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Alpine deformation at the western termination of the axial zone, Southern Pyrenees

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Abstract: Detailed structural and sedimentological research has been conducted in order to unravel the local tectono-sedimentary history of an area located at the western termination of the axial zone, in the Southern Pyrenees. Two cross-sections have been constructed perpendicular to the axis of the orogen, situated above the westward dipping axial zone antiform and the northern part of the south Pyrenean zone, striking perpendicular to the axis of the orogen. The sediments of this area date from Late Cretaceous to Middle Eocene age and display syntectonic characteristics. The cross-sections, together with detailed analysis of key outcrops, reveal that there have been two main phases of deformation. The first phase (D1) is related to the movement along the Lakhoura basement thrust and can be subdivided into three sub-phases, related to activity of two main thrusts, which splay of the basement thrust. The first sub-phase is related to activity along the Lakhoura thrust (D1a), the second sub-phase to the Larra thrust (D1b) and the third sub-phase to reactivation of the Lakhoura thrust (D1c). The second phase (D2) is related to the Gavarnie basement thrust, which resulted in the formation of the axial zone antiform. D1 and D2 show their distinct type of deformation. The first phase is characterised by south vergent ramp-flat thrusting and fault propagation folding, whereas the second phase is characterised by upright to overturned folding and steep reverse faulting. One main structure, the Urzainqui fault propagation fold, shows a synsedimentary relationship and its activity can be dated at Early and/or Middle Lutetian. The total amount of shortening has been estimated at ~ 16.5 km, which represents ~ 49% of the undeformed length. The Lakhoura thrust (D1a, D1c) accounts for ~ 3.5 km of the total amount of shortening. Approximately 8.0 km of shortening can be attributed to the Larra thrust (D1b), whereas the remaining 5.0 km of shortening can be related to the Gavarnie thrust (D2).

Table of Contents

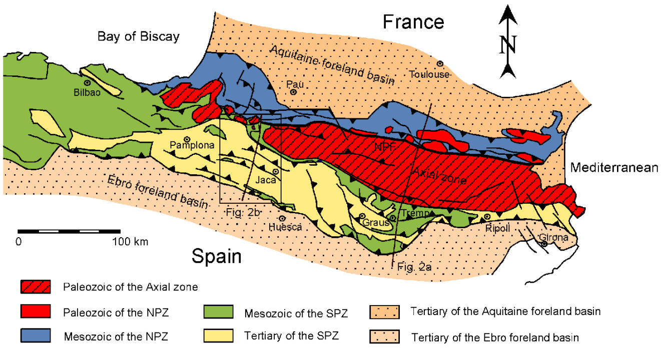
Introduction

The Pyrenees are a mountain belt of Alpine age, situated at the boundary between the Iberian and European plates, and form the westernmost extension of the Alpine-Himalayan chain. They extend from the Cantabrian platform in the west to the Provence in the east and are about 1500 km long with an average width of about 200 km. The Pyrenees formed due to collision of the European and the Iberian plates. This has been realised by the partial subduction of the Iberian plate underneath the European plate [Choukroune et al. 1990], which has been documented by seismic reflection data [ECORS Pyrenees team, 1988], seismic tomography [Souriau and Granet, 1995] and magnetotellurics [Ledo et al., 2000]. Deformation took place from the Late Cretaceous until the Miocene. The area of research is located in the western Central Pyrenees, some 15 km west of the termination of the axial zone. It is located in the Belagua and Minchale valleys, just south of the French border. The valleys strike ~ NNE-SSW, approximately perpendicular to the axis of the orogen. Detailed structural and sedimentological fieldwork has been conducted [e.g. Schellart, 1998] to unravel the local tectono-sedimentary history of the region. The most important results of this research are presented in this paper.

Geological setting

The Pyrenees have a fan like geometry, with north verging thrusts in the north and south verging thrusts in the south. The Pyrenees are generally divided into 5 structural zones [Vergés et al., 1995], which are, from north to south (Fig. 1):

Figure 1. Location map of the Pyrenees



Location map of the Pyrenees (modified from Teixell [1996, 1998]), showing the five structural zones of the Pyrenees. The location of two cross-sections has been plotted (see Fig. 2). The outlined area can be seen enlarged in Fig. 3. NPF, North Pyrenean Fault; NPZ, North Pyrenean Zone; SPZ, South Pyrenean Zone.

The Aquitaine molassic foreland basin which is a foreland basin, situated on the European plate. It formed by flexing under the weight of the thrusts of the North Pyrenean thrust zone, which were thrust on top of the plate from the south. The basin has been filled with tectonically undisturbed synorogenic sediments. The oldest sediments date from the Late Cretaceous.

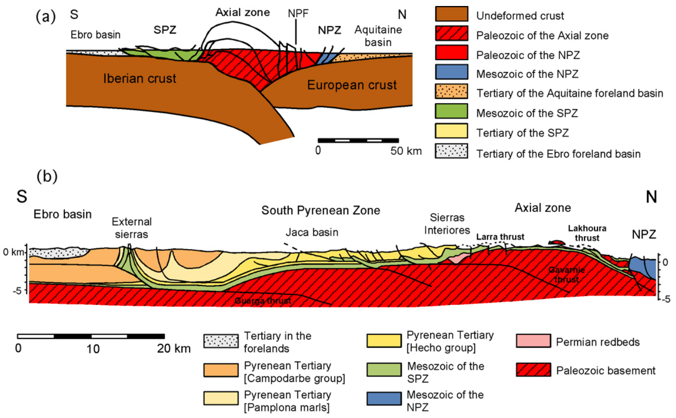
The North Pyrenean thrust zone which is a zone of imbricate north verging thrusts. The thrusts affected the Hercynian basement and the Mesozoic to Late Eocene cover sediments. This zone is bounded to the south by a narrow zone of 1-5 km in width, the North Pyrenean zone [Choukroune, 1976], which consists of highly metamorphosed rocks of HT/LP origin [Goldberg, 1987].

The axial zone which consists of Hercynian rocks and forms a south vergent antiformal stack of basement thrust sheets. The thrust sheets show typical characteristics of thick-skinned tectonics. Occasionally, the thrust sheets are covered by small remnants of Mesozoic sediments. The axial zone consists of three major thrust sheets, which are, from bottom to top, the Nogueres, Orri and Rialph thrust sheet, respectively, in the central and eastern Pyrenees. In the western part of the central Pyrenees these are the Lakhoura, Gavarnie and the Guarga thrust sheet. The rocks in the axial zone have not experienced major Alpine metamorphism.

The South Pyrenean thrust zone, which is a zone of south vergent thrusts. It deformed by thin-skinned tectonics [Muñoz et al., 1986] with a décollement level in the Triassic evaporites overlying the undeformed Hercynian basement. The décollement zone emerges at the South Pyrenean Frontal Thrust (Sierras Marginales), north of the mainly undeformed Ebro molassic foreland basin. The zone includes the deformed Paleogene foredeep (Jaca basin). Deformation of the South Pyrenean thrust zone commenced in the Early Eocene and continued throughout the Tertiary until the Early Miocene [Labaume et al., 1985].

The Ebro molassic foreland basin which formed by flexing of the Iberian lithosphere under the weight of the South Pyrenean thrust system. The Ebro basin is a molassic foreland basin, with relatively tectonically undisturbed sediments of Tertiary age, which rest directly on top of the Hercynian basement.

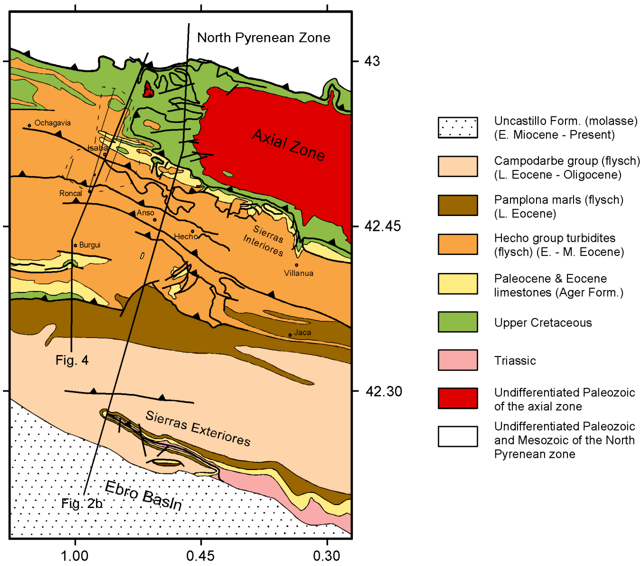
Figure 2. Cross-sections



(a) Crustal scale cross-section based on the ECORS-Pyrenees seismic profile (simplified from Munoz [1992]). See Fig. 1 for location. (b) Balanced cross-section through the western central Pyrenees from the southern margin of the North Pyrenean zone to the Ebro Basin (simplified from Teixell [1996]). See Fig. 1 and 3 for location. Note different scale between (a) and (b).

