Eastern Pyrenees and related foreland basins: pre-, syn- and post-collisional crustal-scale cross-sections

J. Vergés*, H. Millán, E. Roca, J. A. Muñoz and M. Marzo
Dept de Geologia Dinàmica, Geofísica i Paleontologia, Univ. de Barcelona, Martí i Franquès, s/n, 08071 Barcelona, Spain

J. Cirés
Servei Geològic de Catalunya, ICC, Parc de Montjuïc, 08038 Barcelona, Spain

T. Den Bezemer, R. Zoetemeijer and S. Cloetingh
Faculty of Earth Sciences, Vrije Universiteit, Amsterdam, The Netherlands

Received 1 October 1994; revised 15 February 1995; accepted 9 May 1995

A new crustal-scale cross-section through the Eastern Pyrenees shows a minimum of 125 km of total shortening across the belt. Convergence rates of 6 mm/yr (during early and middle Eocene time) between the northern domain of the Iberian plate and Europe can be evaluated from calculated shortening rates in both sides of the orogen. Two stages of orogenic growth can be determined in the Eastern Pyrenean transect. A first stage (from Early Cretaceous to middle Lutetian time) is characterized by a low topography, submarine emplacement of the thrust front, fast rates of south-directed shortening up to 5 mm/yr and widespread marine foreland deposition. This stage is also characterized by equivalent amounts of mountain erosion and detrital foreland accumulation. A second stage (middle Lutetian to late Oligocene) is marked by an increase in structural relief, subaerial emplacement, a decrease in shortening rates and widespread continental sedimentation. This leads towards a non-equilibrium condition in which mountain erosion is almost three times the foreland basin accumulation, leading to a large bypass of sediments towards the Atlantic before the final endorheic stage of the basin. Erosion rates based on area conservation between middle Lutetian and present day sections in a two-dimensional calculation indicate an average of 0.15 mm/yr. This rise is lower than middle Lutetian to early Miocene rock uplift rates in the Eastern Pyrenees, which account for 0.2–0.35 mm/yr, suggesting that erosion has been discontinuous through time. Inferred maximum river incision rates since the middle Miocene opening of the Ebro Basin towards the Mediterranean Sea account for less than 0.1 mm/yr.

Keywords: eastern Pyrenees; balanced cross-sections; shortening rates

The growth of an orogenic wedge and the formation of a related foreland basin is the result of geodynamic processes on a lithospheric scale that are strongly dependent on the forces acting on the plates and the rheology of the lithosphere (e.g. Cloetingh et al., 1989) as well as on the geological processes acting within the orogenic wedge (Davis et al., 1983; Dahlen and Suppe, 1988). Precise control of the major processes acting during the evolution of mountain ranges and their adjacent foreland basins is not always easy to decipher, particularly in ancient orogens. However, inactive orogens occasionally preserve a geological record of major tectonic events, allowing us to better understand how mountain ranges and related flexural basins evolve through time.

A new balanced and totally restored crustal-scale cross-section through the Eastern Pyrenees based on the available geophysical and geological data is presented in this paper. In addition, a restored section during middle Lutetian time is also presented. This restoration corresponds to the final stage of marine evaporitic conditions in the Ripoll Trough.

The aim of this paper is to document and discuss the geological constraints on a crustal cross-section through the Eastern Pyrenees before, during and after contraction. This study is part of an iterative process of cross-section construction and flexural modelling in which both techniques build on one another (Millán et al., this issue).
Eastern Pyrenees crustal-scale cross-section

The Pyrenean orogen has been defined as an asymmetrical, double-wedge continental belt formed by the collision and partial subduction of the Iberian plate beneath the European plate (ECORS-Pyrenees Team, 1988; Choukroune et al., 1989; Roure et al., 1989; Muñoz, 1992), which lasted from Late Cretaceous to Oligocene—early Miocene time (Puigdefàbregas and Souquet, 1986). The southern, and most significant thrust system developed on top of the subducted Iberian plate, whereas the northern thrust system developed on top of the European plate and represents less shortening. From north to south the Pyrenean orogen comprises (Figure 1): (1) the Aquitaine Retro-foreland Basin, related to the northern Pyrenean wedge; (2) the North Pyrenean Thrust System; (3) the Axial Zone of the chain, formed by an antiformal stack of mainly basement rock units; (4) the South Pyrenean Thrust System; and (5) the Ebro Foreland Basin associated with the southern Pyrenean wedge. Brief descriptions of these structural zones are presented below, followed by a discussion of constraints on the deep structure of the Pyrenees (Figures 1 and 2a).

