merging upwards at the bottom of the Mesozoic succession (Cámara and Klimowitz 1985; Teixell 1996; Labaume et al. 2016). The southern tip of the lowermost of these thrust sheets (Guarga thrust sheet) defines the position and trend of the Guarga syncline along the axis of the Jaca synclinorium (Labaume et al. 2016) and can be followed as far as the eastern edge of the Ainsa Basin (Fernández et al. 2012). In between the Guarga thrust and the Gavarnie one different thrusts involving the Variscan basement have been described by different authors from the interpretation of available seismic data (Fig. 9.7). There is not a consensus about the number of thrusts and the given names for each individual thrust but it is widely accepted that they branch upwards into the detachment at the bottom of the Mesozoic succession of the Gavarnie thrust sheet (Cámara and Klimowitz 1985; Millán et al. 2006; Muñoz et al. 2013; Labaume et al. 2016). Seismic data do not rule out that some of these thrusts may breach through this detachment and deform the overlying Mesozoic and Paleogene sediments.

South of the basement of the Gavarnie thrust sheet at the southern edge of the Axial Zone the Upper Cretaceous-Paleogene series are deformed by the Monte Perdido-Larra fold and thrust system, which constitutes the highest structural relief of the so called Sierras Interiores (Teixell 1996). The Monte Perdido system is located in the footwall of the Peña Montañesa-Montsec thrust sheet and is tilted at the northern limb of the Añisclo anticline. So, these relationships constrain the uppermost Ypresian to lowermost Lutetian age for the Monte Perdido system (Muñoz et al. 2013). The Monte Perdido-Larra fold and thrust system continues around the western termination of the Axial Zone and branches with the Lakhora thrust (Teixell 1996, 1998). Thus, this thrust has the same structural position as the Peña Montañesa-Montsec thrust and would have been most probably active during the Early Eocene. In the hangingwall of the Lakhora thrust another south-directed thrust system would be the result of the inversion of the Lower Cretaceous basins that developed in the hyperextended rift domain and above an exhumed mantle (Lagabrielle et al. 2010; Masini