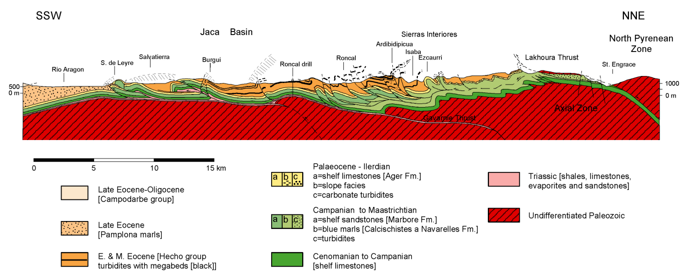
Total shortening in the Pyrenees (calculated from balanced and restored cross-sections) is estimated between 120 km [Roure et al., 1989], 147 km [Muñoz, 1992] and 165 km [Fitzgerald et al., 1999]. These estimates are mainly based on the restoration of the ECORS-Pyrenees deep reflection seismic survey [ECORS Pyrenees team, 1988, see Fig. 2a]. An estimated 125 km of shortening has been proposed by Vergés et al. [1995] for a cross-section, some 50 km to the east of the ECORS cross-section. They account 70 km of shortening to the South Pyrenean zone, ~ 23 km of shortening to the internal deformation within the basement units and ~ 32 km of shortening to the North Pyrenean zone. The difference in shortening between the north and south is probably related to the asymmetrical root of the Pyrenees, where the Iberian plate has been partially subducted underneath the European plate. The estimates mentioned above have been proposed for the central and eastern Pyrenees. However, in the western part of the central Pyrenees (some 150 km to the west of the ECORS section) total shortening is estimated to be in the order of 75-80 km (Teixell [1998]), based on mapping in the Southern Pyrenees (Anso transect [Teixell, 1996], see Fig. 2b and Fig. 3) and seismic reflection experiments in the Northern Pyrenees [Daignières et al., 1994]. This difference is the result of the greater amount of convergence between Iberia and Europe in the east compared to the west, related to the asymmetrical opening of the Bay of Biscay. This resulted in the formation of the antiformal stack of the basement thrust sheets in the east (Fig. 2a), but hinterland dipping basement duplexes in the west (Fig. 2b, 4).

Figure 3. Simplified geological map



Simplified geological map of the area around the western termination of the axial zone. The traces of two cross-sections have been plotted. The eastern one is from Teixell [1996] (see Fig. 2b) and the western one is from Labaume et al. [1985] (see Fig. 4) The study area described in this paper has been outlined by a dashed line (see Fig. 5).

Figure 4. Cross-section running through the Belagua valley and the Roncal valley



Cross-section running through the Belagua valley and the Roncal valley. Redrawn after Labaume et al. [1985]. Location of cross-section can be seen in Fig. 3. The section shows two basement faults (Lakhoura thrust and the Gavarnie thrust). From the section it is evident that in the area between Roncal and the Spanish-French border most thrusts splay of a detachment level present in the marls of Campanian to Maastrichtian age. However, several thrusts near the summit the Ardibidipicua are related to another detachment level, which is probably related to the Lakhoura thrust.

Pre-Pyrenean evolution

The oldest rocks exposed in the Pyrenees form the Hercynian basement and are exposed in the axial zone and in the North Pyrenean zone. The rocks date from Cambrian to Carboniferous age and have been affected by low- to high-grade metamorphism during the Hercynian orogeny [Vissers, 1992] in the Late Carboniferous.

During early Permian times, a first stage of sinistral strike-slip was followed by progressive extension, which led to volcanism and intrusions of granitic bodies into the Hercynian basement [Zwart, 1979]. Deposition of post-orogenic Permian redbeds occurred in small asymmetric fault bound basins as erosional products of the emerged mountains. During Triassic times, sediments were deposited in an intracontinental rift environment, with deposition of clastics, carbonates and evaporites [Puigdefàbregas and Souquet, 1986]. During Jurassic times, a widespread carbonate platform extended over most of the Pyrenees and surrounding areas [Peybernes and Souquet, 1984]. During Early Cretaceous times (Neocomian - Baremian), opening of the Central Atlantic Ocean caused a left lateral, transtensional movement of the Iberian plate with respect to the European plate, resulting in the opening of the Bay of Biscay and formation of pull-apart basins in the North Pyrenean zone [Puigdefàbregas and Souquet, 1986]. Sedimentation during this time was discontinuous and erosion was present. Marine sedimentation was restricted to these pull-apart basins. When transtension changed in left lateral strike-slip during Middle Albian to Early Cenomanian times, smaller scale lozenge to triangular shaped basins started forming [Puigdefàbregas and Souquet, 1986]. These basins were filled by the first Pyrenean flysch and narrow carbonate platforms started forming. The tectonic setting changed to transpression during the following phase. During Cenomanian-Middle Santonian times, a global sea-level rise resulted in basin widening, which led to the retreat of the carbonate shelves, while carbonate turbidite sedimentation continued in the deeper parts of the basins [Peybernes and Souquet, 1984]. During Late Santonian-Maastrichtian times, a transpressional tectonic system was present. This resulted in down-faulting tectonics, causing a general unconformity and creation of retreating erosional margins in the beginning of this period. The deeper deposits consisted of debris and interfingering systems of fan siliciclastics turbidites [Van Hoorn, 1970], whereas shallow water carbonates were deposited on the basin edge. These sediments were followed by a shallowing upward sequence from turbidites to shales and marls (Couche Rouge) and finally non-marine redbeds in the south (Aren and Marboré sandstones) [Puigdefàbregas and Souquet, 1986]. These sediments were deposited during the first phases of folding. The sinistral displacement, which started in the Early Cretaceous, ended in the Maastrichtian and was in the order of ~ 400 km towards the SE [Le Pichon and Sibuet, 1971; Olivet et al., 1981].

Evolution of the Southern Pyrenees foreland basin

During the transition from Mesozoic to Tertiary times, rearrangement of the Iberian and European plate configuration changed the geographic setting. The Mesozoic deep marine basins with corresponding northern and southern margins disappeared and were replaced by a northern and southern Tertiary foreland basin, with the emerging mountain chain in between. In the following section, the attention will be focused on the southern Pyrenees, because it is of main importance to the rest of the paper.

During the Paleocene, emergence of the inner part of the chain led to erosion and deposition of non-marine redbeds, which extended from the northeast to the east and all along the southern foreland from the east to the west near Pamplona. Along most of the southern Pyrenees a shallow carbonate shelf developed (Ager formation [Labaume et al., 1985]), including reef formations in the western Pyrenees. In the west, this shallow marine shelf environment changed northwards into a deeper basin, with deposition of carbonate slope facies and condensed deeper hemipelagics and carbonate turbidites [Léon-Gonzalez, 1972; Plaziat, 1975]. During this period, the first foreland basin geometry established in the east, while Mesozoic type basins still prevailed in the west.

From Eocene-Early Oligocene times, the Pyrenees developed into a real fold and thrust belt orogen [Seguret, 1972]. Successive thrust sheets formed in a piggyback sequence. During this period a complete southern foreland basin developed, which was flanked in the north by active thrust sheets and in the south by platform carbonates [Puigdefàbregas and Souquet, 1986; Barnolas et al., 1991]. A siliciclastic turbiditic trough developed in the east (Vic Basin) and in the west (Jaca Basin), while in the centre deltaic fan deposition took place (Tremp-Graus basin) (Fig. 1). During ongoing deformation and southward propagation of the active thrust sheets of the southern Pyrenees, the southern foreland basin synchronously shifted southwards, with turbiditic deposits onlapping onto the retreating southern platform carbonates [Labaume et al., 1985]. Contemporaneously, the southern foreland basin shifted westwards, due to progressive infilling of the basin from east to west. During Bartonian times, final infilling of the Jaca and Vic basins resulted from delta progradation from the active northern margin. These sediments were covered by a deltaic system prograding westward. The Tremp-Graus basin was at this time a region of erosion or non-marine sedimentation. The extensive southern carbonate platforms were no longer present during this time.

During the Oligocene, sedimentation was mainly fluvial to the south of the progressively emerging axial zone [Puigdefàbregas, 1975]. Southward migration of the basin and the deformation front continued during this period and ended during the earliest Miocene [Teixell, 1996]. Lower Miocene molasses were deposited unconformably above thrusts and folds of the southern Pyrenees deformation front and were later gently folded [Labaume et al., 1985]. From this time onwards the sedimentation shifted to the autochthonous Ebro basin.

