Volume n° 2 - from B16 to B33

Field Trip Guide Book - B16

32nd INTERNATIONAL GEOLOGICAL CONGRESS

WESTERN PYRENEES FOLD-AND-THRUST-BELT: GEODYNAMICS, SEDIMENTATION AND PLATE BOUNDARY RECONSTRUCTION FROM RIFTING TO INVERSION

Leaders:
R. Bourrouilh, L. Moen-Maurel,
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Florence - Italy
August 20-28, 2004

Pre-Congress **B16**

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Front Cover:

1. Thrusts and Folds structures of Les Eaux Chaudes in the Ossau Valley, France. DAY 2, Stop 2.5b.

2. Thrusts and Folds structures of the External Sierras, Spain. DAY 5, Stop 5.3.

3. Thrusts and Folds of Gavarnie at La Estiba, Spain. DAY 6, Stop 6.2.

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Introduction

This field trip along a geotraverse through the north-western and south-western Pyrenees area has three main purposes:

1. To analyse the present-day structures in the fold and thrust belt and to reconstruct their structural evolution in order to better understand the subsurface foreland structures (some of them being oil/gas traps or prospects).
2. To describe the stratigraphic evolution of the Mesozoic and Tertiary formations (with respect to plate tectonics and to the Iberian plate drift) on several transects: along the Ossau Valley, the Aspe Valley, the Mendibelza and Orhy Massifs for the northern side, and from the Orhy Massif to Jaca and Aïnsa for the southern side.
3. To give petroleum geoscientists the recent concepts and keys for the Pyrenean foreland and foothills hydrocarbon exploration.

The scenic landscapes of the Pyrenees mountain range provide an exceptional laboratory for examining the orogenic processes, from rifting to collision, both in the crustal lithosphere as well as in the foreland basins.

A comparison to more classical fold-and-thrust belts will be proposed along the way and in the wrap-up conclusion of the field trip.

Field References:

Topographic and Road maps : for example:
Michelin n°234 Aquitaine 1/ 200,000
Michelin n°573, Regional 1/400,000:
Pais Vasco/Euskadi, Navarra, La Rioja.
Michelin n°574, Regional:1/400,000:
Aragon, Cataluña/Catalunya

Geological maps :
S.N.P.A. (Société Nationale des Pétroles d' Aquitaine) (1972). Carte géologique des Pyrénées, scale 1:250,000, P. Soler ed. 4 plates.
B.R.G.M. (1980). Carte Tectonique de la France, scale 1:1,000,000, B.R.G.M. Ed.
B.R.G.M.: 1/50,000 geological Maps: n° XIII-46, St Jean Pied de Port; XIV-46, Tardets-Sorholus; XIV-47, Larrau ; XV-45, Pau; XV-46, Oloron Sainte Marie; XV-47, Laruns (under press).
Instituto Geologico y Minero de Espana, Geological Maps at 1/ 250,000 and at 1/50,000: n°118 Zuriza, n° 144 Anso, n°145 Sallent.

General geological setting (Plate 1, fig.1 and 2).

Introduction

Situated between the European plate and the Iberian plate, the Pyrenean orogenic belt extends from the NW part of Spain or Galicia, to the Southern part of the Alps and the Gulf of Genoa, in the Mediterranean. It is thus a belt almost 2,000 km long, of which only the central part, or the Pyrenees, is a visible collisional orogen, whereas both the Western part or the Bay of Biscay and the Eastern part or the Gulf of Lion and Gulf of Genoa remain at subsea.

The Pyrenees resulted from an orogeny which developed over a previously thinned continental crust, but without an intervening oceanic crust between the two diverging plates. Underplating of the Iberian plate underneath the European plate led to the sinking of a deep crustal root along a north-dipping plane while coeval delamination of the upper crust led to the propagation of a major fold-and-thrust belt to the south as well as of a conjugate north-verging back-thrust system. Thus the range is characterized by a doubly-verging asymmetric thrust stack wedge:

- the southern fold-and-thrust belt (FTB) represents a south-verging imbricate fan affecting cover formations (Mesozoic and Cenozoic in age) rooting into an antiformal stack of basement rocks (Precambrian to Paleozoic variscan folded series) in the axial zone. The piggy-back thrust sheets indicate a crustal shortening which reaches over 100 km along, i.e. 90% of the total orogen collision;
- the northern fold-and-thrust belt is confined over a narrower zone north of the axial zone or the High Chain, in a conjugate position to the underplating process (Moen-Maurel *et al*., 1999).

The relatively minor inversion in the Northern FTB protected the rifting series and structures from uplift and erosion, thus permitting the examination of the initial processes of the orogenesis.

The genesis of the Pyrenean belt is the result of:

- 1. the break-up of the 250 Ma old Wegener's Pangea and reactivation of its structural inherited fabrics,
- 2. the correlative propagation of the Atlantic Mid-Oceanic Ridge from Central Atlantic to the North, with an Atlantic RRR triple junction along its Eastward opening branch entering Bay of Biscay,
- 3. the Late Jurassic - Late Cretaceous diachronous Eastward opening of the Bay of Biscay which

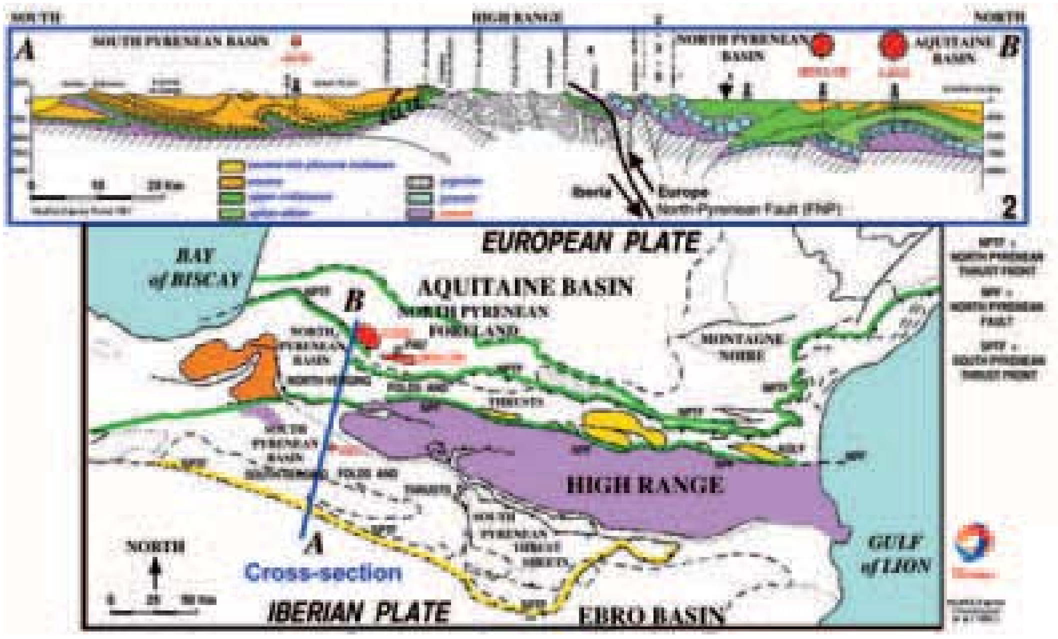
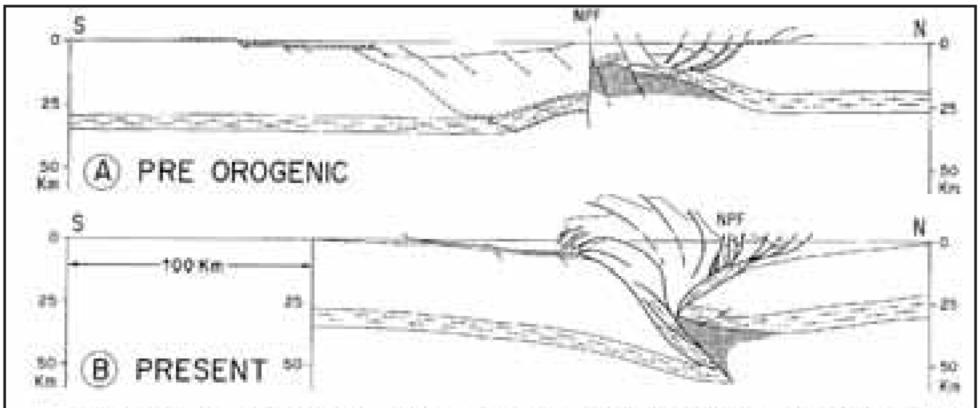


Plate 1 - 1. Structural Map of the Pyrenees with location of cross section AB. 2. Transversal cross section AB of the Pyrenees, from the Ebro Basin (South) to the Aquitaine Basin (North).

resulted in the drift and progressive rotation of the Iberian plate and in the birth of rifted interplate basins, such as the N- Pyrenean basin at the tip of the Bay of Biscay. In Turonian the Eastward Atlantic R junction point aborted (R-->r), the North-Pyrenean rift never produced oceanic crust: it can thus be called "aulacogen",

- 4. the northward push by the large African plate Stopping the free drift of the Iberian plate toward the south-east. This drift is due to the Cretaceous oceanic opening of both the South Atlantic and Indian Oceans, as well as by the closure of the Tethys, which provoked



Crustal scale balanced cross-section along the profile. The upper section (A) shows the pre-orogenic stage during the sinistral strike-slip motion of the Iberian Plate along the vertical NPF and illustrates the tensional tectonics in the North Pyrenean Zone. This major boundary is then deformed (B) by the 100 km crustal shortening. This interpretation is mainly based on the hypothesis that the south-dipping reflectors (SDB in Fig. 2) below the trace of the NPF represent the most southern part of the under-thrust European lower crust.

Figure 1 - Crustal scale balanced cross section, ECORS seismic profile, from Choukroune et al. , 1989.

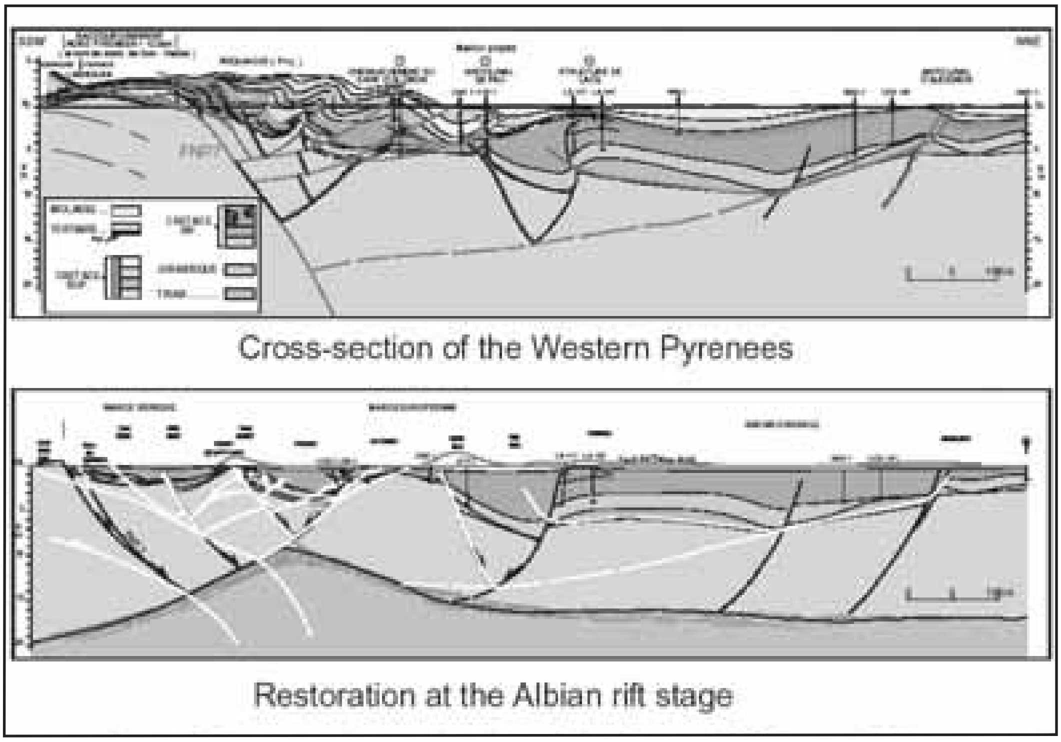


Figure 2 - Cross-section passing through the Lacq and Meillon gas fields (Moen-Maurel et al.1999).

the migration of the African plate towards the NE and then towards the N. This plate motion crushed the open space between Europe and Africa and formed the Pyrenean-Alpine orogenic belt in a diachronous compression which proceeded westwards along the northern margin of the Iberian plate.