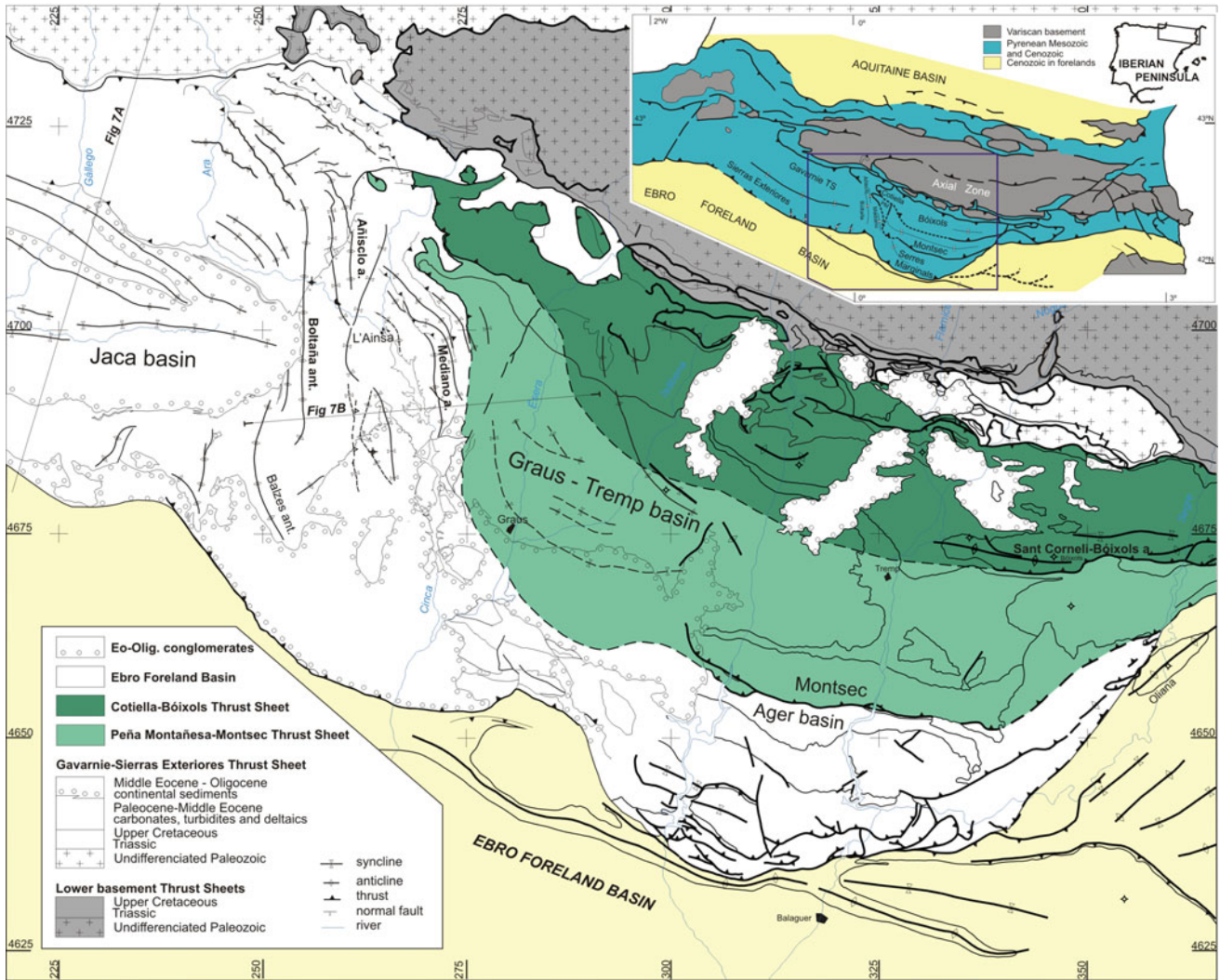


Fig. 9.6 Structural map of the transition from the central to the western Pyrenees. This transition is located at the western edge of the Central Pyrenean thrust salient, which is characterized by N-S trending folds. Modified from Muñoz et al. (2013)

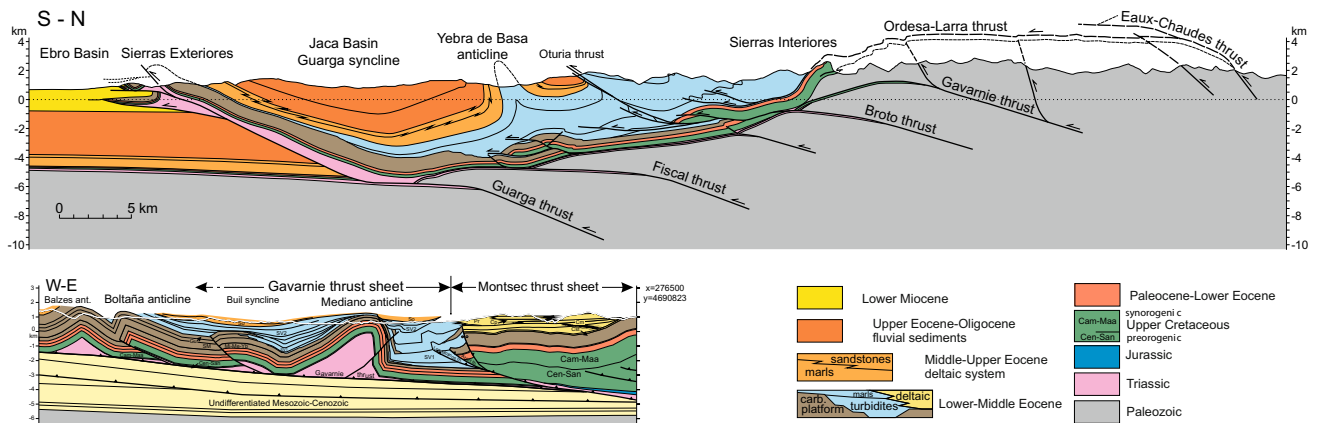


Fig. 9.7 Cross-sections across the Gavarnie thrust sheet and its relationships with the Montsec-Peña Montañesa and Cotiella thrust sheets. The upper S-N section has been modified from Labaume et al. (2016). The E-W section is from Muñoz et al. (2013)

et al. 2014). It would have developed during the Late Cretaceous, synchronously with the displacement on the North Pyrenean frontal thrust.