Stratigraphy

Stratigraphy of the Alpine cover in the area of the western termination of the axial zone

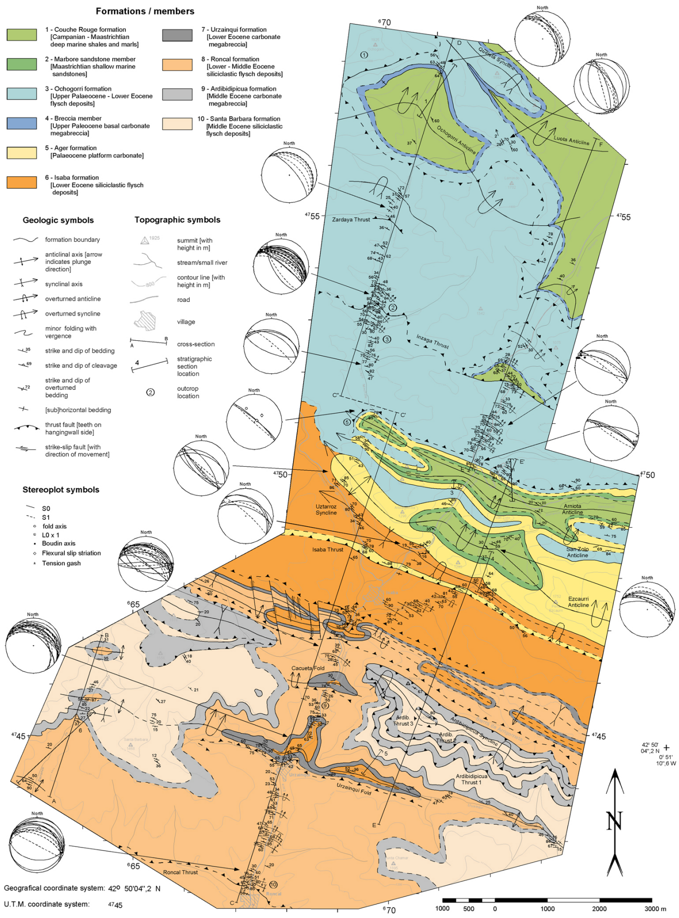
After the Hercynian orogeny, the first sediments deposited in the area of the western termination of the axial zone were discontinuous red beds of Permian age. These sediments are discordantly covered by sediments of Late Cretaceous to Late Eocene age [Teixell, 1990a]. The succession begins with Lower Cretaceous (Cenomanian to Santonian) shallow water limestones and dolomites (Calcaires des cañons [Ribis, 1965; Souquet, 1967]). The covering rocks of Santonian to Maastrichtian age consist of a calcareous shale unit (Couche Rouge), which increases in thickness towards the south. In the north, it is unconformably overlain by an onlapping turbiditic sequence, the Longibar turbidites [Teixell, 1990b], which have been dated as Late Campanian to Early Maastrichtian [Ribis, 1965]. In the south, the calcareous shales are overlain by a unit of shelf sandstones, the Marboré sandstones (Maastrichtian). In the northern domain, rocks of Tertiary age are not preserved, while in the south, the Marboré sandstones are overlain by shelf and slope carbonates of Paleocene to earliest Eocene age (Ager formation [Labaume et al., 1985]). These are followed by turbiditic sediments of Eocene age (Hecho group) which are the actual sediments of the Jaca basin.

The Jaca basin was an Eocene marine foreland basin located in the Southern Pyrenees. Today, it is exposed in the South Pyrenean thrust zone (Fig. 1). During Ypressian and Lutetian time, the basin was an east-west elongated through, filled with a thick turbiditic wedge (Hecho group), that thinned southwards [Mutti et al., 1984]. The northern border was formed by the active Lakhoura (or Eaux-Chaudes) basement thrust sheets [Labaume et al., 1985; Teixell, 1992]. To the south the trough was flanked by carbonate platforms [Puigdefàbregas and Souquet, 1986; Barnolas et al., 1991]. The turbidite wedge is 50 km wide and 4,5 km in maximum thickness [Teixell, 1992] and shows a south directed onlap [Labaume et al., 1985], which is the same direction as the advance of the thrust sheets. The sedimentary infill of the basin consists of terrigenous turbidites and thick carbonate megabreccia sheets or megabeds. The terrigenous turbidites were fed axially by deltaic shelves located to the east, on top of a thrust unit (Tremp-Graus basin) and mainly consist of siliciclastic material. The origin of the megabreccia sheets is controversial. Early papers suggested an origin in the hinterland, from hypothetical carbonate platforms (now eroded away), which were sitting on top of active thrust sheets [Mutti et al., 1984; Séguret et al., 1984; Labaume et al., 1985]. This interpretation has been based on paleocurrent directions in these megabeds and facies associations in carbonate turbidites. However, an origin in the south has also been proposed [Barnolas and Teixell, 1994]. This interpretation is based on the nature of the erosional contacts of the megabeds and the carbonate platforms in the south, the thinning and fining upward trends of some of the megabeds and the existence of the platforms in the south.

Stratigraphy of the Belagua valley

The rocks exposed in the study area are Campanian/Maastrichtian to Middle Eocene sediments. The oldest rocks are situated in the north and become progressively younger towards the south (Fig. 5). The rocks have been subdivided into eight formations. The oldest formation is the Couche Rouge formation, which consists of distal shelf calcareous shales and contains the deep-water foraminifer Navarella joaquini (Campanian to Maastrichtian) [Puigdefàbregas and Souquet, 1986]. In the north, these rocks are overlain by a series of calcareous turbiditic sediments, which mainly consist of limestone beds interbedded with shale beds (Ochogorri formation). Sometimes there are thin-bedded siliciclastic turbiditic layers present between the limestone and shale beds. These deposits have been dated at Late Paleocene age (and possibly up to Early Eocene age) and are interpreted to be the southward continuation of the Longibar turbiditic series, which are of Late Campanian to Early Maastrichtian age [Ribis, 1965]. These rocks are situated above the westward plunging axial zone antiform. In the south, the Couche Rouge formation is overlain by a platform carbonate of Paleocene age, the Ager formation [Labaume et al., 1985; Puigdefàbregas and Souquet, 1986; Teixell, 1990]. This platform carbonate shows a rapid southward increase in thickness from ~ 50 m in the north to ~ 300 m in the south. The northern (thinner) part consists of carbonate slope deposits and the southern (thicker) part consists of platform and shelf carbonates. In the south the platform carbonate is overlain by a series of siliciclastic turbiditic deposits with interbedded resedimented thick carbonate megabreccia sheets or megabeds (Hecho group) of Early to Middle Eocene age [Mutti et al., 1972; Labaume et al., 1985]. These deposits have been subdivided into five formations consisting of three packages of turbiditic material separated by two megabeds, referred to as (from old to young) the Isaba, Urzainqui, Roncal, Ardibidipicua and Santa Bárbara formation. Below, a detailed description of the formations and the separate stratigraphic columns is given.

Figure 5. Geological map of the Belague valley and the Minchale valley

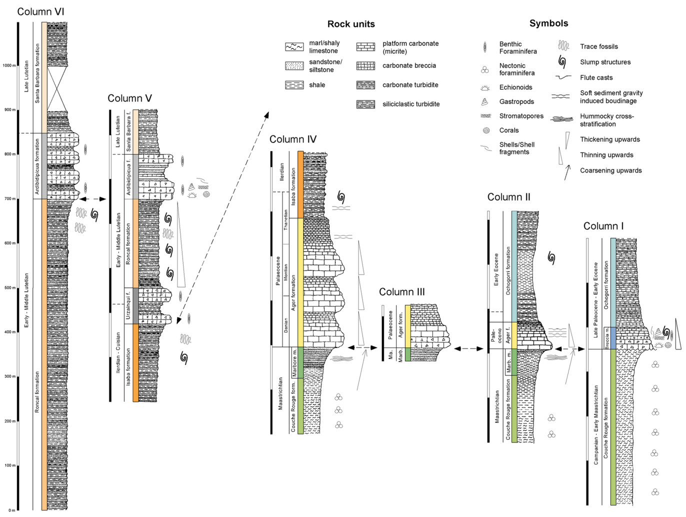


Geological map of the Belague valley and the Minchale valley, located in the Southern Pyrenees, west of the axial zone, between the village Roncal in the South and the French-Spanich border in the north (see Fig. 3 for exact location). The stratigraphic columns are plotted in Fig. 6 and the cross-sections are plotted in Fig. 7.

Correlation of the stratigraphic columns

Six correlated stratigraphic columns have been plotted in Fig. 6. The geographical positions of these columns can be seen on the geological map (Fig. 5). Below a short description of each column is presented, followed by a short section about the correlation of the columns.

Figure 6. Plots of six stratigraphic columns



Plots of six stratigraphic columns. For location of columns see Fig. 5. For description of each individual column see text.