As a result of the collision between the Iberian and European plates, which began during the Turonian and continued:

- 1. the rifted N-Pyrenean basin became a foredeep basin
- 2. a lithospheric compensation responded to the collision and the associated crustal thickening, creating a large intraplate foreland over the Iberian plate, (the South-Pyrenean basin), and a retro-foreland basin on the Northern edge of the Iberian plate and over the

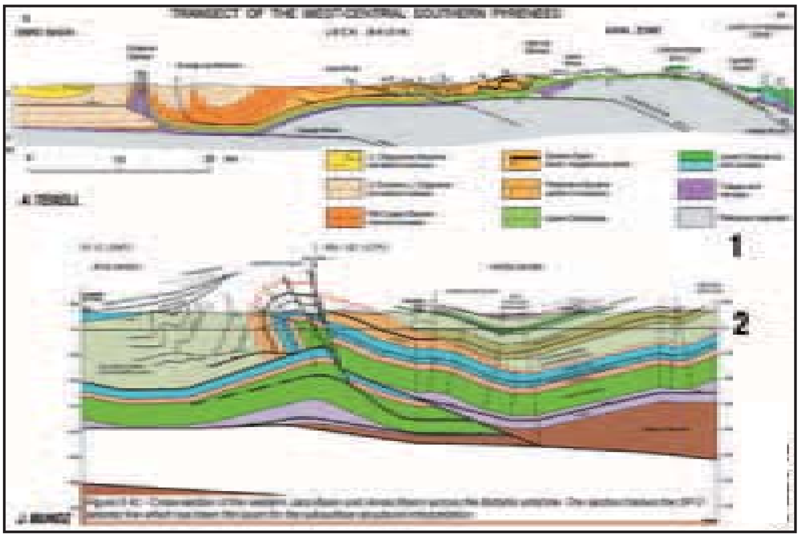


Plate 2 - 1. Transect Of The West-central Southern Pyrenees, from A. Teixell. 2. Cross-section of the Western Jaca basin and Ainsa basin across the Boltaña anticline, from J. Muñoz.

European plate (the Aquitaine basin).
- 3. the Pyrenean orogenic belt displays a fan-shaped cross-section, resulting from the crustal ductile underthrusting or underplating of the Pre-Pangean basement (Precambrian-Paleozoic) of the Iberian plate underneath the European plate. Since the initial stages of the Pyrenean collision at Late Santonian-Campanian times two foreland basins developed, one on each side of the double-wedge (Figs. 1 and 2).
- 4. The post-Pangean sedimentary infill of the interplate (north) and the intraplate (south) basins thrusted toward either the north (Aquitaine basin) and south (Ebro basin) forelands, in successive shallow fold-and-thrust piggy-back sheets.

Deformation of the Pyrenean double-wedge migrated outwards in a piggy-back manner, although synchronous hindward internal deformation has also been documented. The forward propagation of the deformation in the southern Pyrenees was modified, in the last stage of the evolution of the thrust-belt, by a break-back reactivation of the older thrusts and by the development of new, minor out-of sequence thrusts affecting syntectonic deposits (Martínez *et al*., 1988, Vergés and Muñoz, 1990). Thrust transport direction was constantly N-S to NNE-SSW through most of the tectonic evolution as deduced by the map pattern of the structures, kinematic criteria along thrust planes and the absence of significant rotation around a vertical axis along the analyzed cross-section (Dinares *et al*., 1992). This implies a near normal convergence through the main orogenic phase. The strike slip convergence vector between the two plates must be found at depth, decoupled from the cover deformation. A partitioning of the deformation must occur both in map view and with depth in favour of the activation of detachment horizons.
Although an isostatic balance occured all along the Pyrenean orogeny, between compression - erosion -sedimentation, the main apparent mountain uplift has occurred since the Pleistocene, following mainly the glaciations. It mostly affects the Iberian plate, as a result of the isostatic rebound of the crustal Pyrenean subducted root.

Structure of the Pyrenees

The Pyrenees can be considered an immature, incomplete orogen resulting from plate drift rotations which leads to varying deformations along strike. To the west, the space between the European and Iberian plates still remains open in the Bay of Biscay, the northern margin of Spain showing an incipient north-verging subduction. To the east, the Pyrenean orogenic belt extends in the Alps of Provence, exhibiting minor inversion thrusts and folds (Debelmas, 1974). More to the south, the Neogene Mediterranean extension, consecutive to the Betico-Rifean compressional events (Bourrouilh and Gorsline, 1979) overprints the Pyrenean belt.
Between these belt-tip areas, where the European and Iberian plates are in full contact, i.e. in the Pyrenees *sensu stricto*, the collision directly affected the continental lithosphere, since no oceanic floor was present. Using an inherited Precambrian (?) - Variscan fault network, the collision provoked an inversion of the deep fault network in the crustal basement, and the building of a deep-rooted fan-shaped mountain range (Fig.1 and 2, and Plate 1). In the Central Pyrenees the associated strike-slip motion that is well visible along one of the faults of the southern rift margin (North-Pyrenean Fault) marks the plate boundary. Along our geotraverse the NPF is concealed underneath a south-verging thrust (Plate 1). The North-Pyrenean rift structures were little inverted north of the rift axis, and remained north of the North-Pyrenean Thrust Front. South of its axis the tilted blocks of the rifting stage were affected by inversion and uplifted, either towards the south or the north, depending on the original dip sense of the rift faults, either sheared by wrenching, or thrust along the detachments in the Triassic evaporites and the Upper Cretaceous flysch.
The south-verging thrusting along successive shallow basal and internal plastic décollements also affected the cover of the High Chain, as well as the sedimentary infill of the South-Pyrenean intraplate basin. The plate collision resulted in the piggy-back stacking of south-verging folds and thrusts involving the basement as well as its cover.

The crustal and cover structuration is clearly observable in the field, in the Pyrenean mountain belt and foothills. It can be subdivided from north to south into different structural units (fig. 2 and Plate 1, fig.1 and 2) (Castéras, 1974):
The northern foreland basin, in the southern part of the Aquitaine basin, consists of a thick (few km) succession of Upper Cretaceous turbidites overlain by an up to 4 km thick Paleogene series. Most of the latter are represented by continental deposits as only marine platform sediments of Lower Ypresian age are observed (Buy and Rey, 1975). The Aquitaine basin mainly developed in the footwall of the North-Pyrenean Frontal Thrust and was not greatly involved

in the North-Pyrenean thrust system.
The southern part of the Aquitaine Basin is affected by folds with large curvature radii related to the Pyrenean Tertiary tectonic phase. In this area the Pyrenean compression is mainly Eocene in age. This phase enhanced salt tectonics.
The Sub-Pyrenean Zone or northern folded foreland (which lies beneath and north of the North Pyrenean Thrust Front (or NPTF) is characterized by north-verging blind thrusts and asymmetric folds often cored by a salt ridge or an inverted rift graben.
The NPTF is composed of a series of en-echelon N-verging thrust faults running from the Bay of Biscay up to the South-Western Alps.
The North Pyrenean Zone or NPZ, lies to the south of the NPTF and is characterized by north-verging thrusts and folds verging either north or south depending on the dip sense of the inverted normal faults. The southern part of this zone has been deformed by schistose deformation and low-grade Pyrenean-age metamorphism, at the contact with the NPF.
Precambrian to Paleozoic basement outcrops in the NPZ, constituting the N-Pyrenean massifs.
The North Pyrenean Fault or NPF lies to the south of these units, and appears in various ways throughout the belt. Its variable aspects take on the status of the orogenic axis of the range (Iberian-European plate boundary), the change-over zone from northern to southern vergence (Central Pyrenees), the zone of intense deformation (various schistosities), a metamorphic zone, and a lherzolite (mantle-derived ultrabasic granulitic rock) injection zone (fig. 2 and Plate 1, fig. 2).
The fault evolved from an initial Albo-Cenomanian transtensional regime with the formation of pull-apart basins and the development of a thermal metamorphism (Debroas, 1990; Goldberg and Maluski, 1988) to a later transpressional regime during the onset of convergence in Early Senonian time (Puigdefàbregas and Souquet, 1986; Debroas, 1990). Lower crustal granulitic rocks as well as ultrabasic upper mantle rocks (lherzolites) are observed embedded between the Lower Mesozoic metamorphic rocks along a narrow strip parallel to the North Pyrenean Fault (Choukroune, 1976; Vielzeuf and Kornprobst, 1984). These rocks were carried to upper crustal levels during the strike-slip faulting. Apart from this narrow strip parallel to the North Pyrenean fault neither post-Hercynian metamorphic rocks nor lower crustal rocks are observed at the surface in the Pyrenees.
The Axial Zone or High Chain (Plate 1, fig. 2 and Plate 2, fig. 1) is the highest part of the range, and represents the present-day orographic axis. It is composed of Precambrian series, possibly folded (?) during the Late Proterozoïc, and of thick Paleozoic series that were folded during the Variscan (Hercynian) orogeny and redeformed during the Upper Cretaceous to Tertiary Pyrenean tectonic phases. The High Chain basement is overlain by discontinuous Permo-Triassic deposits and by a thick Upper Cretaceous shelf carbonate series; this shelf is overlain by the edge of an Upper Cretaceous to Tertiary synorogenic flysch basin, which developed synchronously with the compression and with the tectonic subsidence and deepening of the South and North Pyrenean foredeep basins.
The basement rocks of the Axial Zone constitute an antiformal stack. This antiformal stack only involves upper crustal rocks, its floor thrust being located 15 km below the top of the basement. In the Central Pyrenees it is constituted by three main structural units: Nogueres, Orri and Rialp thrust sheets (Muñoz, 1992). The regional westward plunge of the basement antiformal stack leaves only the Nogueres – equivalent thrust sheet at outcrop level; it is then called Gavarnie - Eaux Chaudes in the Western Pyrenees. The Nogueres-Gavarnie thrust sheet is the uppermost of the antiformal stack and its southern tip is the basement of the lower cover thrust sheets. In the central Pyrenees, the contact between the cover Upper Thrust sheets and the basement antiformal stack corresponds to a passive-roof backthrust (Morreres backthrust). During the development and southward displacement of the basement antiformal stack the cover units have been wedged by delamination, and on the top of the basement, as described in other orogenic belts (Price, 1986). A mirror image of this delamination of the cover by the basement thrust sheets also exist in the North-Pyrenean foreland at the edge of the Paleozoic Basque Massifs, which make a cartographic salient and a splinter wedge into the northern foreland (Figure 1).
The South Pyrenean Zone (Plate 2, fig. 2) is characterized by South-verging thrusts and folds. The stratigraphy is essentially represented by Permo-Triassic deposits, locally overlain by Jurassic, but mostly covered by Upper Cretaceous shelf carbonates passing to Eocene turbidites (in the western part) and Oligocene molasses conglomerates
In the southern Pyrenees the cover Upper Thrust Sheets (Muñoz *et al.*, 1986) consist of Mesozoic and

syntectonic Paleogene rocks which were initially detached from the basement over the Upper Triassic evaporites. These thrust sheets were later thrusted on top of autochthonous Paleogene rocks in continuation with the Ebro foreland basin (Fig. 2). They are the Central-South Pyrenean thrust sheets (Cotiella, Boixols, Montsec and Sierras Marginales) in the central Pyrenees and the Pedraforca thrust sheets in the eastern Pyrenees (Figs. 1 and 2). The Mesozoic series is only tens of meters thick in the southernmost units and progressively thickens northwards up to 7km. This sedimentary wedge is the result of the progressive southwards thinning and pinch out of the Cretaceous stratigraphic units coupled with the geometry imposed by Cretaceous (mainly Lower Cretaceous) extensional faults. The Upper Thrust Sheets show numerous oblique and lateral structures, probably related to the original Mesozoic basin configuration. From these structures an approximately N-S transport direction can be deduced. Location of thrusts is strongly controlled by previous extensional faults, mainly Lower Cretaceous in age. Inversion tectonics is a structural feature of the Upper Thrust Sheets.