Aquitaine Retro-foreland Basin

The Aquitaine Retro-foreland Basin evolved by flexure of the upper plate in response to the load of the north Pyrenean thrust sheets (Brunet, 1986; Desegaulx et al., 1990). Syn-orogenic sedimentary infilling started in Late Cretaceous time with the deposition of turbidites in the Nalzen Basin (e.g. Dèramond et al., 1993; Figure 2b). On top of the Palaeocene red beds, early Eocene (Ilerdian) deposition consists of shelfal platform carbonates overlain by up to 4750 m of alluvial and fluvial deposits (Buis and Rey, 1975). These continental deposits, the ‘Poudinges de Palassou’, consist of three depositional sequences separated by angular unconformities (Buis and Cugny, 1978; Baby et al., 1988). These deposits are dated as Ilerdian to Bartonian in age (Buis and Cugny, 1978), although a more recent interpretation suggests that the uppermost sediments have been deposited (Fischer, 1984).

Older syn-tectonic sediments of the Aquitaine Basin are involved in the North Pyrenean Thrust System, whereas Eocene deposits are only deformed by a synclinal structure in front of the Sub-Pyrenean thrust wedge (Cavallé and Paris, 1976; Biлотte et al., 1988) (Figures 1 and 2a).

North Pyrenean Thrust System

The North Pyrenean Thrust System consists of an imbricate system of north-directed thrust faults. These thrust faults involve Hercynian basement and Mesozoic to lower Eocene cover rocks (Fischer, 1984; Souquet, 1988; Baby et al., 1988; Dèramond et al., 1993). Basement rocks comprising the north Pyrenean massifs, as at Saint Barthélémy (Figures 1 and 2), are involved in the thrust system as short-cuts of Mesozoic extensional faults (Souquet and Peybernès, 1987; Roure et al., 1989). The frontal structure is comprised of an anticline, which has been interpreted as the surficial expression of a duplex (Baby et al., 1988).

Axial Zone: antiformal stack basement rocks thrust sheets

The Axial Zone of the Pyrenean belt consists of up to 10 km of Cambrian to Carboniferous rocks affected by low- to high-grade metamorphism during the Hercynian orogeny (e.g. Vissers, 1992). These rocks were deformed by Hercynian compressional events in late Carboniferous time and intruded by crustally derived granitic bodies during the early Permian time (Zwart, 1979).

Post-Hercynian rocks consist of up to 2.5 km of continental sedimentary and volcanic rocks ranging from late Carboniferous to early Triassic in age, unconformably overlying deformed Palaeozoic rocks. These rocks crop out on top of the Orri Unit in this transect (Figure 2a). The involvement of these post-Hercynian rocks in the south-vergent system of thrusts indicates that thrusting is Late Cretaceous–Oligocene (Alpine) in age (Séguret, 1972; Poblet, 1991; Muñoz, 1992). Sparse fission track ages also corroborate the Tertiary age for the antiformal stack development as described herein (Garwin, unpublished data; Yelland, 1990).

The Axial Zone is comprised of an antiformal stack of three main tectonic units named, from top to bottom, Nogueres, Orri and Rialp (Muñoz, 1992; Figure 2a). In addition, pervasive deformation of the basement units suggests that internal deformation accounts for up to 50% of Alpine shortening (e.g. Muñoz, 1992). This important internal deformation of the basement units has been taken into account in the area balance.

South Pyrenean Thrust System

The South Pyrenean Thrust System is formed by a stack of three different cover thrust sheets: the Upper Pedraforca, the Lower Pedraforca and the Cadi thrust sheets (Figure 2). Each thrust sheet displays a different stratigraphy with the oldest rocks in the uppermost and first emplaced Upper Pedraforca thrust sheet. The Upper Pedraforca unit is comprised mainly of lower Cretaceous strata and is an inverted lower Cretaceous basin (Vergés and Martinez, 1988). The sole thrust of the Upper Pedraforca unit dips gently to the south, due to later southwards structural tilting of the whole stack of thrust sheets. The Upper Pedraforca unit is the eastern continuation of the Bóixols thrust sheet (Berastegui et al., 1990). The Lower Pedraforca thrust sheet is mainly composed of upper Cretaceous to lower Eocene rocks. These rocks are deformed by an imbricate set of thrusts, detached above an upper Triassic (Keuper) salt horizon. This horizon acts as the major detachment between the cover and basement rocks. Nevertheless, cross-cutting relationships between thrust faults constrain the age and sequence of emplacement of the basement units and their links with cover thrust sheets.