The northernmost column (column 1) begins with a minimum of ~ 350 m of shaly marls, which contain various deepwater foraminifera (Pseudotextularia sp., Globotruncana stuarti, Globotruncana elevata, Hetrohelix sp.). The lowermost part has been dated as Campanian and the uppermost part has been dated as Early Maastrichtian. These sediments are discordantly covered by a package of ~ 50 m of limestone breccias (Breccia member), containing clasts of up to 50 cm. The member contains numerous fossils, such as foraminifera (Discocyclina seunesi, Assilina sp.), algae rodoficeas (Lithotamnium), hexacorals, stromatolites and gastropods. From determination of the foraminifera, the lowermost part of this sequence has been dated as Late Paleocene (Thanetian), indicating that there is a considerable time gap present between this formation and the underlying shaly marls. The breccia deposits are concordantly followed by at least 200 m of well-bedded carbonate turbidites with alternating limestone and shale beds. These deposits are of Late Paleocene age but could also become as young as Early Eocene. The uppermost 250 m of the column show a thinning upward trend.

The second column starts with ~ 150 m calcareous shale (Campanian-Maastrichtian), followed by some 60 m of a coarsening upwards sequence of silty shale at the bottom to siltstone at the top (Maastrichtian). These are covered by nearly 60 m of a thinning upwards sequence of micritic mudstones beds (Paleocene), with thick-bedded to massive beds at the bottom and thin-bedded beds at the top. The intermediate bedded limestones in the middle show patterns of soft sediment gravity induced boudinage (probably caused by the gravitational gliding along the slope on which they were deposited). The limestones are covered by at least 150 m of turbiditic deposits (Early Eocene), with about 10-20% of sandstone beds and a maximum bed-thickness of 10 cm.

The third column starts with some 50 m of siltstones to fine sandstones (Maastrichtian), and are followed by at least 80 m of carbonate rocks (Paleocene). The lowermost 40 m of these carbonates consist of carbonate breccias with a micrite matrix and micrite clasts of up to 1 m. These carbonates are probably slope deposits of the Paleocene carbonate platform situated more to the south.

The next column (column 4) starts with at least 125 m of shaly marls with numerous pelagic foraminifera (Globotruncana sp.), which change to calcareous shales, and finally become finegrained sandstones with hummockey cross-stratification in the last 70 m (Maastrichtian). These siliciclastic sediments are covered by some 290 m of micritic mudstones (Paleocene) in which three fining upwards sequences are present. The uppermost part of these limestones consists of thin-bedded limestones with wavy lamination, indicating a change to a deeper water depositional environment. It contains sporadic benthic foraminifera (Globigerina sp. and Globorotalia sp.) and some small fragments of echinoderms. This part has been dated as Paleocene and possibly Early Eocene. These sediments are followed by some 4 m of shales, which themselves are followed by at least 150 m of siliciclastic turbidites (Early Eocene), with about 40% sandstone beds. The transition of carbonates to turbidites indicates a rapid drowning of the platform. The turbiditic sandstone beds show patterns of soft sediment gravity induced boudinage (especially in the lowermost 50 m), indicating that they were deposited on a slope.

The fifth column (column 5) consists of turbidites and megabeds (Early to Middle Eocene). In this column two megabeds can be recognised. The megabeds predominantly consist of limestone breccia clasts of up to 1m and numerous foraminifera (Numulites sp., Assilina sp. and Discocyclina sp.) within a micrite matrix. The turbiditic deposits mainly consist of siliciclastic material with sandstone beds interbedded with shale beds. The percentage of shales varies along the column.

The last column (column 6) consists of at least 700 m of turbidites of the Roncal formation, covered by the megabed of the Ardibidipicua formation (Cuisian). This is probably covered by at least ~ 250 m of turbidites (Cuisian-Lutetian).

From correlation of the first four columns, it can be concluded that during Late Cretaceous times, the depositional environment changed from a deep water distal marine environment in the north (column 1), to slope environment (column 2), to shallow marine environment (column 3, 4) in the south. During Paleocene times, a deep marine environment remained in the north, where carbonate turbidites were deposited (column 1). Towards the south, the environment changed to platform slope (column 2, 3) and finally to platform interior, with deposition of a thick succession of carbonate mudstones (column 4). During Eocene times, turbiditic deposition shifted towards the south on top of the drowning carbonate platform (column 4). Further to the south, more turbiditic sediments were deposited, with at least two interlayered carbonate megabreccia sheets (column 5). Column 6 shows a very rapid increase in thickness of the turbiditic sediments of the Roncal formation, compared with column 5, which is interpreted to be related to synsedimentary thrusting along the Urzainqui thrust, in front of which most sediments were deposited.

Structure

Structure around the western termination of the axial zone

The study area is located around the western termination of the axial zone, which is in the Mesozoic-Cenozoic cover above the west dipping axial zone and the northernmost part of the South Pyrenean thrust zone. The basement in this area forms thrust sheets with a little degree of overlap (Fig. 2b, Fig. 4) [Laboume et al., 1985; Teixell, 1996, 1998], which is markedly different from the antiformal stack geometry in the eastern part of the Pyrenees (Fig. 2a) [Muñoz, 1992]. The dominant structural trend in the area is WNW-ESE with a SSW vergence. The deformation type is typically of upper crustal levels and shows no metamorphism. The area can be divided into four zones (Fig. 3), which are from north to south [Teixell, 1996]: (1) The Lakhoura thrust sheet, (2) the axial zone, (3) The Jaca basin (deformed Palaeogene foredeep) and (4) the thrust front of the External Sierras.

North of the axial zone, the Lakhoura thrust sheet [Teixell, 1990a; Teixell, 1996] (also named Eaux-Chaudes thrust sheet [Labaume et al., 1985]) places Hercynian basement rocks and Mesozoic cover on top of the Late Cretaceous cover of the axial zone. The thrust sheet consists of Silurian-Carboniferous rocks, covered with discontinuous sedimentary series of Mesozoic age.

The axial zone antiform consists of Hercynian rocks and a thin cover of Late Cretaceous sediments. It has a profound plunge (about 15° [Teixell, 1990a]) to the west, which affects both the Hercynian basement and the Mesozoic-Cenozoic cover. The antiform shows a flat top, which is affected by normal faults, and a homoclinal gently dipping northern and southern limb [Teixell, 1996]. These limbs are characterised by a system of hectometre-scale overturned south vergent folds, which have an associated axial planar cleavage [Teixell, 1996]. No thrusts affect the basement-cover unconformity and there are no signs of shear deformation at the unconformity [Teixell, 1996]. The sedimentary cover, however, contains numerous folds and thrusts. In the northwest of the dipping antiform, there is a series of imbricate thrusts that involve the upper part of the "Calcaires des cañons" and the overlying calcareous shales (Couche Rouge) [Teixell, 1990a]. These thrusts show typical decametre- to hectometre-scale south directed displacements. These thrusts branch from the same décollement horizon, the Larra floor thrust, which is believed to be branching of the Lakhoura basement thrust [Teixell, 1996]. The displacement on the Larra thrust in the north is transferred to the south to another set of thrusts, located in higher stratigraphic levels. Here the main décollement horizon is situated in the Upper Cretaceous marls (Couche Rouge).

South of the Axial zone lies the Jaca basin, which shows the broad geometry of an asymmetric synform. Its southern limit is defined by the External Sierras. The area can be subdivided in two regions, corresponding to two different levels of exhumation, with a northern part, dominated by the Hecho group and a southern part, dominated by the younger Campodarbe group.

The northern part of the Jaca basin shows two main stages of deformation, as found in the axial zone: an early stage, which consists of a system of gently dipping thrusts, and a second stage of deformation, defined by folds and steeper dipping thrusts and reverse faults [Soler and Puigdefàbregas, 1970; Ten Haaf et al., 1971; Labaume et al., 1985]. The first generation of thrusts shows little associated folding and is believed to be the southern continuation of the Larra thrust [Teixell, 1990], which can be observed in the Sierras Interiores, situated at the southern limb of the axial zone antiform (Fig. 2b, 3). The structure of this area is build up of a series of imbricate thrusts. The sediments involved in this thrusting are the Marboré sandstones (Maastrichtian) and the limestones and turbidites of Paleocene and Eocene age [Soler and Puigdefàbregas, 1970; Ten Haaf et al., 1971; Labaume et al., 1985]. The floor thrust in this area is located between the calcareous shale (Couche Rouge) and the overlying Marboré Sandstone and is considered to be the continuation of the Larra floor thrust [Teixell, 1990a]. The individual thrusts in this area show typical fault propagation fold geometries [Alonso and Teixell, 1992]. The second generation of thrusts is steeper and forms a large imbricate fan with intervening folds varying from metric to kilometric in size. These folds show a distinct southward vergence and may be accompanied by a slaty cleavage. The second stage structures are correlated with the antiform of the axial zone.