The South-Pyrenean Thrust Front or SPFT extends from the east to the north-west (Plate 1, Fig. 2) and represents the Frontal thrust of the ramps and duplexes system which affects both the High Chain and the sedimentary infill of the South-Pyrenean foreland basin.

The Ebro Basin represents the autochthonous external foreland south of the Southern Pyrenees Frontal Thrust. It is mainly filled by the last stage continental sediments of the foreland basin after the Lower Priabonian evaporites. These evaporites represent the paleogeographical closing of the Ebro basin. Upper Eocene-Lower Miocene continental clastics filled the enclosed basin and progressively backfilled and buried the south-Pyrenean thrust system during its late development stages (Coney, *et al*., 1996).

Geophysical data and crustal structure

The crustal and lithospheric structure of the Pyrenees has been investigated by different geophysical techniques (deep reflection and refraction seismic profiles, gravity, magnetotellurics, magnetic anomalies, tomography, heat flow). The data that best constrain the Pyrenean crustal structure are from the deep seismic reflection profiles, mainly the ECORS-Pyrenees profiles (Choukroune et el., 1989, Roure 1989). These profiles have been interpreted to show the subduction of the Iberian plate below the European one and has been the basis for the construction of crustal balanced cross-sections (Roure *et al*., 1989, Muñoz, 1992).

Several different interpretations of the crustal structure of the Pyrenees have been given on the basis of the combined geological and geophysical data (Roure *et al*., 1989, Mattauer, 1990, Muñoz, 1992). We believe that an explanation in which the orogenic double-wedge involves only upper crustal rocks provides the best geometry in which to integrate all these data (Fig. 4). Apparently, the crust was decoupled and the lower crust, below the upper crustal double-wedge, was subducted together with the lithospheric mantle into the mantle. This inferred crustal subduction is compatible with other geophysical data such as tomographic analyses (Souriau and Granet, 1995) and by a magnetotelluric profile across the central Pyrenees (Pous *et al*.,1995).

Balanced and restored cross-sections

Balanced and restored cross-sections were constructed not only to integrate geophysical and geological data but also to estimate the amount of orogenic contraction (Fig. 1). A geometrical solution of a crustal cross-section of the central Pyrenees along the ECORS profile gave a total shortening of 147 km (Muñoz, 1992). However, this value increases up to 160 km if the internal deformation of the crust below the sole thrust of the Pyrenean thrust system is restored. Other cross-section restorations of the central Pyrenees have estimated shortening values over 100 km (Deramond *et al*., 1985; Roure *et al*., 1989). A shortening calculation for a crustal cross-section in the eastern Pyrenees yielded a shortening estimate of about 125 km (Vergés *et al*., 1995). No more than 15 km of shortening can be accounted by the north-verging structures north of the North Pyrenean Fault (Moen-Maurel *et al*., 1995). Therefore about 90% of the collision shortening is produced by the south-verging thrusts and antiformal stacks, south of the North Pyrenean Fault.

These shortening calculations are compatible with the estimated separation of the Iberian and European plates as deduced by reconstruction of the past motion of Iberia after paleomagnetic data (Roest and Srivastava, 1991). These paleomagnetic data, as well as cross-sections west of the ECORS one, show that shortening decreases westwards. Shortening values on the order of 80 km are reached westwards from

the ECORS cross-section (Grandjean, 1992, Teixell, 1996, 1998).
The estimated duration of convergence in the central Pyrenees is about 60 Ma, which gives a mean shortening rate of 2.5 mm/yr. A similar shortening rate has also been deduced in the eastern Pyrenees during a shorter period of convergence (Vergés *et al*., 1995). The latest deformation migrated westwards. It Stopped during Middle Oligocene time in the eastern Pyrenees and continued to the Middle Miocene in the westernmost Pyrenees (Vergés, 1993). In the central Pyrenees, along the ECORS cross-section, deformation ended by Early Miocene times.

Stratigraphic and geodynamic evolution

(Fig. 3) (See further on the contribution of A. Teixell and of J. Muñoz for a more detailed study of the Southern Pyrenean side).
According to Bourrouilh, Richert and Zolnaï, 1995, Richert *et al*., 1995, the stratigraphic and structural evolution of the Pyrenees can be summarized as follows:

Precambrian to Paleozoic

The evolution of the Precambrian and Paleozoic domain related to the geodynamics of the Pyrenees, Montagne Noire and Western Mediterranean has

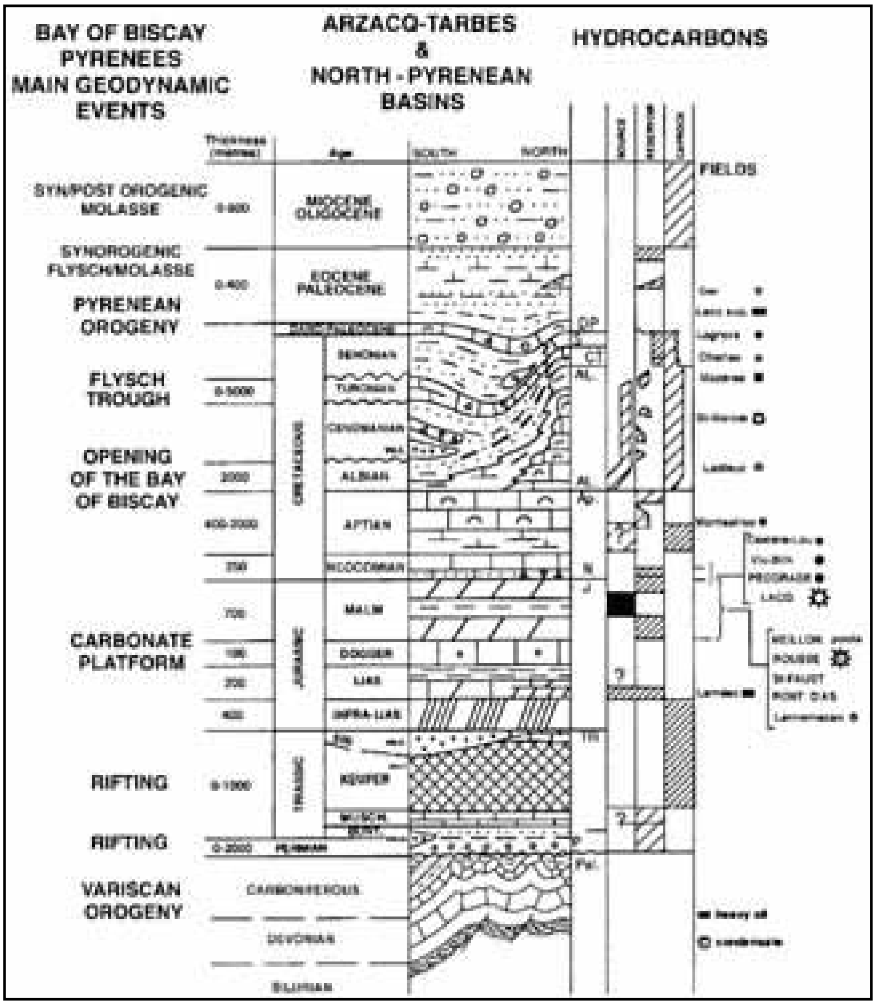


Figure 3 - Geodynamic evolution of Bay of Biscay-Pyrenees and North-Pyrenean basins, from Bourrouilh et al., 1995, modified.

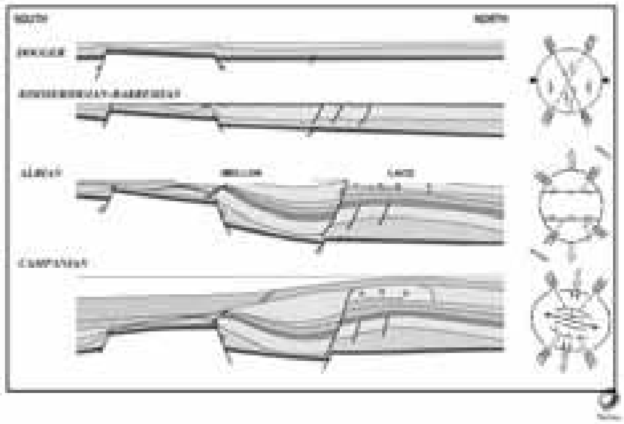


Figure 4 - Tectono-stratigraphic evolution of the Lacq-Meilllon area (Total).

been synthetized by Bourrouilh and al. (1980). The Paleozoic has been studied in the Aquitaine area by Winnock (1971), Autran and Cogné (1980), Bourrouilh and Mirouse, (1984) and others.

The structural framework of the Precambrian basement was acquired during the Late Proterozoïc (Cadomian). Generally, Precambrian series are metamorphosed (Agly massif) and radiochronologically dated (Ursuya-Baigoura massif). The Paleozoic contains six main members: a post-orogenic conglomeratic and then locally stromatolithic Cambrian (Bouquet *et al.*, 1985), a thick fine-grained silici-clastic Ordovician (Gapillou, 1981), a black-shaly Silurian interval, a limy Devonian, a sandy-shaly carboniferous section and the red Permian member. Precambrian and Paleozoic basement were then successively affected by Caledonian and Variscan (or Hercynian) orogenies. The Variscan orogeny was accompanied by large thrusts and folds and metamorphism. Upper Variscan granites locally intruded the series.

The **Permian** lies unconformably on the previously deformed series. It is composed of continental red beds, in which locally an andesitic volcanism provides volcanic debris (Lucas 1985) but also basaltic flows, as witnesses of oceanic expansion.

Triassic to Early Liassic (Fig.3)

The Triassic series are of a typical Germano-type i.e. a three-fold stratigraphy: from base to top: post orogenic red conglomerates and sandstones, marine carbonates and red shales with evaporites and volcanics. The thickest volcanic wedge showing crustal distension, has been emplaced along the southernmost rifted basin edge (over several hundreds of metres), while they taper off to zero towards the Aquitaine present-day basin center.