The Cadi thrust sheet (Figure 2) consists of a 2.5 km thick Stephano-Permian and lower Triassic continental section which unconformably overlies basement rocks (Gisbert, 1980; Speksnijder, 1986). Above a thin Mesozoic section, syn-orogenic Palaeocene and largely Eocene sedimentation took place in a narrow basin, the Ripoll Trough, parallel to the active thrust front. The
Figure 1 Structural map of the Pyrenees combined with the depth of the Moho. Studied crustal section (thick line) passes through the Aquitaine Basin, Eastern Pyrenees and the Ebro Basin. The Eastern Pyrenees consists of North Pyrenean Thrust System (stippled patterns), basement Axial Zone (ruled pattern) and the South Pyrenean Thrust System (stippled patterns). The map also shows the location of deep crustal-scale reflection profiles across the Pyrenees and Catalan Ranges (thin continuous lines are deep seismic reflection profiles and thin broken lines are seismic refraction profiles). Crustal thicknesses come from geophysical data. For the Eastern and Central Pyrenees: refraction data (Daignieres et al., 1982), reflection data (Roure et al., 1989, Berastegui et al., 1993), gravity data (Torné et al., 1989). For the Western Pyrenees: reflection and gravity data (Grandjean-Gregoire, 1992). For the Bay of Biscay: reflection and gravity (Lefort, 1993).
Eastern Pyrenees and related foreland basins: J. Vergés et al.

Figure 2 (a) Post-orogenic balanced cross-section through the Eastern Pyrenees and adjacent foreland basins. The French side of the orogen is mainly based on Bally et al. (1988). The crustal geometry is based on the results of the Pyrenees ECORS profile (Roure et al., 1989; Choukroune et al., 1989; Mufioz, 1992). The upper crustal rocks have been balanced by the length of the strata. Basement units and the lower layered crust have been balanced by area (solid black dot marks the presently exposed base of the Nogueres basement unit). Cross-cutting relationships between lower crustal layering and Moho have been observed in both the Iberian and European plates (Berastegui et al., 1993). P, Puig-reig anticline. See location of cross-section in Figure 1.

(b) Composite restored section showing the unfolding of all the sedimentary basins and basement units. Four different stratigraphic levels have been taken as horizontal reference levels. From south to north, the top of the Cardona Formation, the top of the Beuda Formation, the top of the Palaeocene section and the base of the Late Cretaceous section. The cross-section displays the width and migration of the depocentres. The intrabasinal bulge marks the boundary between Pyrenean and Catalan deflection of the basin during Eocene time. Surface and subsurface geometry between the lower Cretaceous basins is speculative.
Ripoll Trough was later incorporated into the Pyrenean thrust system as a piggyback basin (Puigdefàbrigas et al., 1986). The lower to middle Eocene marine section within the Ripoll Trough consists primarily of five different formations, which correspond in all cases to depositional sequences (Puigdefàbrigas et al., 1986; Giménez, 1989; Figure 3): (1) Cadi (platform carbonates); (2) Coronas (deltaic deposits); (3) Armàncies (slope deposits); (4) Campdevànil (turbidites); and (5) Beuda-Vallfogona (evaporites).

The Armàncies Formation (Almela, 1958) is exposed along the northern flank of the Cadi thrust sheet (Figure 2a). Near our line of cross-section, the Armàncies Formation is comprised of about 750 m of alternating marls and marly limestones, interpreted as talus deposits of Cuisian age (Gich, 1969; Busquets, 1981; Giménez, 1993; Figure 3). The base of this unit is a 210 m thick alternation of marly limestones and dark marly pelites with abundant bitumen, showing oil seeps at the surface (García-Valls et al., 1985; Permuy et al., 1988; Giménez, 1993). Total organic carbon contents and pyrolysis data (Rock–Eval) show that the lower part of the Armàncies Formation has excellent potential for petroleum from marine type II organic matter (García-Valls et al., 1985; Clavell, 1992).

The southward extension of the middle Lutetian marine foreland basin (the Ripoll Trough) is controlled by the southern pinch-out of the Beuda evaporites (Figure 3). The Ripoll Trough (Figure 2b) became progressively subaerial (Bellmunt Formation) during its piggyback motion in the hanging wall of the Vallfogona thrust, the frontal emergent thrust of the south-eastern Pyrenean thrust system. The Ripoll Trough strata are presently folded within the Ripoll syncline, the largest structure within the Cadi thrust sheet (Figure 2a).

The structural stack of the Upper and Lower Pedraforca and Cadi thrust sheets was folded into the Ripoll syncline due to the coeval southward emergence of the Vallfogona thrust and the growth of the antiformal stack to the north (Muñoz et al., 1986; Burbank et al., 1992a). During growth of the syncline, deformation within the Upper and Lower Pedraforca thrust sheets was accomplished mainly by bedding- and thrust-parallel reactivation. The reactivation of older fault surfaces was achieved by backbreak thrusting and backthrusting (Martínez et al., 1988), in many cases overprinting the original contacts.

**Ebro Foreland Basin**

The stratigraphy of the Ebro Basin consists of Palaeocene red beds, lower and middle Eocene platform carbonates, middle Eocene Beuda evaporites overlain by a thick succession of middle and upper Eocene marls corresponding to distal parts of alluvial fans and deltas attached to the thrust fronts of Pyrenees and Catalan Ranges. The change from open marine to widespread continental deposition and confined conditions in the Ebro Foreland Basin took place after the deposition of the Cardona Salt, during early Priabonian time, at ~ 37 Ma (Puigdefàbrigas et al., 1986; Burbank et al., 1992b; Vergés and Burbank, in press). The youngest exposed continental sediments in the eastern region of the Ebro Basin are late early Oligocene (Agustí et al., 1987; Sáez, 1987; Cuenca, 1991).