The Paleogene of the northern limb of the Jaca synclinorium is characterized by a 4 km thick succession of the Ypresian-Lutetian turbidites of the Hecho Group (Mutti et al. 1988). These turbidites are in continuation with the slope turbiditic systems of the Ainsa Basin and were fed axially from the Tremp-Graus and Ager basins (Mutti et al. 1988; Muñoz et al. 2013). The Ypresian lower part of the turbidites developed in a foreland trough in front of the Peña Montañesa-Montsec thrust sheet and most probably in the footwall of its western continuation into the Lakhora thrust, whereas the Lutetian upper part of the turbiditic succession were deposited piggyback above the Gavarnie thrust sheet as it developed westwards synchronously with fold growing of the at present N-S trending Ainsa-Sierras Exteriores fold system described above (Arbués et al. 2011; Muñoz et al. 2013). The turbidites onlap southward the coeval Lower-Middle Eocene platform limestones cropping out in the Sierras Exteriores (Fig. 9.7, Puigdefàbregas 1975; Labaume et al. 1985; Barnolas and Teixell 1994; Muñoz et al. 2013). From Bartonian to early Priabonian times, the turbiditic basin was progressively filled by southward and westward prograding deltaic complexes (Puigdefàbregas 1975; Dreyer et al. 1999) and by a thick continental succession, Upper Eocene-Lower Miocene in age, once the Ebro foreland basin became endorheic. Their accumulation was associated with a southward migration of the Jaca Basin depocenter with respect to the turbiditic stage, synchronously to basement underthrusting below the northern limb of the Jaca synclinorium and emergence of the frontal thrust system at Sierras Exteriores (Hogan and Burbank 1996; Teixell 1996; Labaume et al. 2016).

9.4.4 The Basque-Cantabrian Pyrenees

The Pyrenean orogenic system shows a significant structural change across the Pamplona transfer fault (Fig. 9.1). The Pamplona is an inherited transfer zone of the Early Cretaceous rift system. The overall extension of the hyperextended rift system was transferred across this fault from the Mauléon basin in the east to the Basque-Cantabrian basin in the west (Roca et al. 2011; Masini et al. 2014; Tugend et al. 2014). The Pamplona fault was probably inherited from the Variscan orogeny and not only segmented the Early Cretaceous rift system but also affected the entire orogenic system during the contractional deformation. Nonetheless, it is revealed at surface as a diffuse zone of deformation with different structures. In the south it has been interpreted as a hangingwall drop fault resulting from accommodation of the different thickness of Mesozoic sequences during

southwards displacement above the south Pyrenean frontal thrust (Larrasoña et al. 2003). Diapirs aligned along the Pamplona transfer zone would have been developed during the Early Cretaceous and reactivated during the Cenozoic deformation. In the north, the Paleozoic Basque massifs crop out on both sides of the Pamplona transfer zone (Fig. 9.1). The Cinco Villas massif is the westernmost and largest of the Basque massifs. It is involved in the northern directed thrust sheets of the Basque Pyrenees and shows a pop-up structure between inverted Lower Cretaceous basins. South of the Cinco Villas massif, the Leiza fault follows a narrow strip of lower crustal and mantle rocks as well as metamorphic Mesozoic rocks (Martínez-Torres 1989). It was considered as the western continuation of the North Pyrenean fault into the Basque Pyrenees (Choukroune and Mattauer 1978; Rat 1988) but, following the new ideas on the hyperextended non-volcanic margins, the Leiza fault has been recently interpreted as the inverted extensional detachment that produced the mantle exhumation in the Basque-Cantabrian basin (DeFelipe et al. 2017).

There is not a consensus about the terminology used by different authors to refer to the realm of the Pyrenean orogenic system located westward of the Pamplona transfer fault. Recent papers on the area refer to such realm as the Basque-Cantabrian basin (Abalos 2016; DeFelipe et al. 2017) in order to respect this widely used term in the old literature (Rat 1988) or to avoid the use of the adjective Pyrenean. This is a confusing terminology, as the area deformed by the Pyrenean structures involves a stratigraphic record from the Paleozoic to the Miocene with sedimentary units that were deposited in different basins with different aerial distribution (some of them exceeding its limits and others being more reduced) and having distinct tectonic significance. Herein, the term Basque-Cantabrian basin is used only to refer to the Lower Cretaceous extensional basin that formed as a result of the Bay of Biscay-Pyrenean rifting at the North Iberian margin.