Jurassic

The Lower Liassic anhydrites grade upwards into a shallow-marine dolomitic, then mainly limestone platform sequence (Dogger-Malm). The Oxfordian through Early Kimmeridgian time period can be associated to a more active extensional phase leading to structural and paleogeographical differentiation on the platform (Canérot, 1987). The local Pangean platform is subdivided into two major provinces: an inner part with dolomites and limestones of the Meillon and Bayseres Formations to the east versus a more open marine environment to the west : Oxfordian Ammonite marls and Lower Kimmeridgian shaly limestones of the lower Cagnotte Formation. Following this phase of differentiation, the Middle to Late Kimmeridgian period represents a stable environment of deposition, affected only by local condensed sedimentary deposits, related to synsedimentary salt tectonics along basement faults. At the end of the Jurassic, the Neo-Kimmerian tectonic phase leads to a general regression. The dolomitic facies of the Mano Dolomite (200 m thick in average) are associated to this regressive cycle. In the Late Portlandian to Berriasian, the platform is again separated into two domains, along the

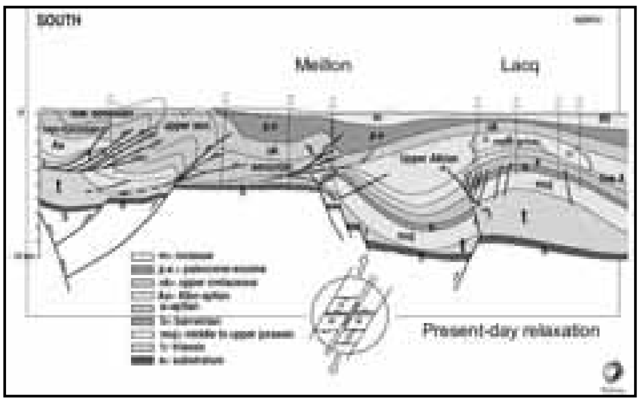


Figure 5 - Cross-section of the Lacq-Meillon area (Total).

submeridian trend defined here above. At the Jurassic-Cretaceous boundary large areas became emergent as demonstrated by the presence of anhydrite, some coal, red shale, clastics (sand), sand-rich carbonates and intra-formational (desiccation) breccias. The first salt-movements (gentle swells and salt-cushions) seem to have started as early as in the Portlandian (Winnock and Pontalier 1970). Massive sand-bodies were deposited in the Parentis Basin during the Purbeckian-Wealdian period. The equivalent series of the southern Aquitaine sub-basins are carbonates, intraformational breccias and shales (Lacq). The sedimentation over the south-western part of the basin indicates a deeper environment of deposition and a more continuous sedimentation during the Late Jurassic – Early Cretaceous transition (Hauterivian – Berriasian). Montagné's thesis (1986) evidenced rifting during Late Jurassic, south of Bayonne. Later, ODP Leg 103 (1988) found turbidites in the off-shore Hauterivian sequences in the westernmost part of the basin near Bilbao, indicating yet another opening (rifting) attempt of the Bay of Biscay.

Early Cretaceous

From Late Jurassic onward, a new paleogeography developed, provoking the formation of en-echelon diamond-shaped grabens in relation with the activation of inherited structural fabrics.

Within these basins three major phases of subsidence associated to extensional phases can be distinguished:

Barremian to Early Aptian.

The depocenters are associated to very thick black shale deposits, while the surrounding carbonate platforms show a very low rate of sedimentation. One of the main salt diapiric event also occurs during this time along the border faults of the newly formed basins by migration of the Triassic evaporites from the subsiding depocenters towards the edges.

Latest Aptian to Early Albian.

The shelf-basin configuration is acquired during this period. The basins are yet characterized by a very high rate of subsidence and sedimentation of black shales. Halokinesis is still going on, such as at the edges of the Aptian-Albian pull-apart basins and rhomb-grabens, e.g. the huge Audignon anticline to the north of the Arzacq Basin. The shelf-basin transition is marked by a system of limy patch reefs while the platforms surrounding the shelves show a low rate of sedimentation.

Late Albian.

The paleogeography changes completely, and is entirely controlled by the important tectonic movements taking place along the plate boundaries (Peybernès and Souquet, 1984). As a result, the subsiding zones shift to the south towards this very mobile zone, while a massive transgression develops over the entire area. The Late Albian sedimentation overlaps the earlier paleogeographies, demonstrating the change of structural stress at this time. Locally salt tectonics enhanced erosion and subsidence, producing the main hydrocarbon traps as they are known today.

Late Cretaceous

An important inversion occurred diachronously all along the Europe-Iberia plate boundary. While the western part remains extensive in a rift mode, the eastern part of the domain progressively became in compression and in a foredeep mode. Correlatively, two extensive carbonate platforms developed, covering to the north the major part of the Aquitaine Basin, and to the south the persistently high Iberian plate, both carbonate and separated by the N-Pyrenean flysch-furrow in between. Thus, the Late Cretaceous paleogeography shows three major domains from North to South:

On the Aquitaine platform, to the north, the syntectonic sedimentation of the foredeep evolves laterally to a carbonate shelf (Dubois and Seguin, 1980) of Cenomanian to Maastrichtian ages. This domain is located in a more distal and more stable environment within the Late Cretaceous paleogeography. It is important to point out that, along the southern limit of the shelf (South Aquitaine margin), the Cenomanian carbonates unconformably rest over the Upper Albian sediments. This unconformity fades out northwards into the platform. ***Salt-lineaments*** of varied orientations as well as circular diapirs were also enhanced in this apparently stable platform and in the Parentis Basin. They are located mostly along major faults and fault-intersections, such as the diapirs near the city of Dax, and along the oil-bearing structures of the Arzacq Basin, (Fig. 6).

The N-Pyrenean basin progressively becomes equivalent to a Cenomanian to Maastrichtian foredeep located immediately to the north of North-Pyrenean Fault (NPF), thus bordering the European-Iberian plate boundary. The entire fast-subsiding, turbidite-filled, mobile belt becomes progressively the North Pyrenean Zone, where the major part of the orogenic thin-skinned deformation of the North-Pyrenees will take place (Zolnai 1971). During the

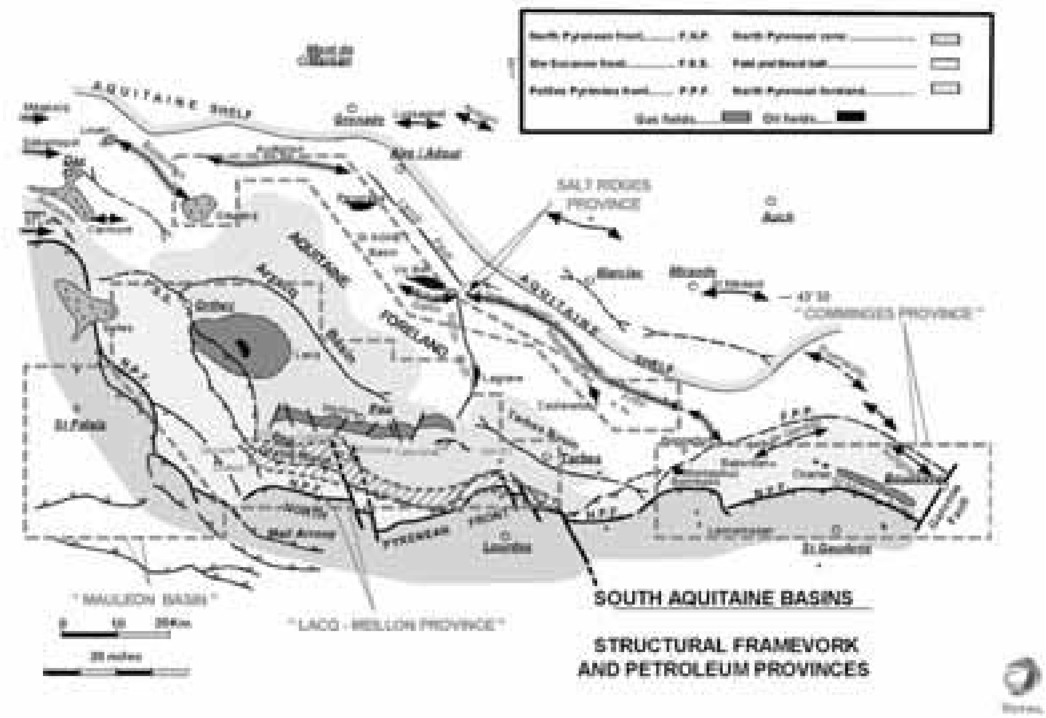


Figure 6 - Structural framework and Petroleum Provinces of the South Aquitaine and N-Pyrenean basins (Total).

Cenomanian, the en-echelon rift basins merged into a single N-Pyrenean basin, which corresponds to an E-W asymmetric and narrow trough or furrow (Moen-Maurel *et al*., 1999).

As compression between the two plates became prominent, the axis of the foredeep migrated northwards, from Turonian to Lower Maastrichtian. The flysch sedimentation in the deep North Pyrenean furrow by now became fine-grained, dominated by marls and ended with a thick mudstone series during the Maastrichtian. Transpression provoked wrenching and salt tectonics. Olistostromes composed of Triassic material (salt, shale) are also emplaced, mainly into the mobile domains of the foredeep (Stevaux and Zolnai 1975). Within the North Pyrenean furrow the salt-structures acted as paleogeographic barriers and influenced the distribution of the turbiditic systems.

To the South, on the Iberian plate, an Upper Cretaceous carbonate platform developed. The massive, thick (about 500m) Iberian Upper Cretaceous carbonate series of the “calcaires des Eaux Chaudes” or “calcaires des cañons” overlay the earlier sediments and/or the basement, with a sharp angular unconformity.

However, as a result of the ongoing north-westward Pyrenean compression and of the formation of the intraplate retro-foreland basin, the Southern Upper Senonian flysch domain progressed over the former Iberian carbonate platform along the area of the High Chain in the present-day central and western Pyrenees. Its connection with or independence from the North Pyrenean equivalent flysch basin is under study.

In conclusion, the two Upper Cretaceous carbonate p7latforms to the north (Aquitaine) and to the south (Iberian plate) are separated by a flysch foredeep then a retro-foreland basin. No transitional facies are known between the platforms and the North Pyrenean Trough sediments, as a result of the progressive cannibalisation of the platform edges.

Tertiary

The Paleocene was a time with relatively low plate convergence between Europe and Africa (Roest and Srivastava, 1991). After a Dano-Paleocene period of relatively reduced tectonic activity, the Tertiary paleogeography and its basin evolution are under the control of the ongoing compression, which is diachronous and oblique, progressing from east to west again.

The eastern and central parts of the foreland basins changed from marine to continental as topography developed and the amount of eroded material was sufficient to fill the basins. In the western Pyrenees the crust would have recovered its initial thickness later than in the central and eastern Pyrenees as a result of a greater extension of the Pyrenean crust during Early Cretaceous times coupled with a younger and lesser amount of convergence. In this area Paleocene rocks are represented by deep-water carbonate and siliciclastic sediments deposited in continuity with the Upper Cretaceous turbidites (Pujalte *et al*., 1989).

Two main events will arise :

1. the N-Pyrenean foredeep basin will retreat from east to west, and thus the related sedimentary facies synchronously will do the same, recording the orogenic evolution, horizontally and vertically. From east to west and from bottom to top, facies will pass from syntectonic flysch facies to tidal deposits and then to syn- and post-orogenic fanglomerates.

In detail: the Late Cretaceous-Paleocene time period is still marked by margin instability, with large submarine collapses of carbonates mass-flows (see here below, Stop 1.2 at Le Tucq).

In Late Paleocene-Eocene (Ypresian) rhythmic sedimentation continued in the N-Pyrenean basin, constituting a new thick flysch sequence.

During the Early-Middle Eocene thrusting rate increased. Both foreland basins experienced a deepening which resulted into the widest extension of marine deposits in the Pyrenean foreland basins (Puigdefàbregas *et al*., 1992, Burbank *et al*., 1992a). The thrust front in the southern side of the central Pyrenees strongly advanced because deformation of the Mesozoic wedge on top of a weak detachment level (Triassic evaporites). Shallow marine deposits were deposited in the foreland as well as in the piggy-back basins which demonstrate a subhorizontal mean topography over the southern frontal wedge. Strongly subsident troughs filled by turbiditic sequences were developed south of the uplifted basement in the footwall of the upper thrust sheets. Some relief existed hindwards as evidenced by the N-S river systems supplying basement clastics and by the geometry and location of proximal alluvial fans. A maximum topography of 1-2 km has been calculated based on paleotopographic reconstructions and flexural modelling (Vergés *et al*., 1995, Millán *et al*., 1995).

In Oligocene along the future Pyrenean foothills i.e. the orogenic foreland, sedimentation became progressively more terrigenous from east to west, with large masses of coarse continental fanconglomerates (“poudingues”) deposited in successive sequences, in a suite of fans.