South of the Vallfogona thrust, seismic studies suggest the existence of a blind thrust separating well-imaged subhorizontal reflectors beneath folded conglomerates (Vergés et al., 1992). The Puig-reig anticline, in the northern part of the fold and thrust system, represents the surficial expression of a thrust–ramp geometry at depth (Figure 2a). This ramp geometry allows the Pyrenean sole thrust to climb from the lower Beuda décollement to the structurally higher Cardona décollement. A detached fold and thrust system is developed above the Cardona décollement level south of the Puig-reig anticline (Vergés et al., 1992).

The variation in thickness of the middle and upper Eocene marine strata below the originally horizontal upper Cardona salt level shows a double flexion of the basin with an intrabasinal high (‘bulge’ on Figure 2). This middle and upper Eocene double-wedge basin geometry may be related to the effect of the Pyrenean load to the north and the load of the Catalan Ranges to the south-east (Figure 2b). Development of a late Lutetian intrabasinal bulge was roughly coeval with increased deformation along the Catalan margin of the Ebro Basin (Anadón et al., 1985; Guimerà, 1988; Colombo and Vergés, 1992).

The distribution within the Ebro Basin of sediments derived from the Pyrenees and Catalan ranges is of great importance in calculating sediment flux in the basin. Beneath the Cardona salt, the intrabasinal bulge has been taken as the boundary between north and south derived distal deltaic sediments (Figure 2). Above the Cardona salt, the distribution of palaeocurrent directions in the alluvial and fluvial Solsona depositional sequence exhibits a marked limit between north and south provenance coinciding with the Súria anticline (Malmsheimer and Mensink, 1979; Sáez, 1988).
The boundary between the Pyrenean and Catalan provenances has thus migrated = 12 km southwards with respect to the previous provenance boundary (Figure 2).

The regional northwards dip of the southern part of the Ebro Foreland Basin resulted from Neogene extension in the Catalan Ranges and Valencia Trough (Morgan and Fernández, 1990; Janssen et al., 1993).

**Constraints on the deep structure**

We have projected the deep structural data from the ECORS profile (Muñoz, 1992) due to: (a) the proximity of the two sections (55 km apart); (b) the subhorizontal E–W disposition of the Moho in this segment (Daignières et al., 1982); (c) the similar distribution of Bouguer gravity anomalies (Casas et al., 1987); (d) the almost identical amount of shortening (Muñoz, 1992; Vergés et al., 1992), despite changes in geometry along-strike related to geometry inherited from Mesozoic extensional structure (Vergés, 1993); and (e) the cartographic continuity of the main thrusts involving basement rocks in the inner part of the chain (Figure 1).

The depth to the Moho has been constrained using different geophysical sources (Daignières et al., 1982; Choukroune et al., 1989; Torné et al., 1989; Surínach et al., 1992). In our transect, the Iberian Moho increases from 33 km depth beneath the undeformed Ebro Basin to a maximum depth of = 55 km of the northern boundary of the Iberian plate. However, two-dimensional gravity modelling along the ECORS–Pyrenees profile suggests greater depths (Torné et al., 1989), which is supported by magnetotelluric data (Pous et al., 1995). At the southern margin of the European plate, the Moho is located at ~ 32 km depth and at ~ 30 km depth beneath the Aquitaine Basin.

Reflection data shows a very consistent, layered lower crust in both Iberia and Europe (Choukroune et al., 1989; Rouge et al., 1989). The thickness and position of the layered lower crust in our cross-section is based on reflection data from the ECORS–Pyrenees profile (Berastegui et al., 1993).

**Ages of thrusting: North and South Pyrenees**

Both the North and South Pyrenean Thrust Systems display a general forward or piggyback, thrusting sequence (Puigdefábregas et al., 1992; Démamond et al., 1993), which developed synchronously with the growth of local breakback sequences (Martínez et al., 1988; Vergés and Muñoz, 1990) in the South Pyrenean Thrust System.

Compressional deformation related to plate convergence started earlier in the previously thinned European plate and ended later in the south-directed thrust system overthrusting the Iberian plate (Desegaulx et al., 1990). Although the timing of the initial thrust motion is not very well constrained in this particular transect, the available data indicates a roughly continuous shortening involving both the South and North Pyrenean Thrust Systems at least from Late Cretaceous to middle Eocene time. However, it is difficult to relate specific thrusts on each side of the belt.