In the southern part of the Jaca basin, the first stage of deformation is of little importance and the second stage of folds and steep thrusts is dominant. This second stage of deformation defines the Guarga synclinorium [Puigdefàbregas, 1975]. The folds in this area are upright, with a poorly defined vergence and are of kilometric scale. The folds generally show angular geometries in the cores while the outer arcs are more rounded. The anticlines can be very tight.

The structure of the External Sierras consists of a narrow complex anticline, the Santo Domingo anticline [Almela and Rios, 1951; Nichols, 1987], whose core is disturbed by thrusts. It separates the highly deformed Jaca foreland basin with its flysch sediments from the (more or less) undeformed Ebro molassic foreland basin.

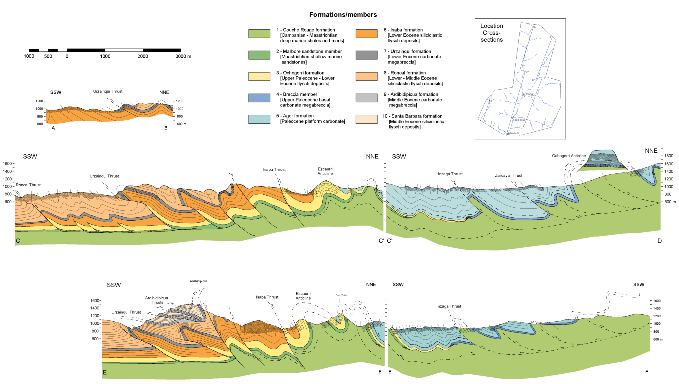
Timing of deformation

Deformation in the Southwestern Pyrenean Thrust Zone shows an in-sequence development of hinterland dipping basement thrusts [Teixell, 1996]. In general, the cover sediment thrusts developed in an in-sequence order, although out-of-sequence thrusting has also been reported in the Southern Pyrenees [Vergés and Muñoz, 1990; Burbank et al., 1992]. There are three main stages of deformation related to the three basement involved thrusts (Fig. 2b), which are from north to south: the Lakhoura (or Eaux-Chaudes) thrust, the Gavarnie thrust and the Guarga thrust [Teixell, 1996]. The first thrust that formed was the Lakhoura thrust, which resulted from inversion of the North Pyrenean basin margin, slicing off a thin piece of basement. Initiation of the Lakhoura thrust maybe started somewhere in the Late Santonian [Teixell, 1996]. Since the Larra thrust is believed to branch of the Lakhoura thrust, the age of the movement on the Larra thrust could be the age of movement of the last translation of the Lakhoura thrust as a whole. The activity of the Larra thrust has been dated from Middle/Late Lutetian to Bartonian times [Teixell, 1996]. The second basement thrust to be activated was the Gavarnie thrust, resulting in the formation of the axial zone antiform and deformation both the Lakhoura and the Larra thrust and their overlying sediments. This thrusting event was the cause of the second stage of deformation in the Jaca basin, where folds and steep thrusts dominate. The thrust was active from Priabonian to Rupelian times [Puigdefàbregas, 1975: Teixell, 1992]. The main emergence of the External Sierras can be regarded to the activation of the Guarga thrust. This stage of deformation led to the deformation of the entire Jaca basin and has been dated as Middle Rupelian to latest Oligocene - earliest Miocene [Puigdefàbregas and Soler, 1973; Pocoví et al., 1990; Hogan, 1993].

Structure of the Belagua valley

The area of investigation is situated some 15 km to the west of the termination of the axial zone. The area contains several major thrusts, reverse faults and folds. The axes of these folds and thrusts run more or less WNW-ESE (Fig. 5). These thrusts and most of the folds have a vergence towards the SSW (Fig. 7). Structural relationships at key-localities and reconstruction of the cross-sections in the area indicate that there are two main phases of deformation, the first one related to the Lakhoura basement thrust (D1) and the second one to the Gavarnie thrust (D2). The first phase can be subdivided into three subphases, the first subphase (D1a) is related to the first activity of the Lakhoura thrust, the second subphase to the Larra thrust (D1b) and the third subphase to reactivation of the Lakhoura thrust (D1c). In the following paragraph, the large-scale structures of the area will be described in reference to the geological map (Fig. 5) and three cross-sections that have been constructed (Fig. 7).

Figure 7. Cross-sections

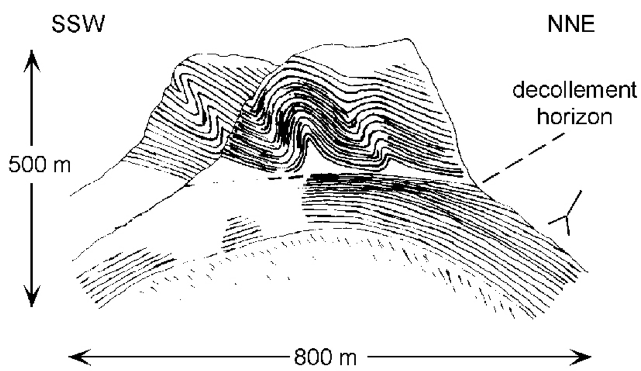


Cross-sections. For location see inset and Fig. 5.

General structures

The axial trends of the folds show a rotation of strike from NNW-SSE (320-340°) in the north to WNW-ESE (~290°) in the south (Fig. 5). The northernmost part of the area contains one syncline and two anticlines with fold axes running approximately NNW-SSE to NW-SE from which the southernmost one (Ochogorri anticline) has a relatively large wavelength (~ 2 km) and amplitude (~ 1 km). The Ochogorri anticline folds a thrust plane (Fig. 8), which is probably the southern continuation of the Lakhoura floor thrust. The décollement horizon has been interpreted to be related to an early phase of deformation (D1a), while the folding of the décollement horizon has been interpreted to be related to a later phase of deformation (D1b).

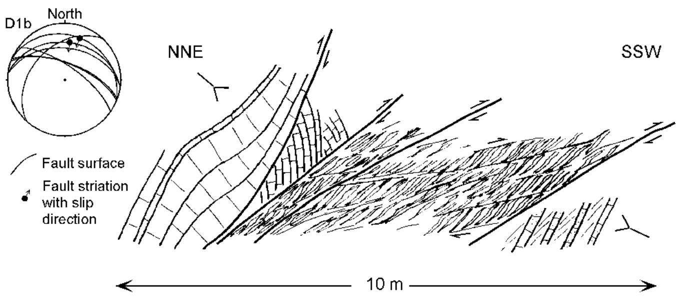
Figure 8. Sketch of outcrop 1



Sketch of outcrop 1 showing the summit the Ochogorri with marls of the Couche Rouge formation at the bottom covered by carbonate turbidites of the Ochogorri formation. The sketch shows a large anticline (D1b) (the Ochogorri anticline), which folds the sediments and a thrust plane, indicating that the thrust plane (D1a) must have been formed before the large fold. The folds above the detachment horizon point to a SSW vergence. In the Couche Rouge formation, only some cleavage is visible.

Further to the south there is a region in which the beds are dipping about 30-40° towards the NE. In this zone there is one thrust (Zardaya thrust), running more or less parallel to the bedding. This thrust could be a splay, which branches of the Larra floor thrust. This structure is followed by a zone defined by ~ 100 m scale south vergent folds and one important fault zone, the Inzaga thrust. All the kinematic indicators in the fault zone point to a top-to-the-SSW movement, in agreement with the interpretation that it is a thrust. The thrust is found in the Minchale Valley (Fig. 9) as well as in the Belagua Valley and probably originates from the Larra floor thrust.

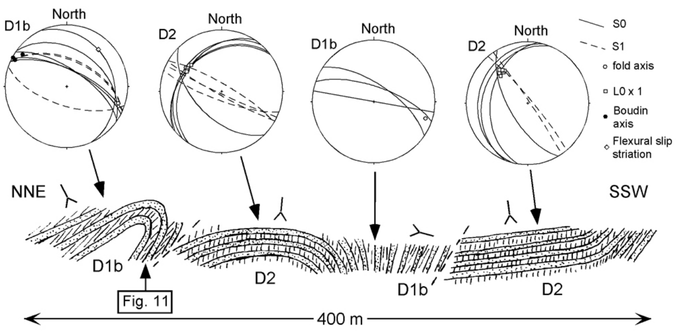
Figure 9. Sketch of outcrop 2



Sketch of outcrop 2 showing carbonate turbidites of the Ochogorri formation. The sketch shows a fault zone (Inzaga thrust) with several fault planes and in between fault breccias in which smaller scale shear zones and sigmoidal shaped shear lenses are visible. All the structures have a SSW vergence. To the north of the thrust the rocks consist of medium bedded limestones while to the south of the thrust the rocks consist of sandy limestone beds, interbedded with shale beds (at least 50%). The orientation of several faults is plotted in the stereonet. Fault striations and kinematic indicators point to a top to the SSW movement.