The Ebro basin became closed and separated from the Atlantic because the tectonic relief growth during the inversion of the Lower Cretaceous basins in the western Pyrenees. Erosional debris of the Pyrenees and other surrounding chains of the Ebro basin (Iberian and Catalan Coastal Ranges) progressively filled the basin and then backfilled, to bury the flanking thrust belts on its margins (Coney *et al*., 1996). This progressive backfilling forced deformation to migrate hindwards and as a consequence reactivation of previously developed thrust structures and break-back thrust sequences occurred (Vergés and Muñoz, 1990, Burbank *et al*., 1992b). The southern Pyrenees were almost completely buried by Early-Middle Miocene times.

The greatest part of the Aquitaine Basin, to the north of the orogenic foreland, nevertheless, remained shallow marine, until the general filling of the basin caused the sea to progressively retreat to the west.

The abrupt and subsequent Miocene-Pliocene re-excavation of the southern Pyrenees and the Ebro basin to develop the present fluvial system was some combination of the Miocene rifting of the western Mediterranean and the Messinian desiccation crisis.

Structural history of the Pyrenees

Inherited structures of Precambrian to Paleozoic age affected the tectonic basement of the Pyrenean orogeny, and had a fundamental influence on the geometry of the Upper Jurassic to Upper Cretaceous extensional (grabens and rift margins) and on the Upper Cretaceous to Tertiary compressive structures (frontal and lateral ramps). This combination of pre-rifting events produced three trends:

- ***N 20°E to 40°E:*** corresponding to the Garonne - Villefranche fault zone crossing through Toulouse city. It is active in Late Variscan times as a left-lateral strike slip fault.
- ***N 90°E to N 120°E:*** its corresponds to the Celt-Aquitaine hinge-line and to the North Pyrenean fault zone which acts as a left-lateral strike-slip fault in Late Variscan times.
- ***N 160°E*** represented in the Axial Zone by kilometer-high folds and incipient cleavage of Variscan age.

A succession of events affected the N-Pyrenean domain (fig. 4 and 5), after the Variscan orogeny.

The **first post-Variscan rifting** attempt took place as soon as the **EARLY PERMIAN**, i.e. during the late stages of the post-variscan peneplanation. The Permian lies unconformably on the previously deformed series. Roughly N-S and E-W oriented, grabens are known in the Pyrenees as well as under the basin itself; they are filled with red continental clastics and volcanic flows, with thicknesses up to 1,000 m. One of these grabens is the transverse (NE-SW-oriented) Elizondo trough in Basque country (Gapillou 1981) near St Jean Pied de Port.

Triassic - Jurassic

The **Triassic** was widely deposited in a flat-lying basin in the centre of the area, while **a second rifting event** prevailed in both the Parentis Basin, and along the southern basin edge, (near the North Pyrenean Fault Zone). High emerging blocks (e.g. Arbailles, Grand Rieu) were formed along the southern basin-edge, which alternated with deep corridors already during the Late Triassic. Submarine (ophitic) volcanic flows and intrusions took place at the Triassic-Jurassic transition (approximately between the Keuper salts and the Liassic anhydrites), near the Iberian-European plate boundary, but also to the north.

The thick Keuper evaporitic and shaly potential detachment horizon, which locally can be more than 1 km thick, allowed diapirism, folding and thrusting, as well as gravity tectonics (Zolnaï, 1975).

Jurassic- Early to Mid-Cretaceous rifting

This period is characterized by the beginning of the opening of the Central Atlantic Ocean and the propagation to the North of the Mid-Atlantic Ridge. The Bay of Biscay rifting corresponds to the extensive phase of the Late Jurassic (Montagné, 1986) - Early Cretaceous "revolution" or **third rifting episode** which creates E-W striking normal faults connected by N 20°E right-lateral and N 160°E left-lateral transtensional faults. A regional sinistral wrenching between the two plates is related to that deformation phase. In Aquitaine, the Jurassic paleogeography is characterized by diffuse W-NW/E-SE extension. As a result, the Jurassic sedimentation appears to be controlled primarily by the inherited basement faults (essentially the N 20°E, N 50°E-70°E and N 160°E sets of faults).

Transtension and pull-apart basins :

The Early Cretaceous geological evolution is marked by the paroxysm of the general north-south extension and anticlockwise rotation of the Iberian Plate, which resulted in the opening of the Bay of Biscay. Although the major transcurrent motion takes place along the European-Iberian plate boundary in a Baja-California style (Miranda Avilès, 2002), the entire region from the Parentis basin (north of the Aquitaine Basin) to the Ebro platform is affected by diffuse strike-slip movements. The high rate of sedimentation is controlled by the activation of N 110°E inherited faults while the N 40°E and N 160°E directions are used as minor relay faults. The Jurassic platform is dislocated along these main fault zones (Canérot, 1989), allowing for the development of a set of NW-SE rapidly subsiding, diamond-shaped basins or potentially pull-apart basins (Bourrouilh, Richert and Zolnaï, 1995, R. Miranda Avilès, 2002) and uplifts (with E-W and NNW-SSE trending edges. The W-E to WNW-ESE oriented **North Pyrenean Fault** was initially a braided fault-system. It stretches over 500 km in successive segments, between the easternmost Pyrenees to Basque country, where it splits into several E-W oriented branches, separating the different elements of the Basque Massifs. Its trace is lost beyond the transverse NE-SW Pamplona- fault array.

North-Iberian margin and oceanisation:

During Late Jurassic and Early Cretaceous, as the opening of the Bay of Biscay progressed eastward, faulted tilted blocks (of ten-km width), calved from the Iberian margin and landmass. Large areas of the present-day High Chain became emergent e.g. to the south of the Mendibelza massif in Basque country (Boirie and Souquet 1982), or in the Central Pyrenees, offering source areas for coarse clastics, which deposited into the deep sea-arm ("aulacogen", Souquet and Debroas 1981) to the north. Above the basal wildflyschs, the first thick sandy-shaly turbidite (flysch) sequence was deposited (Albian-Cenomanian).

In the meantime tholeitic basalt masses (pillows and lava flows as well as sheeted dykes and sills) were emplaced within the flysch furrow, between Bilbao and the area south of Tarbes. Correlative to the Bay of Biscay sea-floor spreading, the strike-slip tectonics and the differential drift movements of the Iberian and European plates (Plate 1, figs. 1 and 2), provoked the remobilization of lherzolites and a peculiar shear metamorphism all along the N-Pyrenean fault zone, known as "N-Pyrenean metamorphism".

Late Cretaceous – Tertiary diachronous extension/collision

Compression comes from the East...

Since at least the beginning of Late Cretaceous, the Iberian plate became in compressional contact with the NW moving African plate. Iberia progressively crushed the space between its northern margin and the European plate. Compression was diffuse during Late Albian to Cenomanian; it became effective as early as in the mid ***Late Cretaceous*** (Turonian-Senonian) in the eastern part of the Pyrenees (Henry and Mattauer 1972, Souquet and Deramond 1989), and propagated progressively to the west during the Senonian. Compression provoked gigantic linear gravity slope sequences or "Evolutionary-mass flow-megaturbidites" (Bourrouilh *et al*. 1983), which can reach 90 km in length and a thickness of about 60 m. Similarly, in the southern Pyrenean Basin, but during the Eocene, compression is accompanied by the emergence of a syntectonic cycle of gigantic mass-flow-megaturbidites extending over 140 km in map view, and more than 200 m thick (Soler *et al*., 1970, Johns *et al*., 1981).

The main north-south compressional event of the Pyrenees began nevertheless from **MIDDLE EOCENE** onward. The main ductile deformation occurred at the plate boundary, within the North Pyrenean Fault Zone, which represents the most strongly compressed, schistosed, crushed part of the belt (Choukroune, 1974), and along the northern edge of the Iberian block. Former normal faults and hingelines were inverted into overthrusts, the northern limit of the highly compressed zone corresponding to the earlier boundary between the flysch furrow and the northern carbonate plaform. This long-lived tectonic edge has thus become the North Pyrenean Thrust Front.

The main structural events created or re-activated in this period from Late Cretaceous (Maastrichtian) to Oligocene are:

N 110°E folds, cleavage, stylolites, reverse faults and thrust planes either to the north or to the south.

N 20°E to 40°E, left-lateral lateral ramps.

N 160°E, right-lateral lateral ramps.

Normal faults are also present: in some areas they are striking N-S; some minor sub-E-W faults are also created in close relation to major thrusts trending E-W and salt ridge keystone collapse.

During the orogeny, several compressional episodes followed each other, the younger, more external faults (sole-faults of allochthonous units or rafts) deforming the earlier, more internal ones in a piggy-back fashion. The deformation and structural styles appear controlled by the rheology of the material involved: (a) the already fractured deep basement complex was involved in the compression, setting the sites for the major dislocations, and (b) major formations of the sedimentary column (e. g. Keuper salt, Lower Cretaceous marls, Albian to Tertiary flysch and molasse series) were still extremely incompetent during the period of compression. Large, open synclines and anticlines were first formed; the synclines later became overthrusted, overturned and/or thinned, sheared by the overriding anticlines. Specific structures or " nappes " were thus generated, which would seem uncommon in areas of greater crustal rigidity and with more competent sedimentary sequences (there are very few true flat-and-ramp thrusts and "duplexes" in the northern Pyrenees, Moen-Maurel *et al*., 1999).

The North and South Pyrenean Thrust Fronts are actually tectonic envelopes, and not continuous, unique, sole faults. They are composed by segments which relay each other, sometimes with important offsets. On the Northern Pyrenean side, several structural arcs exist: the Basque Arc, along the north-western Pyrenees, the Lannemezan Arc in the central northern Pyrenees etc., but also on the southern side, such as the Graus-Tremp Arc in the southern Pyrenees; their tips all lay in lateral or oblique ramps which were inherited from the rift basin edges. Two well-explored surface anticlines are centered in the northern arcs : the Ste. Suzanne structure to the west, and the "Petites Pyrénées" in the Central Zone. Thrust sheets do not exceed 12 km.

The Southern thrust sheets are therefore the most spectacular ones:

From Early Eocene to Oligocene, the Upper Thrust Sheets (Muñoz *et al*. 1986) or the South Pyrenean Central Unit (Seguret, 1972), composed of Mesozoic to syntectonic Tertiary formations, were successively transported along a sole of Keuper evaporites, argilites and marls (Bresson, 1903, Seguret, 1972, Muñoz *et al*. 1986). They are, from East to West, the Pedraforca, Boixols, Montsec, Cotiella Thrust sheets; the front of all outlines the South-Pyrenean Front Thrust, which produced the Sierras Marginales (plate 1).

In the High Chain, these cover thrust sheets root into the Upper Paleozoic basement, along with the deeper thrusts which are the Nogueras and the Gavarnie-Mont Perdu Units.

The total shortening of the Pyrenees is evaluated, depending on the authors, between 100 to 160 km.

Pyrenean orogeny and inversion continued during the **MIOCENE.**

Partitioning of the deformation (see Miranda Avilès, 2002) as well as rotational constrains induced along the basement faults produced "N-40°E" dip-slip (mostly left-lateral due to S-compression) and "N-160°E" (right- and left-lateral).

Neotectonic relaxation and uplift

The very latest, **PLIOCENE to HOLOCENE** episode of the Pyrenean orogeny seems to have been an important **relief development**, which uplifted the former southern (Iberian) margin to form the present-day High "axial" mountain Chain (+ 2,000 to +3,500 m MSL.), where Paleozoic outcrops were exhumed by erosion. This uplift movement, possibly a late- or post-orogenic isostatic adjustment, also deformed the former overthrust sheets, back-steepening those of southern vergence to near vertical.