The North Pyrenean Thrust System has undergone nearly continuous shortening from Late Cretaceous to late Eocene time (Démamond et al., 1993). The North Pyrenean Frontal Thrust was active during Late Cretaceous time, whereas Palaeocene to upper Eocene shortening was propagating forward to the Sub-Pyrenean Frontal Thrust at the southern margin of the Aquitaine Basin (Crochet, 1989; Démamond et al., 1993; Figure 2). The youngest deposits exposed at the southern margin of the Aquitaine Basin are, however, gently folded (Cavallé and Paris, 1976; Biolote et al., 1988), suggesting a post-Eocene age for the end of deformation (Figure 2).

Deformation in the southern side of the present day Pyrenees started during the latest Cretaceous time (Simó and Puigdefábregas, 1985; Puigdefábregas and Souquet, 1986; Démamond et al., 1993). In the studied transect, the Upper Pedraforca thrust sheet was fossilized by uppermost Palaeocene conglomerates (Vergés and Martinez, 1988; Figures 3 and 4). The emplacement of the Lower Pedraforca thrust sheet below sea level is proved by marine syn-tectonic strata linked to the front of the nappe (Solé Surañes and Clavell, 1973). This emplacement took place between the deposition of the Cadi and the Beuda formations (Figures 3 and 4). Following emplacement = 47 Ma (Burbank et al., 1992a; Vergás and Burbank, in press), the thrust sheet was covered by up to 1600 m of continental conglomerates of the Bellmunt Formation (Figure 3). Motion along the emergent Vallfogona thrust and its blind continuation, beneath the foreland basin marginal conglomerates, took place from middle Eocene to the latest early Oligocene time (Vergés, 1993). Detached folds located south of the Vallfogona thrust evolved partially after deposition of the uppermost foreland deposit in the Eastern Ebro Foreland Basin to attain the present disposition (Vergés et al., 1992). Although there was deformation after the younger preserved strata on this part of the basin, shortening since the lower Oligocene accounts only for ~ 1 km, based on deformation on these strata (Vergés et al., 1992; Vergés and Burbank, in press).

Fission track dating of zircons and apatites from the basement units of the Axial Zone suggest a general decrease in cooling ages from ~ 50 Ma in the north
Pre-collisional restored cross-section

The pre-collisional restored cross-section has been made using line-length and area balancing methods (Figure 2b). The strata involved in the foreland fold and thrust belt and cover thrust sheets have been line-length balanced (Baby et al., 1988; Vergés, 1993). The basement units have been areally balanced to allow changes in thickness during their emplacement (Figure 2b).

The restored cross-section is thus constructed based on three additional constraints. (1) Linkages between cover and basement thrust sheets, deduced from branch line maps and cross-cutting relationships among thrusts constrain the position and time of emplacement of structural units (Vergés, 1993). The upper and lower Pedraforca thrust sheets link with the Nogueres basement unit, the Cadi thrust sheet with the Orri basement unit and the detached foreland fold and thrust belt with the Rialp basement unit. (2) Undefomed basement units in the restored cross-section are longer and thinner, accounting for the observed internal Alpine deformation. In the restored section, the basal thrust of each basement unit displays a gentle dip to the north, resulting in a significant superposition of units. (3) The brittle upper crust is detached above a basal detachment at ~15 km depth, although in the restored cross-section rocks exposed at the base of the Nogueres basement unit are located at ~11 km depth (see solid black square; Figure 2). This maximum burial depth is consistent with relatively weak metamorphism (green-schist facies) affecting Cambro-Ordovician rocks at the base of the Nogueres basement unit (Bons, 1989).

Restoration of the deep geometry of the crust below the intracrustal detachment level is speculative. The section shows a ~32–35 km thick crust where not extended, which is the average thickness for the Iberian crust (Banda, 1987). However, the reconstruction of the central region of the section is difficult due to the lack of data constraining the geometry of the Early Cretaceous extensional basins in this transect.

The Bóixols and Tarascon lower Mesozoic basins are separated by ~75 km in our reconstruction. Whether they were the southern and northern expressions of a continuous extensional basin, or two small basins separated by a structural high, is important in reconstructing the geometry of the deep portions of the pre-collisional section, especially where seismic data along the ECORS profile shows a layered lower crust (Choukroune et al., 1989; Berastegui et al., 1993). Assuming that the present cross-cutting relationships between the layered lower crust and the Moho have been preserved since Early Cretaceous time, the reconstructed lower crust has been slightly thinned below the restored position of the Mesozoic basins, beginning at the point where the internal layering seems to be oblique to the base of the crust in both the Iberian and European plates. In this restoration, the preserved lower crust in the present cross-section only accounts for a small amount of the area between the Bóixols and Tarascon basins (shaded layered lower crust in Figure 2b).