South of the Inzaga thrust there is a zone with two 100 m scale south vergent overturned folds (D1b), which are overprinted by a younger deformation event with associated upright folds and vertical cleavage (Fig. 10) (D2). These younger folds have a steep plunge towards the NW of up to 30¼.

Figure 10. Sketch of outcrop 3



Sketch of outcrop 3 showing medium bedded sandy carbonate turbidites of the Ochogorri formation. Four different zones can be distinguished with each zone belonging to one of two deformation phases. D1b structures have a SSW vergence with overturned folding. D2 structures are characterised by upright folds and associated vertical cleavage with relatively steep fold axes plunging towards the WNW to NW (see stereoplots).

Figure 11. Overturned turbidites with boudin structures

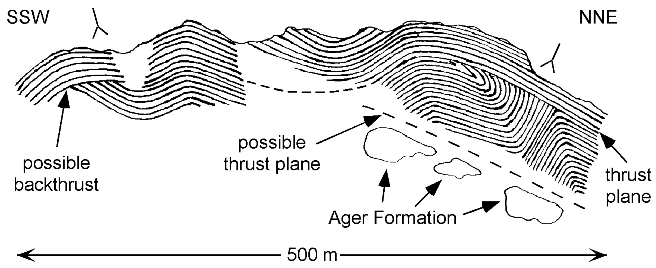
Overturned turbidites with boudin structures from which fold axis can be deduced. Boudins mainly occur in the thicker sandstone beds. Scale is indicated by geological hammer. For location see Fig. 10.

Further to the south the Ager formation comes to the surface where it is folded into three major anticlines (D2). The northernmost two anticlines are slightly overturned to upright and have a wavelength of some 400 m (the Arniota and the San Zolo anticline). The Arniota anticline folds a thrust plane (San Juan thrust), which is probably the southern continuation of the Lakhoura floor thrust (D1a) (Fig. 12). The southernmost anticline is a south vergent overturned fold (Ezcaurri anticline) with a wavelength of some 1200 m, a fold-axis running WNW-ESE and a ~ 10° plunge towards the WNW. The Ezcaurri anticline shows a great lateral continuation towards the ESE of more than 10 km.

The Uztarroz syncline, located south of the Ezcaurri anticline, also plunges towards the WNW.

At the southern flank of the Uztarroz syncline, a south vergent reverse fault crops out (Isaba Thrust, D2), which runs more or less parallel to the axis of the folds. Here, the Ager formation reappears and is thrust over folded Eocene turbidites of the Hecho group. These folded turbidites have wavelengths of 10 - 100 m, with southward verging folds in the north and symmetrical to slightly northward verging folds near Isaba.

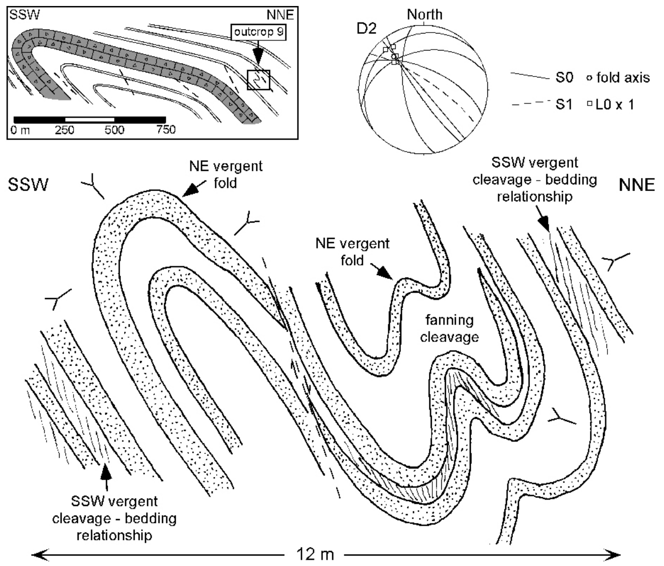
Figure 12. Sketch of outcrop 5



Sketch of outcrop 5 showing turbidites of the Ochogorri formation and limestones of the Ager formation. The sketch shows a SSW vergent thrust plane indicated by drag of the bedding along the thrust plane. Another possible SSW vergent thrust plane and a possible backthrust are also visible. The thrust planes (D1a) and bedding planes have later been folded into an open anticline (D2) (Arniota anticline).

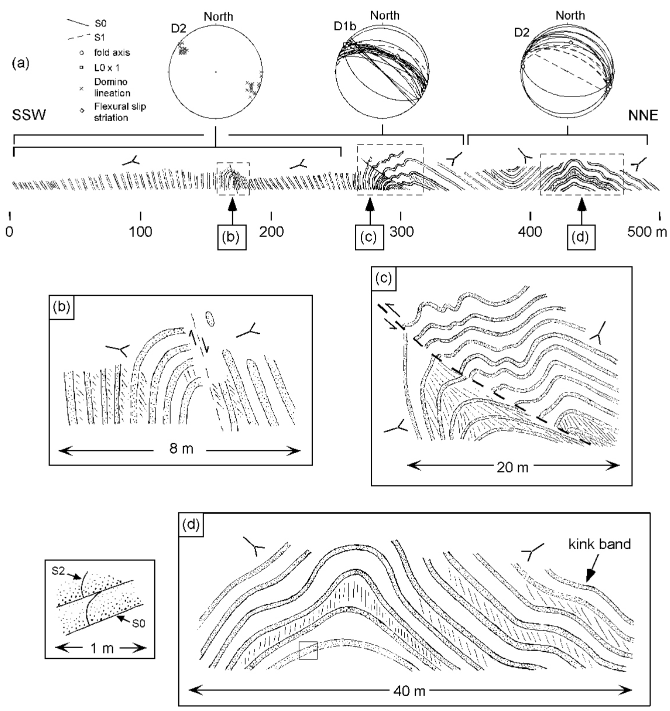
Further to the south, between Isaba and Urzainqui, the structure is dominated by two folds, which also display a great lateral continuation. The first one is an overturned syncline (Ardibidipicua Syncline, D2), which strikes ~ WNW-ESE. The southernmost one is the Urzainqui fault propagation fold (D1b). In the zone between this syncline and fault propagation fold, there is a stack of several thrust slices, with lengths of a few km and thicknesses of a few hundred metres. The thrusts slices have a vergence towards the SSW. The uppermost three thrust slices are probably the southward continuation of the Lakhoura thrust (D1c), whereas the Urzainqui fault propagation fold is probably the southern continuation of the Larra thrust. On the northern flank of the south vergent Urzainqui fold (D1b), younger north vergent structures have been recognised (D2) (Fig. 13). These structures show NE vergent detail folds, which fold the S1 cleavage. The folds plunge towards the NW at ~ 15-25¼.

Figure 13. Sketch of road section (outcrop 9)



Sketch of road section (outcrop 9) showing turbidites of the Roncal formation. The sketch shows NE vergent detail folds (D2) in between beds with a SSW vergence (D1b). These detail folds sometimes fold the cleavage related to the SSW vergent structures. The D2 fold axes display a distinct plunge towards the NW (see stereoplot). Location of outcrop with respect to Urzainqui fault propagation fold is shown in inset.

Figure 14. Sketch of outcrop 10

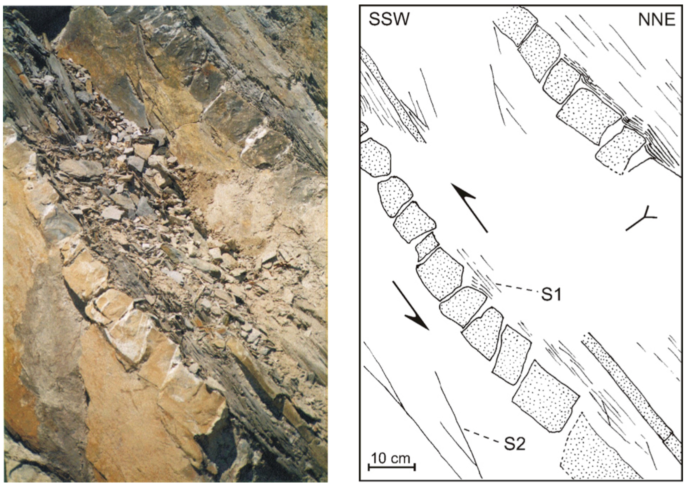


(a) Sketch of outcrop 10 showing a 500 m long road section located north of the vilage of Roncal with deformed turbidites of the Roncal formation. The section shows an open anticline and syncline in the NNE (D2) and a fault propagation fold in the middle (D1b). The part, south of the fault propagation fold contains numerous domino structures in the sandstone beds (see Fig. 15). (b) North vergent cut through detail fold with overturned beds at both sides of the fault (determined from stratigraphic markers and relationship between sedimentary bedding and cleavage. (c) South vergent fault popagation fold which runs ~ parallel to the beds in the lower right corner, forming flats in both hangingwall and footwall. In the middle and upper left corner the thrust cuts through both hangingwall and footwall, forming ramps. Here the beds show drag along the fault. (d) Roughly symmetrical open anticline with vertical cleavage and meter scale kinkbands, which seem to dominate in the northern part. The small inset shows refraction of the cleavage in some of the thicker beds, rotating more parallel to the bedding planes with increasing grain size.