Petroleum exploration framework

History

The Aquitaine area represents the largest oil and gas province of France. Although seeps have been known since Roman times, petroleum exploration of the South Aquitaine basins (foreland and foothills) started in the 1930's and resulted in the discovery of the St-Marcet Gas Field in 1939 (8 Gm3 of gas – 290 BCF) (Comminges Basin). The discovery well was drilled on a surface anticline. The area's positive potential was then confirmed by the discoveries of the upper Lacq oil field in 1949, the deep giant Lacq gas field in 1951 (260 Gm3 gas – 9.2 TCF) (discovery wells were drilled on gravimetric anomalies) and the Meillon gas field in 1965 (65 Gm3 gas – 2.3 TCF) (this discovery was made using 2D seismic data showing a surface anticline). In the same Arzacq basin, several smaller sized fields (Ucha/Lacommande, Rousse, Cassourat …) with accumulations ranging from 3 to 7 Gm3 gas (110 to 250 BCF) were also discovered during the same period. In the 1970's, exploration interest moved northward towards the basin edges, resulting in the discoveries of five sizable oil fields (Pecorade, Vic Bilh, Lagrave, Castera Lou and Bonrepos-Montastruc, fig. 6). Two hydrocarbon trends are clearly identified: a gas trend to the south along the leading edge and the proximal foreland of the Pyrenees Fold and Thrust Belt (the "Lacq-Meillon" domain), and an oil trend to the north along the edges of the foreland basin (the "salt ridges province") (fig. 6 and 7).

Reservoirs (Fig. 7).

Jurassic dolomites and Barremian limestones

They represent the main petroleum objectives; their occurrence is closely linked to the Jurassic and Earliest Cretaceous paleogeographic setting. Two domains can be distinguished in accordance with the paleogeographic provinces described earlier.

On the Jurassic eastern shelf the reservoirs are represented by the Lower Kimmeridgian Meillon dolomites (200 m thick in average), the Portlandian Mano dolomites (150 to 200 m thick) and Garlin breccias. These reservoirs are totally or partially gas-bearing on the Meillon gas-field trend as well as along the Ucha, Lacommande and Rousse trend of structures. For these two reservoirs the petrophysical characteristics are rather poor (2 to 4% matrix porosity for the Mano dolomites and 4 to 8% for the Meillon dolomite). For these formations, effective permeability is primarily due to joint fractures and vugs (due to secondary diagenesis), allowing for good well productivity.

In the Jurassic outer shelf to the west (Lacq, Pecorade and Vic Bilh fields), only the Mano dolomites are preserved as reservoirs within the Jurassic sequence. Even though the petrophysical characteristics are better, they remain poor (2 to 10%). Production is again primarily associated with intense fracturing and intervals characterized by a high energy of deposition. Within this same area the reservoirs also include the Upper Barremian limestones, which display porosities ranging from 10 to 15%. Permeability is again relatively poor but enhanced by the intense fracturing.

The Lower Barremian is also one of the reservoirs of the Lacq giant field. Its low porosities (2-11%) and permeability are also enhanced by intense fracturation.

The Upper Cretaceous reservoirs

They correspond to the Lower Senonian limestones (200 m to 250 metres thick) of the Upper Cretaceous platform. The good reservoir characteristics of these limestones (15 to 20% matrix porosity) are closely linked to the high energy facies distribution and to a secondary dolomitization in the vicinity of the main Pyrenean faults. It is demonstrated by the example of the Lagrave Field where dolomitization and therefore reservoir capacity decrease significantly away from the immediate setting of the Tertiary Seron transcurrent fault system.

Seals (Fig. 7).

Barremian and Jurassic plays

Hauterivian to Valanginian shales, Barremian shales and anhydrites and Lower Aptian shales represent the most efficient seals for the oil and gas fields in the

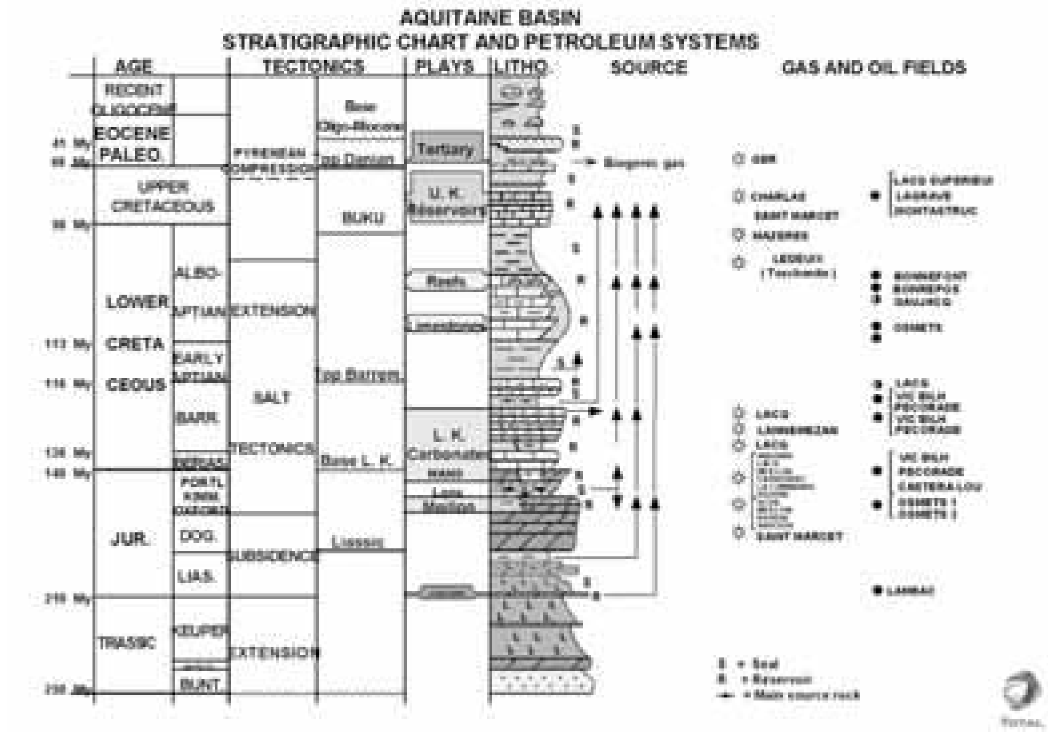


Figure 7 - Aquitaine basin : stratigraphic chart and Petroleum Systems (Total).

Arzacq basin.

Upper Cretaceous plays

Seals are composed of the Upper Senonian shaly or marly intervals and by the Lower Tertiary slope shales.

Source rocks (Fig. 7).

Important reserves of oil and gas have been discovered in the Aquitaine Basin, implying that sources with a regional extension and good petroleum potential are present in order to expel significant quantities of oil and gas. Although some source potential has been recognized in the Tertiary, Albian and Liassic shales, these formations are low-quality source rocks.

The main source rocks are clearly associated to the Barremian and Kimmeridgian formations (Lons Fm) as shown by the source-to-oil condensate and source-to-gas correlations.

The Barremian organic matter is primarily of marine origin including woody material (type II-III). Across the basin the oil window is reached at – 3,000 metres on average, while the gas window is reached at – 4,000 metres (fig. 8). It is noteworthy that expulsion of hydrocarbons started during the Late Albian at the level of the Lower Cretaceous depocenters of the Arzacq Basin and occurred during the Tertiary on both sides of the basin (fig. 8).

The Kimmeridgian source rocks appear to have the best potential. The organic matter is again primarily of marine origin (type II-III).

Trap types and hydrocarbon charge (Fig. 8).

The traps for the different fields are clearly related to structures which are inherited from the Early Cretaceous extension (platform to basin transition) and its associated salt tectonics along the limits of the basins (erosional pinch-out of the Jurassic and Barremian reservoirs). These traps have been enhanced by the Pyrenean orogeny. Some of them are located in the Pyrenean fold and thrust belt (Saucede and Ledeuix); but most occur within the proximal (Lacq, Meillon, Ucha, Lacommande, Rousse fields) and distal forelands (Vic Bilh, Castera Lou, Lagrave). Exceptional productive traps can be even found in episyenites (Moen-Maurel *et al*., 1996).

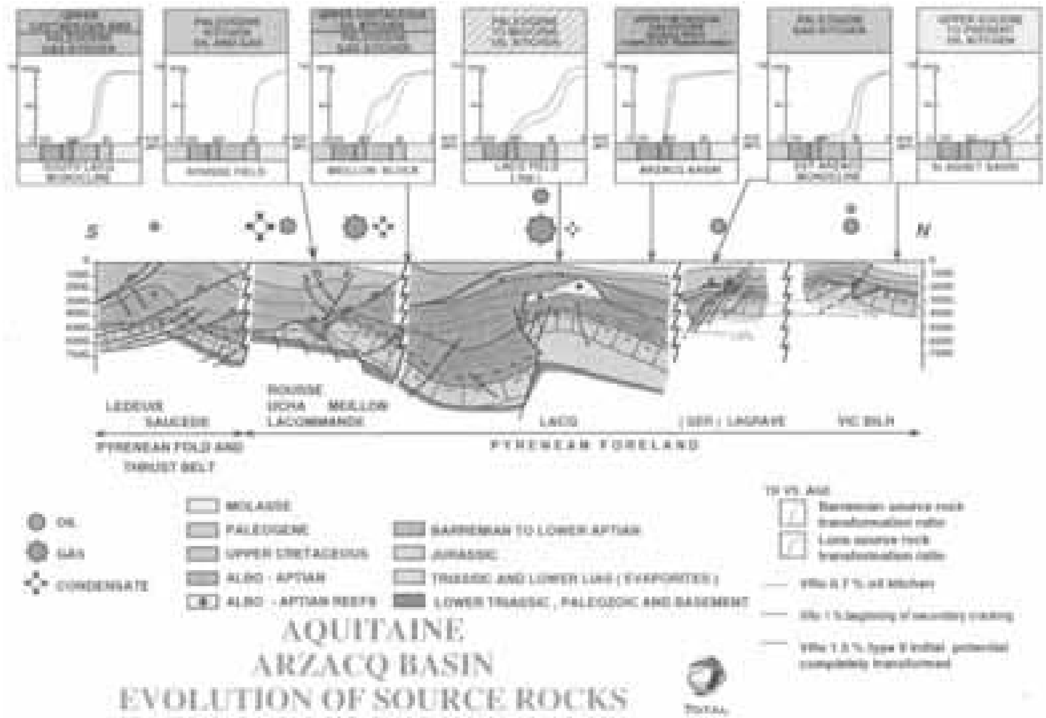


Figure 8 - Aquitaine-Arzacq Basin : Evolution of Source Rocks (Total).

Comparison with other Fold-and-Thrusts Belts

In comparison to others FTB, the following striking characteristics should be pointed out :

1) Crustal characteristics :

The continental crust here underwent diffuse stages of rifting from Triassic to Albian at the western tip of the tethyan domain over a broad area (400 km N-S wide) extending from the Parentis basin to the north to the Ebro basin and Cantabrian domain to the South. At the paroxysm of the Albian rifting, many rifts that can be investigated in the subsurface today, were en-echelon, as a result of the control of previous oblique basement fabrics that were acquired in Paleozoic times. However extension was never sufficient east of the Bay of Biscay to produce an oceanic floor and two differentiated passive margins.

Pyrenean (Late Cretaceous to Eocene) inversion mainly affected the Iberian plate, leaving the rift basins of the European plate in a little deformed retro-arc position (and preserving its petroleum systems, Moen-Maurel *et al*., 1999).

2) Plate motion vector evolution :

The anticlockwise rotation of the Iberian plate with respect to the European plate from Early Cretaceous to Cenomanian led to variations in trend, magnitude and sense of the displacement vectors, producing at the same time extension in the west and compression or strike-slip in the east.

The oblique extension between the Iberian and the European plates from Early Cretaceous to Cenomanian led to the partitioning of the deformation between strike-slip which focused on specific shear fault zones (supposedly reaching tens of km in strike-slip), and diffuse extension which produced en-echelon grabens with minor horizontal displacements along faults (no more than 5% horizontal stretching, Moen-Maurel *et al*., 1999).