The structure of the Basque Pyrenees is dominated by the north-verging Bilbao anticlinorium, a more than 100 km long structure located in the central part of the orogenic double wedge, above the Iberian subducted slab (Figs. 9.8 and 9.9, Pedreira 2004; Ferrer et al. 2008; Quintana et al. 2015). The anticline resulted from the inversion of the extensional faults that bounded to the north the depocentre of the deepest Lower Cretaceous basin of the Basque-Cantabrian basin. The southern limb of the Bilbao anticline is a southward tilted panel involving a thick succession of Cretaceous sediments, up to 15 km thick, that connects with a flat lying panel and the northern limb of the Miranda-Urbaña syncline and its westward continuation in the Villarcayo syncline (Figs. 9.8 and 9.9). This syncline is another main structural feature of the Basque-Cantabrian Pyrenees. It has a structural continuity with the Jaca

synclinorium across the Pamplona transfer fault and it is cored by Oligocene-Lower Miocene fluvial sediments showing growth geometries at its both limbs (Riba and Jurado 1992)

The southern limb of the Miranda-Urbaña syncline consists of several imbricates in the hangingwall of the Sierra de Cantabria frontal thrust, characterized by a relatively thin Mesozoic succession that reduces southwards, as observed in the western and central Pyrenees along the Sierras Exteriores and Serres Marginales thrust sheets. The frontal imbricates involve Upper Cretaceous carbonates, and Upper Albian sandstones (Utrillas Fm) at the bottom, unconformably overlying the Triassic evaporites. Thus, a succession characteristic of the North Iberian margin south of the Pyrenean rift system. These imbricates as well as the Miranda-Urbaña syncline lie on top of an autochthonous Paleogene and Upper Cretaceous succession of the Ebro basin as evidenced by seismic data (Carola et al. 2013).

North of the Bilbao anticlinorium a north directed thrust system involves the Mesozoic and Paleogene rocks (Fig. 9.9). This thrust system and related structures of the Basque Pyrenees are referred as the Basque Arc (Feuillée and Rat 1971) because of their concave geometry in map view (Fig. 9.8). This arc is delineated by different structures, which from south to north and apart from the Bilbao anticline include: the Biscay Synclinorium, the North-Biscay Anticlinorium and the Guipúzcoa Monocline (Abalos 2016). The northernmost Basque thrust sheets are thrust on top of the Tertiary rocks of the Landes High, which was an uplift area during the Early Cretaceous between the Parentis basin in the north and the Basque-Cantabrian basin southwards (Ferrer et al. 2008).

The transition to the Cantabrian part of the Pyrenean orogenic system occurs across the Santander transfer fault

system and related accommodation zone that was inherited from the Early Cretaceous rift system (Roca et al. 2011; Carola et al. 2015). Offshore, this transfer zone is aligned with the western edge of the Landes High and the Parentis basin and their transition to the Bay of Biscay abyssal basin (Fig. 9.1). The transfer zone follows distinct N-S trending topographic features in the Cantabrian slope (Torrelavega and Santander canyons) and produces a shift on the location of the North Pyrenean frontal thrust of more than 100 km. In the Basque Pyrenees it follows the slope close to the coast line, whereas in the Cantabrian Pyrenees, west of the Santander transfer zone, it is located further to the north in the abyssal plain (Fig. 9.1). Onshore, the N-S trending Ramales fault represents the boundary between the north-vergent structures of the Bilbao anticlinorium and the south-vergent structures of the Cantabrian Pyrenees at Santander (Cámara 1997).

Along the coast, between Santander and Oviedo, southward-directed thrusts involve basement and cover rocks and result from the reactivation of extensional Early Cretaceous faults. These thrusts are bracketed between the NW-SE trending Ventaniella fault westwards and the N-S trending Ramales fault eastwards (Fig. 9.8, Cámara 1989). This south directed thrust system continues into the Cantabrian platform where it incorporates the Tertiary Ribadesella basin (Sánchez 1991). The edge of the Cantabrian platform is characterized by north vergent thrusts and folds affecting the Mesozoic succession of the Danois basin and the basement rocks at the slope (Boillot et al. 1979; Álvarez-Marrón et al. 1995).