The reconstruction also indicates a 75 km wide gap in the lower crust (Figure 2), as previously noticed (Roure et al., 1989; Muñoz, 1992). This absence of lower crust below the Iberian–European plate boundary could be explained by two different solutions. One possibility would be a configuration in which the lower crust was only thinned in very localized areas beneath the Early Cretaceous basins, similar to the northern margin of the Bay of Biscay (Le Pichon and Barbier, 1987; Boillot and Malod, 1988) where there is a well developed lower crust. The second possibility would be a generalized thinning of the lower crust as has been interpreted in the ECORS profile (Muñoz, 1992). Although both solutions may be possible, the first possibility seems more reasonable in this transect because extension related to the counterclockwise rotation of the Iberian plate in Early Cretaceous time was less intense than in the western regions of the Pyrenees adjacent to the Bay of Biscay. The interpretation of the crustal geometry presented here is, however, compatible with subduction of the lower crust beneath the European plate (Muñoz, 1992), in agreement with recent magnetotelluric studies indicating the existence of crustal rocks to depth of ~100 km (Pous et al., 1995).

Shortening rates

Total shortening calculated from balanced and restored cross-sections is ~125 km (47%), between the calculated 120 km (Roure et al., 1989) and 147 km (Muñoz, 1992) along the ECORS profile. This total shortening includes 70 km in the South Pyrenean cover thrust system (Vergés, 1993), ~23 km of internal deformation within the basement units and ~32 km in the North Pyrenean cover and basement thrust system (Baby et al., 1988).

Rates of shortening for the South Pyrenean Thrust System, based on balanced cross-sections and palinspastic maps (Vergés, 1993), show a very slow rate of less than 0.5 mm/yr for Late Cretaceous and Palaeocene time, followed by a relatively fast rate of shortening up to 4.5 mm/yr during early to middle Eocene time. From the middle Lutetian (~47 Ma) to the end of thrusting, the rates are nearly constant at ~2 mm/yr (Figure 4). The North Pyrenean system exhibits an average shortening rate of almost 1 mm/yr throughout the period of deformation, although the lower–middle Eocene time was a period of intensified deformation. These shortening rates support a convergence rate between the north-eastern part of the Iberian plate (the present Ebro Foreland Basin) and Europe or between Africa and Europe, it is necessary to take into account the widespread deformation within the eastern domain of the Iberian plate.
Middle Lutetian syn-collisional restored cross-section

Palaeotopography is an important factor during the growth of an orogenic belt in that it drives in the flexure of the underlying lithosphere. However, topographic reconstructions of ancient stages of mountain belts are always difficult. In the external parts of the Pyrenees, preserved syn-tectonic sediments overlying older palaeoreliefs can be used to reconstruct topographies at the time of thrusting (Mellere, 1993; Burbank and Vergés, 1994). However, the scarcity of syn-tectonic sediments in the internal parts of the chain necessitates the use of indirect observational data (structural and stratigraphic constraints and palaeo-botanical assemblages), providing some insights about palaeotopography. Although these data are limited and not conclusive, they allow a first approximation and a comparison with the results obtained from flexural modelling techniques (Millán et al., this issue).

Structural constraints

Partitioning of shortening in the South Pyrenean Thrust System indicates a minimum of ≈ 43 km of shortening from the beginning of contraction (~ 80 Ma, Late Cretaceous) to the end of the Lower Pedraforca emplacement in middle Lutetian time (Martínez et al., 1988; Burbank et al., 1992a; Vergés, 1993).

Experimental and analytical models involving a partial lithospheric subduction (Malavielle, 1984; Koons, 1990; Quinlan et al., 1993) indicate a maximum topographic elevation located above the leading edge of the non-subducted plate, which acts as a backstop to further overthrusting. These models also suggest that the topographic slope is steepest on the back side of the orogen. These features are consistent with the marked asymmetry of the Pyrenean belt, the subduction of lower crust and the generally forward progression of deformation in both sides of the orogen.

Stratigraphic constraints

Our middle Lutetian reconstruction corresponds to the final stages of the marine evaporitic infilling of the Ripoll Trough. The trough was narrow and elongated parallel to the Pyrenean thrust front. The Atlantic connection of the trough was structurally restricted by the emplacement of the large South Central Unit (Vergés, 1993). This restricted basin presumably entrapped all of the sediments derived from the south-eastern Pyrenees. For this reason it seems reasonable to assume an approximate balance between clastic input and erosion of the south-eastern side of the Pyrenean chain during middle Lutetian time.

Middle Lutetian marine sediments pinch-out southward of the Puig-reig oil well (Figures 5 and 6), grading into fine-grained continental sediments to the south (Puigdefabregas et al., 1986). Thicknesses of lower and middle Eocene deposits allow calculation of the geometry of the deflected basin during this time. The calculated deflection in the restored cross-section agrees with the minimum depth of burial for the lower Armancies source rocks during the emplacement of the Lower Pedraforca thrust sheet that has been calculated as minimum as 3 km (Figure 6). This is consistent with $T_{\text{max}}$ from Rock-Eval analysis, which indicates that the source rocks moved downward through the upper part of the oil window area (Permanyer et al., 1988; Clavell, 1992).