During collision the use of potential detachments in the Triassic evaporites and in the Upper Cretaceous flysch led to varying structural styles making it difficult to measure the amount of strike-slip in the shortening phases from Turonian to Miocene. Nevertheless most thrust tips still exist at the outcrop or in the subsurface, therefore the cover deformation cannot accommodate any major oblique collision.

3) Salt tectonics

The existence of Triassic evaporites led to halotectonics from Late Jurassic onward. Salt pillows and ridges were re-activated as anticlines during compression while evaporite-rich widespread units acted as detachments and thrust soles. The salt mobility triggered enhanced amplitude and curvature in fold axes, non-cylindricity and virgation, as well as a diachronous record of the tectonic stress events. When salt is involved onset and end ages of tectonic events may vary through time and space making it difficult to tie deformation to stress and/or plate tectonics.

4) Fold-and-Thrust belt structural style:

Thick-skinned thrusting affected the Iberian plate margin along the underthrusting plane, resulting in antiformal stacks of basement (Paleozoic) producing the axial zone of the mountain range. The steep dips of the basement ramps may have been facilitated by the oblique component of the collision initial vector and the existence of steep and weak shear zones from the Early Cretaceous drift stage (such as the North-Pyrenean Faut system).

Thin-skinned thrusting was possible when cover detachments could be activated over large distances: for the southern FTB it resulted in transferring the uplift and the 100 km range shortening in piggy-back fashion over a few large thrust sheets towards the external sierras, and for the northern FTB it resulted in producing a retroarc and in accommodating the axial overload with thrusts and nappes mostly affecting the Upper Cretaceous to Eocene flysch basins.

The shape and magnitude of the thrust sheets was strongly influenced by the attitude and potential friction of the detachment stratigraphic horizons: salt ridges in the north favoured large amplitude (up to 5 km) upright and overturned when they slid over flysch units. Rift grabens and flysch basins morphology influenced the location of both frontal and lateral ramps. This stratigraphic control leads to structural styles that are common along the alpine (tethyan) orogen, unlike craton margin orogens where isopach formations rather, led to flat-and-ramp thrust propagation such as in the Appalachians and the Rocky Mountains.

5) Petroleum evaluation

The North-Pyrenean retro-arc basin was a prolific oil and gas province with giant fields (Perrodon, 1980) because:

- the petroleum system trilogy (source rock, reservoir and seal) belonged to pre- and syn-rift series which are usually characterized by extensive and stable marine sedimentation making hydrocarbon migration easy and exploration more predictable even in areas of poorly imaged seismic prospecting (Moen-Maurel *et al*, 1995).
- the petroleum system trilogy series were protected from uplift and erosion over the Parentis and North-Pyrenean basins since they remain situated in the outer rim of the collision, in a little affected retroarc position. After rifting subsidence, the additional burial which was achieved with the foredeep deposition led to reaching the gas window and to the development of large gas traps.
- salt tectonics eventually enhanced the amplitude of traps and their closure. Such providence makes the Aquitaine oil and gas structures quite similar to North Sea oil and gas fields.

To conclude, it should finally be emphasized that first of all, the structural history of the whole area is controlled by a N110°E plate boundary crustal anomaly, running from West Greenland to the Tethys and then to the modern Mediterranean (R. Bourrouilh, 1970, 1973). This lineament is evidenced by the N-Pyrenean Fault and its peculiar metamorphism, the thinning of the crust, the Biscay rifting, the injection of the mantle lherzolites. The plate boundary crustal anomaly seems to be Pre-Triassic, putting in contact an Iberian deeply structured (with pre-existing crustal detachment faults) and thick continental lithosphere to the south, and a thin, less structured European one to the north. The Late Cretaceous to Tertiary tectonic shortening is asymmetric, being estimated, depending on the authors and the area, at 100 to 160 km over the Iberian plate, and at 5 to 12 km over the European one.

In its western part (Bay of Biscay), the Pyrenean orogen is incomplete. To the east, it interferes with the Betico-Kabylo-Alpine orogen and is overprinted by the extensive Neogene Mediterranean structures.

The result of the collision in the Pyrenees produces an asymmetric fan-shaped orogen, resulting from the crustal ductile underthrusting or underplating of the Iberian Precambrian-Paleozoic basement underneath the European plate.

The Pyrenean orogenic complex of the Pyrenees is devoid of granitic intrusions and syn-orogenic volcanism, there are no ophiolitic masses and the known metamorphism is restricted to the trace of the North Pyrenean Fault Complex (Choukroune 1974). This indicates that the major orogenic event in the area was the variscan one (tectonics with major crustal shortening, penetrative metamorphism and granitization). Parts of this structural crustal heritage (or regional fabrics) were activated during the Mesozoic extensions, re-utilized by the Late Cretaceous to Tertiary Pyrenean orogenic convergence.

Field Itinerary

DAY 1

Pau

Welcome to the field trip and the Pyrenees

By R. Bourrouilh, L. Moen-Maurel, J. Muñoz And A. Teixell.

Briefing and introduction to the Field Trip.

Stop 1.1:

Panorama of the Pyrenees : figure 1.1.

I. A transect of the northern pyrenees

By R. Bourrouilh* and L. Moen-Maurel**,

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DAY 2

from Pau to Oloron Sainte Marie

Stop 2.1:

Le Tucq : the North-Pyrenean margin and the Upper Cretaceous flysch : figures 2.1 a, 2.1 b, 2.1 c, 2.1d

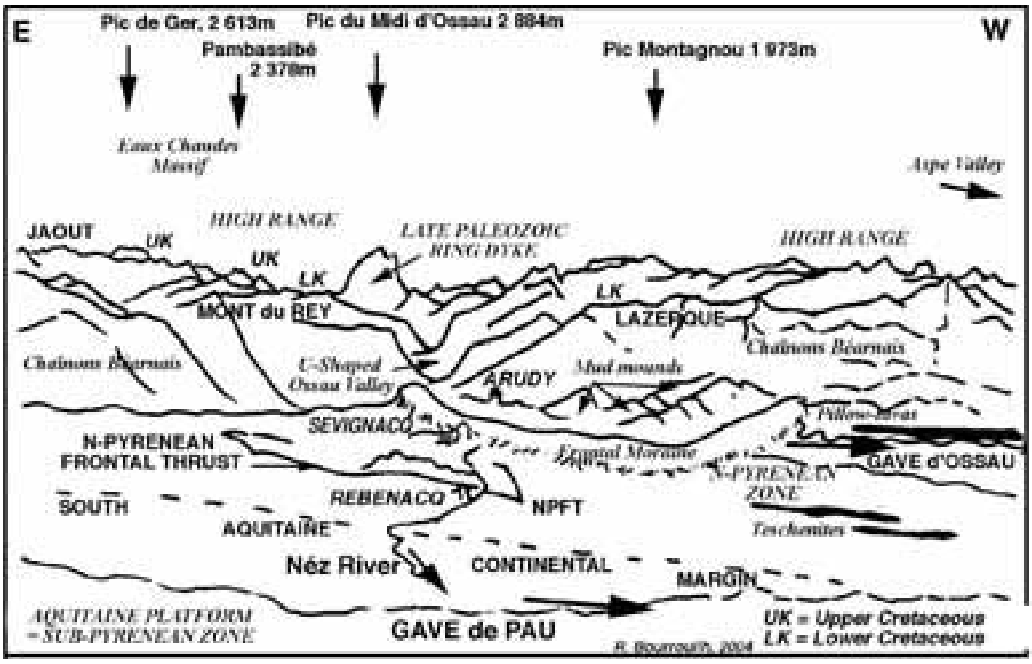
Introduction: the South Aquitaine margin developed as a passive rifted margin during the opening of the Bay of Biscay. The N-Pyrenean rifted basins were created and filled by syn-rift Lower Cretaceous black shales. During Cenomanian (-96 Ma), rifted basins merged in a long and narrow furrow or the N-Pyrenean basin or trough, which was rapidly filled up by synrift and drift calcareous Cretaceous Flysch. However, the area underwent compressional tectonics as recently as 80 Ma ago, during the Late Cretaceous and the furrow rapidly retreated westwards, from 80 Ma to today, forming, up to now, the modern Bay of Biscay. The North-Pyrenean flysch sedimentation, can reach 7 km in thickness.

The section studied in this Stop pertains to Upper Cretaceous-Paleocene sedimentation (but the KT limit is now covered by road work) which is coeval with the closure of the basin, and the westward progradation of the silici-clastic Paleocene sedimentation.

Access: the Stop is on the D 934, on the right side of the road, just at the locality “Le Tucq”. The outcrop section is located in the curve of the road (figure 1.1a) and the measured section (figure 2.1b, Tucq section and 2.1c, Guillampeau section) along the side of the adjacent car track.

Structural and sedimentological landscape: the D

Figure 1.1 - Geological view of the Pyrenees, from the Boulevard des Pyrenees, Pau.



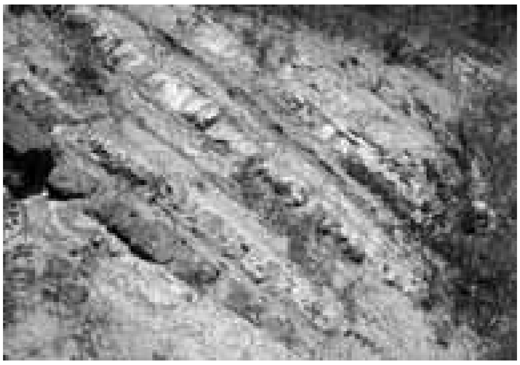


Figure 2.1a - Flysch section at Guillampeau, top is on the right.

934 crosscuts the Upper Cretaceous flysch. The curve of the river, as well as that of the road are provoked by the top of the Upper Cretaceous flysch, containing large white mudstone mudflows, as well as reworked Triassic (first diapiric) gravity slabs mass-flows, silici-clastic mass flows etc. In the landscape, the curve is marked by an old chalk oven, and prolonged by a smooth wooded cliff.

Along the western side of the road two field sections have been measured in the gently north-dipping turbiditic and mass-flow sequences with internal slumps (Bourrouilh, Montagné and Doyle, 1993, unpublished data).

The first one, or the Tucq section is just on the curve of the road, in front of the Tucq restaurant, but is now covered by road work (figure 2.1b):

The Upper-Cretaceous to Paleocene section is 27 m thick and represents a mixed silici-clastic-carbonate sedimentation. Five sedimentary units can be distinguished. On the four basal units, the sedimentation is mainly constituted by successive silici-clastic mass-flow deposits, reworked mudstones and anoxic mud clasts, and by silici-clastic Bouma-type turbidites. Measured sedimentary fold axes are, successively, from bottom to top : N 60°, dip 35°E (unit 2, lower part), N 83°, dip 10° SE (small slump, unit 3), N 175° (large slump, unit 3), N 90° to 110° (clast, base of unit 4), N 115°, dip 25°N, (base of unit 5), N 115°, dip 22° N-NE.

However, carbonate gravity sedimentation comes back again and forms unit 5, considered the base of a large carbonate mud-flow.

Figure 2.1b - measured section at Le Tucq, from R.Bourrouilh et al., 1993, unpub. data.

Figure 2.1ca - measured section 1 at Guillampeau, from R.Bourrouilh et al., 1993, unpub. data.

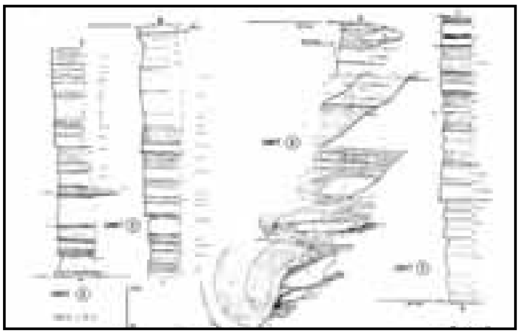


Figure 2.1cb - measured section 2 at Guillampeau, from R. Bourrouilh et al. 1993, unpub. data.