The Cantabrian Pyrenees are characterized by an accretionary prism deforming Tertiary and Mesozoic sediments of the Bay of Biscay oceanic basin in front of the Cantabrian slope (Fig. 9.2, Roca et al. 2011). The accretionary prism is

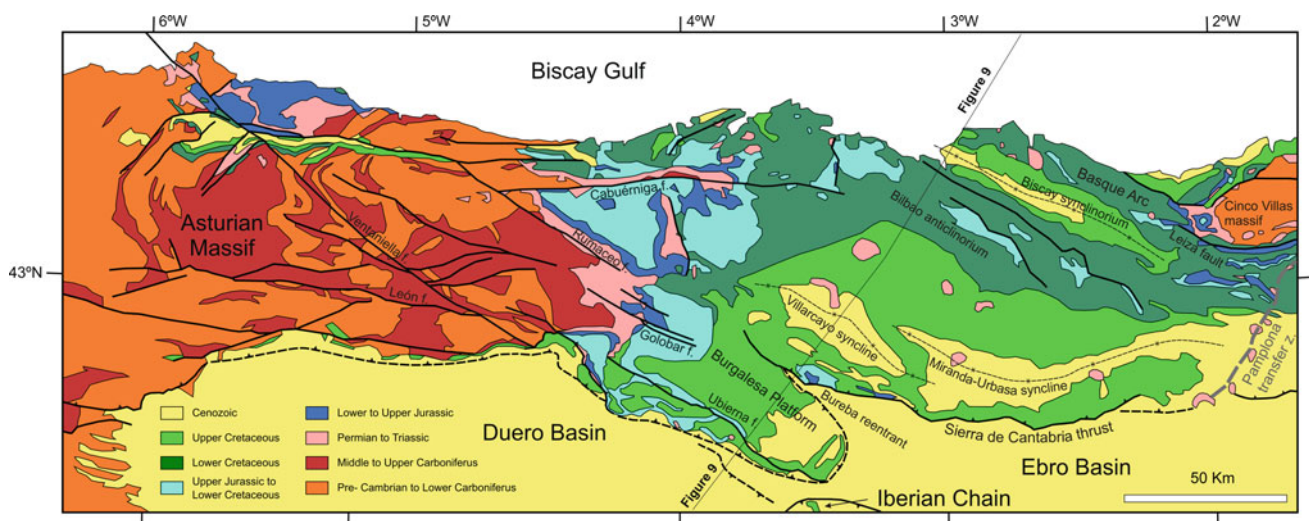


Fig. 9.8 Structural map of the Basque-Cantabrian Pyrenees

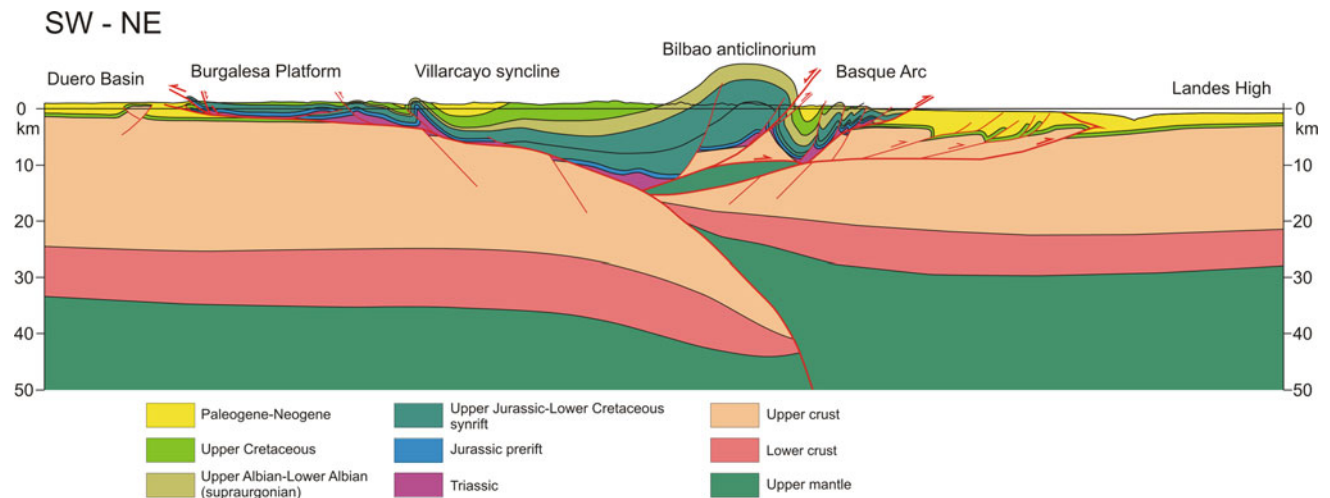


Fig. 9.9 Crustal cross-section across the Basque-Cantabrian Pyrenees. Surface geology based on Quintana et al. (2015) and Carola et al. (2015). See Fig. 9.8 for location

60 km wide in the east (Fig. 9.1) but reduces in width westwards (20 km) changing also its internal structure and morphology (Álvarez-Marrón et al. 1997).