To the south of the Urzainqui fault propagation fold, there are numerous short wavelength folds, with wavelengths varying between a few and several hundred metres. They strike ~ WNW-ESE and have a SSW vergence. North of the village Roncal, two open upright folds (D2) strike WNW-ESE (Fig. 14). The folds have an associated vertical cleavage and meter scale kink bands. Occasionally, the cleavage shows refraction in some of the thicker beds, turning more parallel to the beds with increasing grain size (Fig. 14d, inset). Just south of these folds there is a south vergent fault propagation fold (Roncal thrust) with a similar trending axis (D1b) (Fig. 14).

Figure 15 shows domino structures, found in sandstone beds in the southern part of Fig. 14. The thick sandstone beds show domino structures, which have resulted from sinistral shear. For this reason, they are not related to the fault propagation fold of the first deformation phase (D1), because flexural-slip along the individual beds during formation of the fault propagation fold would impose a dextral sense of shear on the southern side of this anticline. The oldest cleavage (S1) shows that it has been deformed by the rotating blocks of the sandstone beds. A younger cleavage can also be observed (S2). Therefore, the domino structures are most likely related to the symmetrical syncline located towards the north (D2), which would impose a sinistral shear flexural slip on its southern flank.

Figure 15. Domino structures



(a) Photo and (b) interpretation of overturned turbidites of the Roncal formation. The picture shows domino structures in sandstone beds. The dominos deform an older cleavage (S1). There is also a younger cleavage present (S2). These structures point to sinistral shear along the bedding planes, indicating that they are related to the symmetrical syncline in Fig. 14a and not to the fault propagation fold, which would impose dextral shear on the sandstone beds.

Relation between sedimentation and tectonics

The Roncal formation (turbiditic flysch deposits) shows a great change in thickness across the Urzainqui fault propagation fold (Fig. 6). Its thickness increases from some 200 m in the hangingwall to at least 700 m in the footwall (Fig. 7, cross-section A-B and E-F). This increase in thickness can probably be attributed to syntectonic sedimentation, where the emergence of the thrust sheets in the north led to a decrease in accommodation space above the thrust sheet. This has subsequently led to more abundant sedimentation south of the Urzainqui thrust. The Roncal formation was deposited during the Middle Lutetian [Puigdefàbregas and Sánchez Carpintero, 1978a,b], indicating a similar age for the Urzainqui thrust. The activity of the Urzainqui thrust probably seized after deposition of the Ardibidipicua formation, since it is also deformed by the underlying structure (Fig. 5, 7), but to a lesser degree compared to deformation of the Urzainqui formation. Since the Larra thrust was probably active during Middle-Late Lutetian to Bartonian times [Montes, 1992] and the Gavarnie thrust during Priabonian to Rupelian times [Puigdefàbregas, 1975; Teixell, 1992], the Urzainqui thrust is most likely connected to the Larra floor thrust.

The timing of this syntectonic sedimentation and activity of the Urzainqui thrust has a profound influence on the timing of the thrusting activity of the Ardibidipicua thrusts. These thrusts cut through the Ardibidipicua formation, which has been deposited on top of the Roncal formation, and are therefore younger than the Roncal formation. This leads to the conclusion that the Ardibidipicua thrusts have to be younger than the Urzainqui thrust. Furthermore, the Urzainqui thrust has been related to the Larra thrust [Labaume et al., 1985] and the Ardibidipicua thrusts have been related to the Lakhoura thrust [Labaume et al., 1985], where the Larra thrust is thought to be younger than the Lakhoura thrust [Teixell, 1996]. Therefore it can be concluded that the Lakhoura thrust has been reactivated after thrusting activity of the Urzainqui thrust and during or after thrusting activity of the Larra thrust as a whole. Such out-of-sequence thrusting in the Southern Pyrenees has been reported before [Vergés and Muñoz, 1990; Burbank et al., 1992] and would result in an increase of the taper of the South Pyrenean Thrust Zone.

Phases of deformation

Based on the general structure of the study area and several key outcrops it is concluded that the region has been deformed during two phases of deformation (D1 an D2), which are both related to shortening and share a similar orientation of maximum shortening (~ NNE-SSW). D1 can be subdivided into three sub-phases (D1a, D1b and D1c) and shows ramp-flat style and fold propagation style of deformation. This phase is interpreted to be related to the Lakhoura basement thrust, which splays of into the Lakhoura thrust (D1a, D1c) and the Larra thrust (D1b). The first activity is related to the Lakhoura thrust (D1a), the second to the Larra thrust (D1b) and the last activity (D1c) is related to the reactivation of the Lakhoura thrust. The deformation style of the last phase (D2) shows symmetrical to asymmetrical upright to overturned folds and steep reverse faults. Most of the folds are south vergent. The folds of this deformation phase plunge towards the WNW, where the plunge gradually decreases southwards from ~ 30¼ to ~ 0¼ (e.g. Figs. 10, 13, 14). D2 is interpreted to be related to the Gavarnie basement thrust, which resulted in the formation of the axial zone antiform and plunges towards the WNW. This could explain the decrease in plunge of the fold axes of D2 from north to south, where the northernmost part of the area is located centrally above the Axial zone, while the southernmost part is located south of the Axial zone (see Fig. 4).

From cross-section E-F (Fig. 7) and relations between thrusting and sedimentation it can be concluded that out-of-sequence thrusting took place when the Lakhoura thrust was reactivated after thrusting along the Larra thrust took place (e.g. Urzainqui fault propagation fold). This reactivation phase (D1c) resulted in the formation of the Ardibidipicua thrusts. Since these thrust sheets are folded in the Ardibidipicua syncline (D2), it can be concluded that this reactivation is older than D2.

Restored cross-section

The amount of shortening in the area of investigation has been estimated by balancing and restoring the two main cross-sections (Fig. 7, 16). However, this is a rough estimate since the stratigraphic control is not optimal in several parts of the area where only turbiditic sediments are exposed (i.e. between the Ochogorri anticline and the Arniota anticline, and south of the Urzainqui thrust).

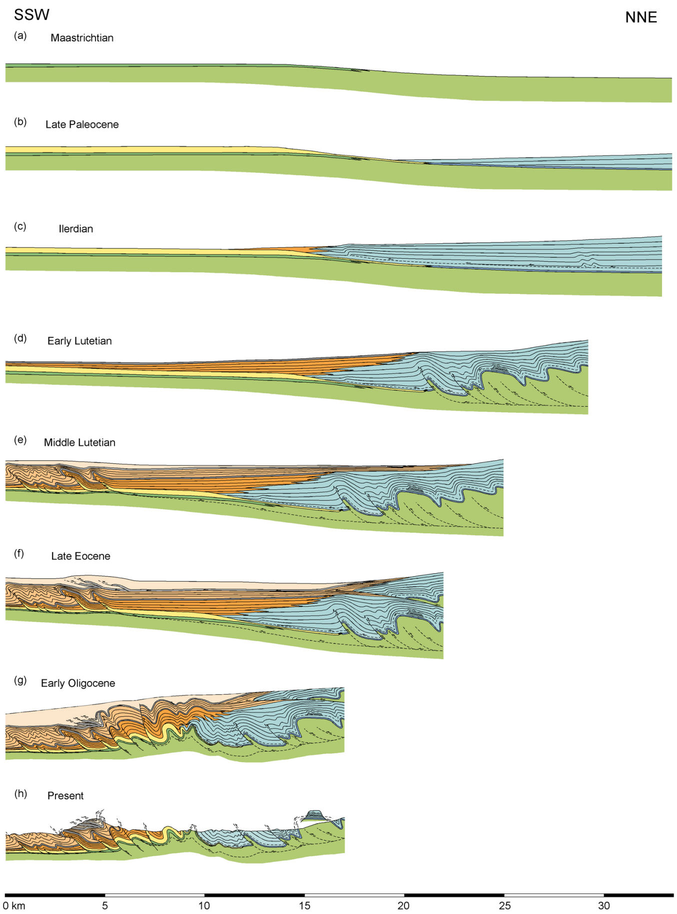
Figure 16. Schematic structural and sedimentological evolution of the Belagua-Minchale valley area, southwest central Pyrenees. The stratigraphic record in this area dates from Campanian-Maastrichtian to Late Lutetian. Deformation started somewhere in the Paleocene (c) (first activity of the Lakhoura thrust) and ended in the Early Oligocene (g) (Gavarnie thrust). For description of the individual figures see text. (Click for enlargement)

The total amount of shortening for this part of the Southern Pyrenees has been estimated at ~ 16.5 km. The total length of the present day section is some 17 km. Therefore the amount of shortening represents some 49% of the undeformed length of the section, which would have an original length of ~ 33.5 km. The total amount of shortening can be subdivided into three components, related to the Lakhoura, Larra and Gavarnie thrusts. The Lakhoura thrust (D1a, D1c) accounts for ~ 3.5 km of the total amount of shortening. This estimate has mainly been based on the minimum amount of displacement of the Ardibidipicua I, II and III thrusts (D1c). The Larra thrust (D1b) accounts for ~ 8.0 km of the total amount of shortening. The Urzainqui, Cacueta and Inzaga thrusts contribute to most of the shortening. The Gavarnie thrust (D2) accounts for ~ 5.0 km of the total amount of shortening. Main contributions to this total amount originate from the Arniota, San Zolo and Ezcaurri anticlines and the Ardibidipicua syncline.