Analysis of conglomeratic clasts of the Ripoll Basin indicates an abrupt change in composition after middle Lutetian time (Busquets, 1981; Puigdefabregas et al., 1986). This change is marked by a massive arrival of basement rocks in the Ripoll Trough, indicating unroofing of the inner part of the Pyrenean chain.

Palaeo-botanical assemblages

Studies of Palaeogene floral assemblages carried out in the eastern part of the Ebro Foreland Basin indicate an increase in local specimens, corresponding to periods of isolation linked to the Pyrenean uplift, during Lutetian (?) and middle Bartonian time (Anadón et al., 1992). In addition, limited floral data indicate the development of an altitudinal zonation, possibly related to the formation of subaerial Pyrenean relief, during upper Lutetian and Bartonian deposition, but not before (Busquets et al., 1992).

Crustal-scale restored section

Considering the problems inherent in two-dimensional

---

**Figure 5** Syn-orogenic middle Lutetian crustal-scale restoration. This reconstruction assumes that erosion (ruled pattern) is equivalent to the early to middle Eocene clastic basin-fill. The Nogueres basement unit has been internally thickened during its emplacement.
Figure 6. Reconstruction of the middle Lutetian basin at the end of the Lower Pedraforca thrust sheet emplacement (cross-section with no vertical exaggeration). Source rocks were buried to 3 km depth (minimum) during the emplacement of the Lower Pedraforca thrust sheet and passed through the upper part of the oil window area, calculated by $T_{\text{max}}$ from Rock-Eval analysis.
balancing between erosion and basin infill, we have
constructed the middle Lutetian crustal-scale cross-
section assuming an erosional product (hatched area in
the syn-collisional cross-section from the inferred water
divide to the south, Figure 5) equivalent to the detrital
sedimentary infill of the basin of \( \approx 40 \) km\(^2\). As in the
pre-collisional restored section, the middle Lutetian
reconstruction preserves the line-length of the foreland
basin strata and the area for the basement units (Figure
6). The Nogueres unit is thickened during thrusting,
whereas the underlying units are slightly flexed down-
wards due to flexural subsidence of the Iberian litho-
sphere. Fission track analysis in apatites of the Orri unit
indicating a partial annealing, probably caused by
tectonic stacking (Yelland, 1990), corroborate our
results.

The middle Lutetian restoration shows a maximum
structural topography of \( \approx 1 \) km following erosion.
This maximum topography is located above and a little
south of the leading edge of the European plate and
progressively decreases towards both sides of the
orogen. Although different topographies (under
\( \approx 2.75 \) km) match well with the calculated middle
Lutetian deflection of the top of the basement, we
chose \( \approx 1 \) km of topography based on the results of
flexural modelling (Millán et al., this issue). For
maximum topographies ranging from 1 to 2 km, the
calculated flexural rigidities of the non-deformed
portion of the Iberian lithosphere are similar to the
values obtained for the same region of the present day
section (effective elastic thickness of 27–28 km, Millán
et al., this issue).

**Discussion**

**Orogenic growth**

Our syn- and post-orogenic crustal-scale cross-sections,
made by combining geological and geophysical data,
suggest two stages of orogenic growth. The first stage,
during early and middle Eocene time, involved
submarine emplacement of the Lower Pedraforca
thrust sheet subsequent to the tectonic inversion of the
Upper Pedraforca thrust sheet (Figures 5 and 6). The
passage from marine to continental conditions in the
Ripoll Trough is marked by a decrease in the rate of
south Pyrenean shortening from 4.5 to 2.0 mm/yr
(Figure 4), synchronous with the deposition of an
extensive evaporitic sequence (the Beuda Formation,
Figures 3 and 6). Our middle Lutetian reconstruction is
caracterized by low–medium topography (maximum of
1–2 km) near and above the leading edge of the
European plate, progressively decreasing towards the
external parts of the orogen (Millán et al., this issue).
During this first stage of orogenic growth, internal
thickening of the Nogueres basement unit, and corre-
sponding increase in relief was in part compensated by
downward flexion of the Iberian plate.

The second stage of orogenic growth resulting in the
present day geometry, roughly coincides with wide-
spread continental sedimentation in the Ebro Foreland
Basin (beginning in middle Lutetian time in the Ripoll
Trough). Although rates of shortening after middle
Lutetian time decreased, the Ripoll Trough was rapidly
infilled by continental conglomerates. These record a
massive arrival of clasts coming from the Orri unit,
suggesting that the Nogueres unit was already deeply
eroded. Comparison between the syn-collisional
section in middle Lutetian time and the present day
post-collisional section allows the calculation of rates of
erosion.