The asymmetry of the Cantabrian realm of the Pyrenean orogenic double-wedge changes again with respect the Basque Pyrenees. Southward directed structures predominate onshore, whereas northward directed ones are mainly restricted to the accretionary prism in the Bay of Biscay abyssal plain.

The Cantabrian Mountains are characterized by a thick-skinned structural style. Upper Albian-Upper Cretaceous sediments unconformably overlie Paleozoic rocks and the absence of Triassic evaporites has determined the coupling of the basement with the cover succession. Pyrenean contractional deformation caused the reactivation of the Variscan faults and the tightening and steepening of previously developed folds (Pérez-Estaún et al. 1988; Alonso et al. 1996; Pulgar et al. 1999; Alonso et al. 2009). These structures have been interpreted to lay in the hangingwall of a shallow north-dipping intracrustal sole thrust, which at surface corresponds to the fold propagation fold and related frontal thrust at the northern boundary of the Duero basin (Fig. 9.2, Alonso et al. 1996; Pulgar et al. 1999; Gallastegui et al. 2002; Martín-González and Heredia 2011). The thick-skinned style of the Cantabrian Mountains is also evidenced by the syncline geometry of the Duero basin which contrasts with the northward tilted floor of the Ebro foreland basin in front and underneath the Pyrenean thrust front. The Paleozoic rocks of the Cantabrian Mountains show an eastern termination at surface as a result of the regional easterly plunge of the structures. Thus, the map gives us a cross-sectional view of the transition of the Cantabrian Mountains to the Basque-Cantabrian Pyrenees (Tavani et al. 2013). There, a system of WNW-ESE faults

(i.e. Cabuérniga, Rumaceo and Golobar faults) involve the basement and the cover and have a right-lateral strike-slip component (Carola et al. 2013; Tavani et al. 2013; Tavani and Granado 2014). They formed during the Permo-Triassic and were reactivated during the Late Jurassic-Early Cretaceous rift event before the contractional deformation (Espina et al. 2004; Tavani and Muñoz 2012).

The high-angle faults with a reverse and right-lateral kinematics produced an eastward lateral extrusion of the basement-involved structural units (Tavani et al. 2013). However, there is not a consensus about the structural style at the transition between the Cantabrian Mountains and the Basque Pyrenees. Different proposed models differ about the involvement of the basement of the frontal structures and consequently the degree of coupling between the basement rocks and the Mesozoic succession and the fault kinematics (See Carola et al. 2015 for a summary of different models). A thick-skinned structural model, either with dominant strike-slip kinematics (Tavani and Granado 2014) or with dip-slip kinematics (Quintana et al. 2015), has been interpreted to occur along the Plataforma Burgalesa as far east as the Sierra de Cantabria thrust front (Fig. 9.8). Nonetheless, seismic data demonstrate that the Mesozoic succession was decoupled from the basement in the areas where the Triassic salt layer was thick enough, both during the Late Jurassic-Early Cretaceous extensional deformation and the subsequent Pyrenean tectonic inversion. This is revealed by the onlap attitude of the synrift sediments above the Pre-rift Jurassic carbonates.

Thus, the transitional area between thick- and thin-skinned tectonics and from a strike-slip to a dip-slip framework, and the region accommodating the lateral extrusion of the Cantabrian area, coincides with the southern prolongation of the Santander Mesozoic Transfer Zone

(Figs. 9.1 and 9.8). This transitional area is a first order discontinuity in terms of the style of the deformation in the Pyrenean Orogen at the western limit of the Triassic evaporites.

The Basque-Cantabrian Pyrenees are characterized by salt structures involving Triassic salt (Cámara 2017). These structures are nicely preserved in the southern fold and thrust belt. There, diapirs have been moderately reactivated and even have not squeezed, suggesting the decoupling of the Mesozoic succession from the basement during the contractional deformation. On the contrary, in the northern wedge along the Basque Arc, salt structures have been significantly squeezed during the reactivation of the basement-involved faults (Fig. 9.9).

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