Structural and sedimentological evolution

The area of research shows a strong relation between sedimentation and tectonics with the development of a southward shifting foreland basin in the south bordered by south vergent thrust sheets in the north. The geological development of the area has been schematically plotted in

Figure 16. Additional figure



The description below refers to this figure.

During Maastrichtian times (Fig. 16a), pelagic sediments were deposited in the north (Couche Rouge formation) while shallow marine sedimentation of sand and silt took place in the south (Marboré sandstone member). This was followed in the Paleocene (Fig. 16b) by the development of a carbonate platform in the south (Ager formation) on top of the Maastrichtian shallow marine sediments, while slope and deep marine (carbonaceous) turbiditic sedimentation took place to the north, prograding towards the south. During Ilerdian times (Early Eocene) (Fig. 16c), southward thrusting of the Lakhoura thrust (D1a) led to the initiation of emergence of the thrust belt in the north, drowning of the carbonate platform in the south and continuing sedimentation of turbidites towards the south. In the Early Lutetian (earliest Middle Eocene) (Fig. 16d), thrusting activity of the Larra thrust (D1b) led to further emergence of the thrust belt in the north and continuing southward shifting of the foreland basin with deposition of more siliceous turbiditic deposits. During Middle Lutetian times (middle Middle Eocene) (Fig. 16e), continued development of thrust sheets successively towards the south led to a shift in the depocentre south of the Urzainqui thrust with sedimentation of the turbidites of the Roncal formation. When the thrusting activity the Larra thrust nearly ceased, deposition of the second megabed (Ardibidipicua formation) took place. Reactivation of the Lakhoura thrust (D1c) in the Late Eocene (Fig. 16f) led to the formation of the Ardibidipicua thrusts and further increase of topography in the hinterland. Finally, during the Early Oligocene (Fig. 16g), activation of the Gavarnie thrust (D2) resulted in folding and steep reverse faulting in the central part of the area. Also, the northern part of the section was tilted towards the south, resulting from the formation of the axial zone antiform. Sub-aerial exposure followed, which led to erosion of the area (Fig. 16h).

Discussion

Comparison of this work with the geological maps of Ochagavia and Navascues [Puigdefàbregas and Sánchez Carpintero, 1978a,b] reveals several differences. The Minchale valley cross-section of Puigdefàbregas and Sánchez Carpintero [1978a] shows south vergent folding with a 500 m wavelength just south of the Ochogorri anticline, while in the field a continuous zone of some 2 km has been observed with north dipping beds without indication of any folding. Also, the Inzaga thrust has not been mapped in the Minchale valley by Puigdefàbregas and Sánchez Carpintero [1978a]. Another important difference, is the reappearance of the Paleocene platform carbonate south of the Ezcaurri anticline, which has not been reported by Puigdefàbregas and Sánchez Carpintero [1978a]. The work presented in here shows stacking of several thrust sheets between the village of Isaba and Urzainqui, while the geological map and cross section III and IV of Ochagavia show only two thrusts [Puigdefàbregas and Sánchez Carpintero, 1978a]. Although the presence of several megabeds superposed on top of each other can be recognised on the Ochagavia map, the sequence of megabeds is interpreted to be of sedimentary origin (i.e. sequence of different megabeds) [Puigdefàbregas and Sánchez Carpintero, 1978a].

When looking at the work of Labaume et al. [1985], there more similarities to be discovered. First of all, the style of interpretation is much more alike, with clear defined low angle thrusts and décollement levels (see Fig. 4). There is, however, one important difference. Labaume et al. [1985] have interpreted the first massive carbonate unit to the south of the Ezcaurri anticline (see megabed 1 in Fig. 4 between Isaba and Ezcaurri) to be the first megabed of the Hecho group and be located on the southern flank of the Ezcaurri anticline. In the work presented here, this outcropping formation has been interpreted as the reappearance of the Paleocene carbonate platform (Ager formation). Numerous field measurements have indicated that a tight syncline is present between this outcropping formation and the Ezcaurri anticline in the north, which gradually widens towards the WNW, due to the plunge of its fold axis towards the WNW.

It is possible to correlate some of the major structures mapped by Teixell and García-Sansegundo [1994] , Teixell et al. [1994] and Teixell [1996] in the Anso valley with structures in the Belagua valley to the west. Several folds in the Sierras Interiores, such as the Arniota, San Zolo and Ezcaurri anticlines, and the Ardibidipicua syncline can be correlated with folds in the Anso valley. Also, the Urzainqui, Ardibidipicua and Roncal thrusts can be correlated with structures in the Anso valley. However, some structures in the east are not outcropping in the west, due to the westward plunge of the axial zone of ~ 15°. For example, the décollement level of the Larra floor thrust has not been observed in the Belagua region.

Conclusions

The geological map and the structural cross-sections presented here show the northern part of the South Pyrenean fold and thrust belt located west of the termination of the axial zone. The structures show a ~ SSW vergence, with a rotation of the structural axes from NNW-SSE in the north, to WNW-ESE in the south. The sediments in this area have been deposited syntectonically and have been deformed during two main phases of deformation.

The first phase of deformation (D1) is related to the Lakhoura basement thrust, which separates the North Pyrenean zone in the north and the axial zone in the south. This deformation phase can be subdivided into 3 sub-phases, the first one related to the Lakhoura thrust (D1a), the second to the Larra thrust (D1b) and the third to the reactivation of the Lakhoura thrust (D1c). Both the Lakhoura thrust and the Larra thrust branch of the Lakhoura basement thrust. The Lakhoura thrust has a décollement level in the lower part of the Ochogorri formation from which a few thrusts branch of in this formation. The Ardibidipicua thrusts are the result of the reactivation of the Lakhoura thrust (D1c). The Larra thrust has a décollement level in the incompetent sediments of the Couche Rouge formation. An important structure is the Urzainqui fault propagation fold, which shows synsedimentary activity during deposition of the Roncal turbidites. It forms several other thrusts, where the most important ones are the Zardaya, Inzaga and Cacueta thrusts.

The second phase (D2) is related to the formation of the axial zone, which resulted from thrusting along the Gavarnie basement thrust. In this area, the structures related to this thrust shows some upright to overturned folding and related steep reverse faulting in the Mesozoic-Cenozoic cover sediments of the Hercynian basement. The most important structures are the Arniota, San Zolo and Ezcaurri anticlines and the Ardibidipicua syncline. The fold axes related to this phase plunge towards the WNW, where the plunge decreases from ~ 30° in the north (located above the axial zone) to ~ 0° in the south (located south of the axial zone).

A minimum amount of shortening of some 16.5 km can be derived from the constructed cross-sections. About 3.5 km of this total amount can be attributed to the Lakhoura thrust (D1a and D1c) Some 8.0 km of the total amount of shortening can be attributed to the Larra thrust (D1b). The remaining 5.0 km can be attributed to the Gavarnie thrust (D2).

The sedimentation of this area of research is strongly influenced by the tectonic activity with the development of a foreland basin in the south and the formation of thrust sheets in the north. During evolution of the basin, progressive thrusting from the north resulted in the shifting of the foreland basin towards the south, where turbiditic sediments of Eocene age show onlapping geometries towards the south over a drowned Paleocene carbonate platform. The work suits with the general ideas about the development of the paleo-foredeep of the southern Pyrenees, its syntectonic sedimentation and its structural development into becoming a foreland fold and thrust belt. However, although most thrusts probably developed in an in-sequence order, evidence suggests that some thrusts

(Ardibidipicua thrusts) developed during out-of-sequence thrusting.

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