**Rates of erosion**

Estimates of orogenic erosion are always a first
approximation in any two-dimensional approach.
Nevertheless, from the initiation of thrusting, total
erosional product from the southern side of the orogen
is \( \approx 354 \) km\(^2\) (southwards of the inferred water divide
line for the present day crustal section; Figure 2a). Of
this total, \( \approx 314 \) km\(^2\) have been eroded from middle
Lutetian time to the present. By contrast, middle
Lutetian to Oligocene sedimentary infill of the Ebro
Basin accounts for only 117 km\(^2\), \( \approx 2.6 \) times smaller
than the erosional product on this transect (increasing
to 2.9 times if we use mass balance instead of area
conservation). However, during this period of time,
sedimentation on the south side of the Pyrenees was
characterized by a large westward bypassing of
sediments into the Atlantic Ocean, perhaps accounting
for the deficit sediment. Bypassing took place in both
the Ebro Foreland Basin and Aquitaine Retro-foreland
side of the Pyrenees, bypassing was coeval with
extensive late Eocene–Oligocene depocentre progra-
dation from east to west and before the Miocene
deforation which affected the western end of the
chain (Riba and Jurado, 1992), thereby closing the
connection between the South Pyrenean Foreland
Basin and the Atlantic Ocean.

Our calculated maximum erosion rate in the highest
topographic culmination of the chain from middle
Lutetian time (\( \approx 47 \) Ma) to the present is
\( \approx 0.3 \) mm/yr, although \( \approx 0.15 \) mm/yr could represent
the mean erosion rate for the same interval. These
values are comparable with calculated rock uplift rates
in the eastern Pyrenees throughout its evolution.
Cuisian Armâncies marls (Figure 3) containing
petroleum were uplifted to the syn-orogenic topo-
graphic surface during the late Eocene–early
Oligocene, as deduced by the position of syn-tectonic,
slightly deformed, basement source conglomerates
covering the previously formed relief. The relative
position of these unconformable subaerial
conglomerates constrain the minimum rate of uplift of
these rocks from 0.2 to 0.35 mm/yr between \( \approx 46 \) and
\( \approx 37–32 \) Ma (Figure 7). Rock uplift rates calculated by
fission track analysis between \( \approx 33 \) and 20 Ma in the
Eastern Pyrenees basement units also indicate
0.2–0.3 mm/yr (Garwin, unpublished data).

As deduced from the previous data, calculated rock
uplift rates are roughly homogeneous from the middle
Eocene to the early Miocene and slightly higher than
the mean erosion rate from the middle Lutetian to the
present, calculated from our reconstructions. This,
together, with the fact that sedimentation in the basin
slowed down coevally with the end of shortening,
suggests that erosion rates might decrease after the end
of thrusting in uppermost early Oligocene time (Vergès
and Burbank, in press). Continued rock uplift in early
Miocene time, after the end of contraction in the
Eastern Pyrenees and related foreland basins: J. Vergés et al.

Conclusions

Present day and middle Lutetian crustal-scale cross-sections through the Eastern Pyrenees and Ebro Foreland Basin have been constructed based on available geological and geophysical constraints. Our middle Lutetian crustal section reproduces final emplacement Lower Pedraforca and Nogueres thrust sheets, coeval with thick evaporitic deposition in the Ripoll Trough. Subaerial topography within the orogenic belt is based on stratigraphic, structural and palaeo-botanical data. Furthermore, the section agrees with the results of flexural modelling (Millán et al., this issue). From the sections, two stages of orogenic growth can be deduced. The first stage of growth is characterized by submarine emplacement of the front of the thrust sheets, low-medium subaerial relief in the internal regions of the belt, widespread marine sedimentation in the Southern Foreland Basin and mass balance between erosion and detrital deposits, despite the high rates of shortening corresponding to this period of time (up to 5 mm/yr). The second stage of orogenic growth, after middle Lutetian time, is marked by a decrease in shortening rates, increase of subaerial relief, massive continental sedimentation and almost three times more erosion from the range than mass detrital deposition in the basin, consistent with significant westward bypass of material. We have calculated average erosion rates (0.15 mm/yr), based on cross-section reconstructions, and we have compared erosion rates with rock uplift rates (0.2–0.35 mm/yr), deduced from geological, geochemical and fission track studies.

Acknowledgements

We thank our colleagues X. Berastegui, M. Torné, M. Fernández, P. Busquets, A. Permanyer, M. Arcuëllo, P. Hogan and C. Lewis for their constructive discussions and F. Roure, J. L. Mugnier and an anonymous reviewer for their critical and helpful comments. P. Hogan and C. Lewis greatly improved the English of the manuscript. Work was founded by IBS Project, Joule II Programme (JOU2-CT92-110) and DGICYT projects PB91-0252 and PB91-0805.

References


Eastern Pyrenees and related foreland basins: J. Vergés et al.


Busquets, P., Ramos-Guerrero, E., Moyà, S., Agustí, J., Barcelona, 1–543.


De Sitter, L. U. (1952) *Pliocene uplift of Tertiary Mountain Chains*.


De Sitter, L. U. (1952) *Pliocene uplift of Tertiary Mountain Chains*.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.

D’Orey et al.