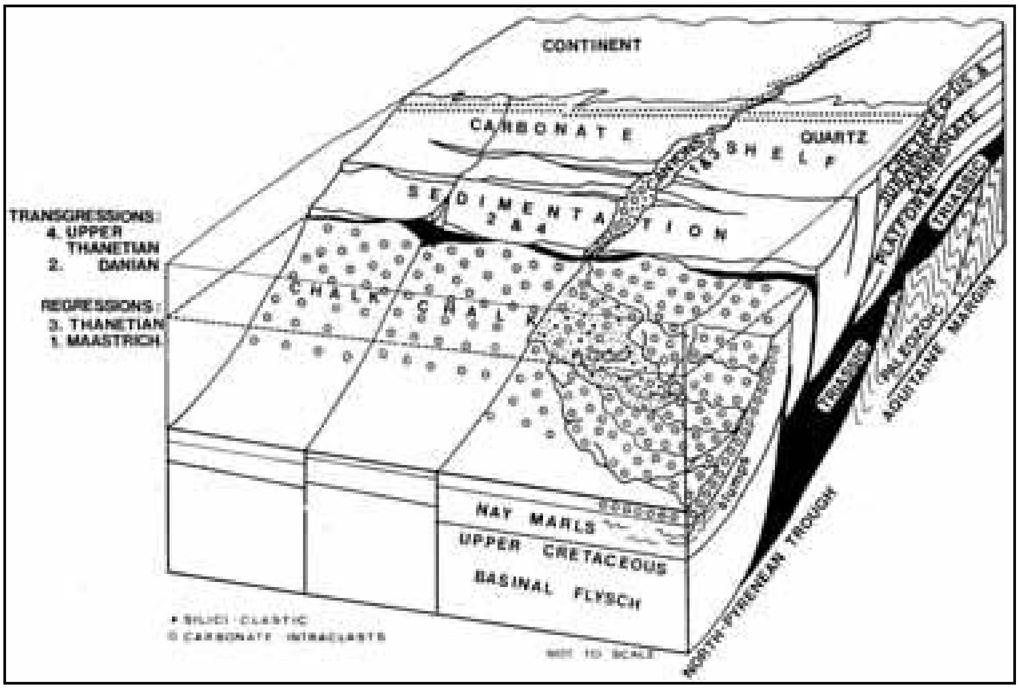


Figure 2.1d - Schematic reconstruction of the Dano-Paleocene Aquitaine margin at Le Tucq, from R.Bourrouilh et al. 1993, unpub. data.

The second one or the Guillampeau section is measured along the side of the road (fig. 2.1ca and 2.1cb)

Conclusions: figure 2.1d : the margins of the N-Pyrenean basin appear to be instable during the late Late Cretaceous to the Early Paleocene even though thicknesses across the North-Pyrenean domain do not vary as much as below in the Campanian and above in the Ypresian. This relative instability is due to a high amount of carbonates deposited on the edge of the shelf and to the progressive closure of the N-Pyrenean basin-Bay of Biscay, due to the progressive westward migration of the interplate compression.

For further information: Stevaux and Zolnaï, 1975, G. Montagné, 1986, Seyve, 1984.

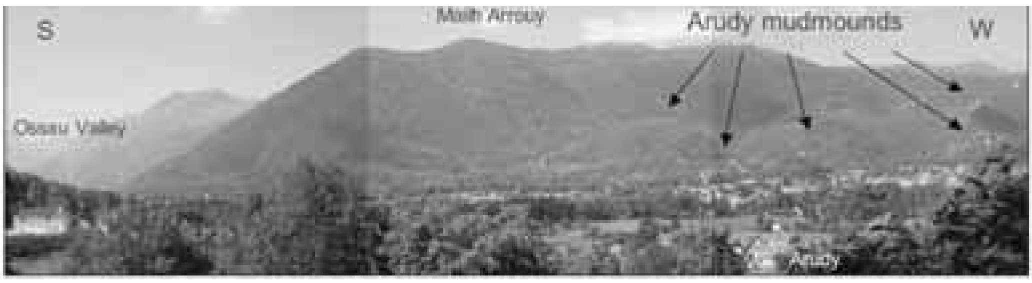
Stop 2.2:

the Ossau Valley: uplifting, glaciers and neotectonics: Sévignacq (figure 2.2)

The small village of Sevignacq is built on the morainic frontal vallum of the glacier, which went down from the High Range and the Pic du Midi d'Ossau, flowing northward through the Ossau Valley (in Béarnais, which is the local language: Aüssaü = high valley).

Access: leaving the main road D 934 from Pau to Laruns, at Sevignacq, on the right take the small road

Figure 2.2 - Panoramic view of the Mailh Arrouy and Ossau Valley (Total).

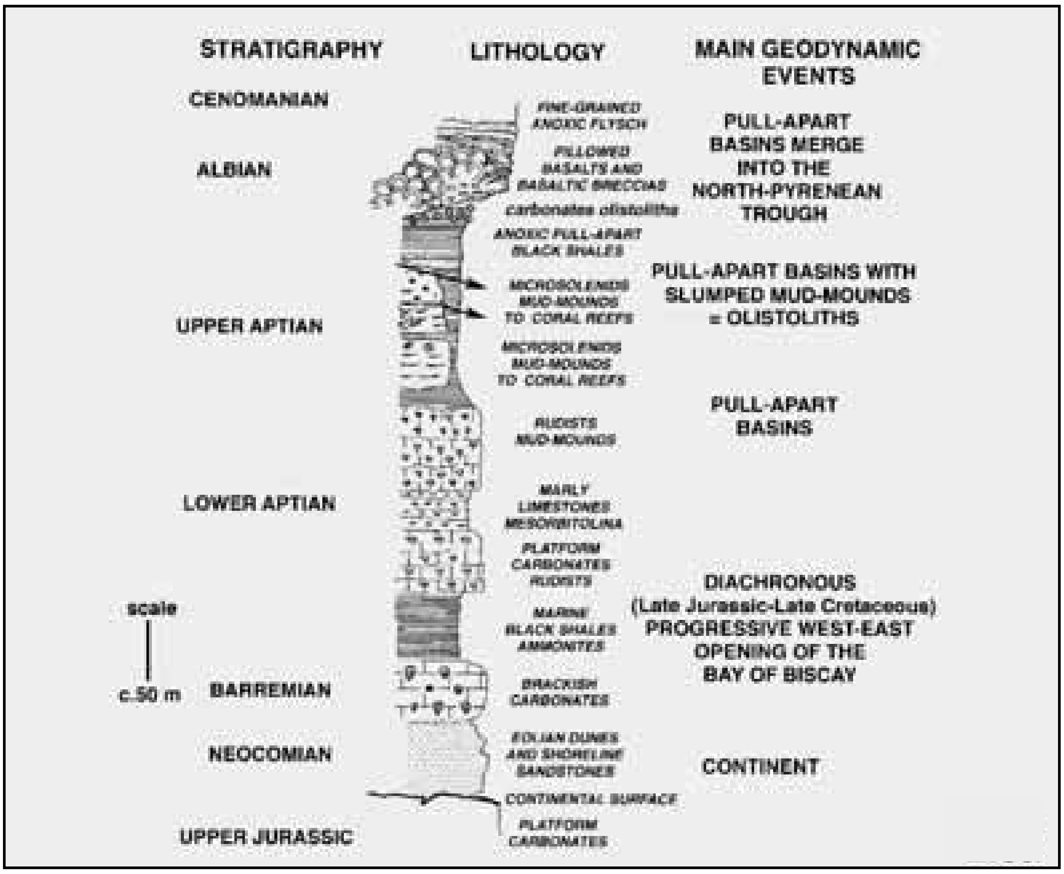


D 232, from Sevignacq to Bescat. The Stop is located after leaving Sevignacq and a few hundred metres down, after the cemetery, in the plain.
Structural landscape: the panorama shows, from left to right: the Mesozoic high mountain of Mont du Rey, the glacial U-shaped valley of Ossau, with the high chain and the faulted caldera of Pic du Midi d'Ossau in the background, and on the right, the Mesozoic Lazerque (1,368m) and Mailh Arrouy (1,251m) mountain range.
Interpretation: the Ossau glacier was one of the most important Würmian Pyrenean glaciers, which flowed south to north, from the High Chain to Sevignacq. At its front, it produced a large moraine, which surrounds us as a morainic vallum. Deglaciation began some 11,000 years ago and the glacier was about 1,000 m thick here. Melting of the ice provoked a reduction of the ice weight on the crustal lithosphere, which began an upward isostatic buoyancy and slowly rose up. The mountain range is still being uplifted under relaxation at present. Rising of the mountain all around provoked in turn the Ossau river to change its course from north to west, cutting the moraine and the Urgonian limestone to the West.
The Ossau river now flows towards the west after cutting the frontal moraine and the cretaceous bedrock in a narrow deep neo-tectonic canyon, located in the western outskirts of Arudy village.
But an underground northward paleocirculation of the Ossau waters still exists and flows to the north, under the frontal moraine and producing a resurgent spring in the karstified limestone north of Rébénacq. There, the river is called "Luy du Néez".

Stop 2.3:

Arudy: opening of the rift as recorded by mud mound markers (fig. 2.3a and 2.3b)
Introduction: at the village of Arudy, several Aptian-Albian mud-mounds are or were worked for marble. Geological studies (Bouroullec *et al*., 1979), generally

Figure 2.3a - Stratigraphy and geodynamic evolution of the Arudy basin, from R. Bourrouilh, 2000.



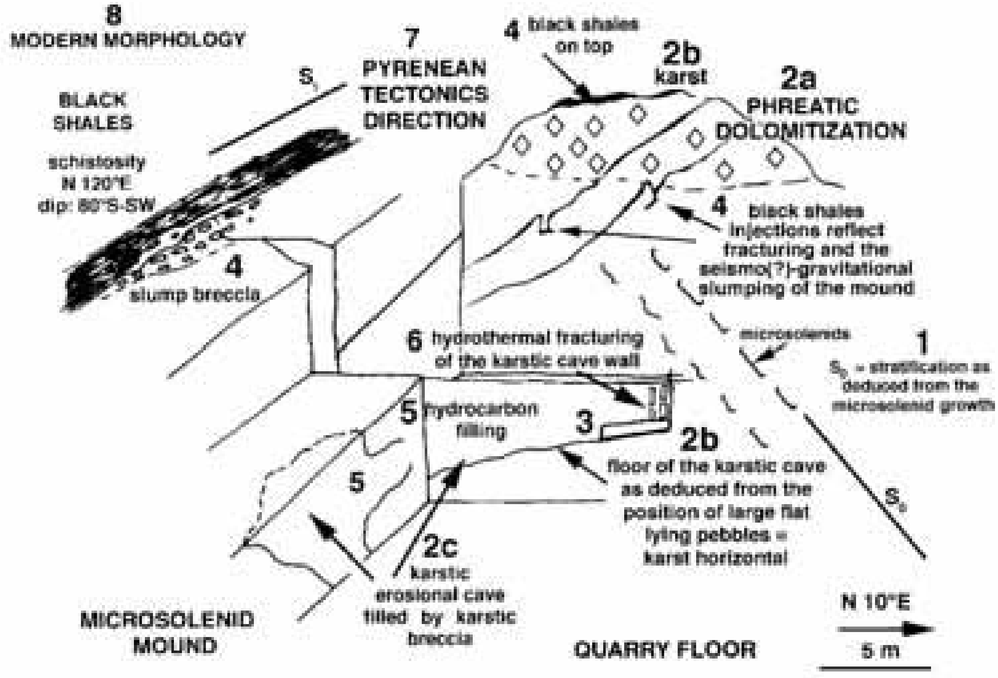


Figure 2.3b - Sketch of the Bois du Bager quarry, as visible in 1997, from R. Bourrouilh, 2000.

dealt with stratigraphy and micropaleontology. Digbehi (1987) was the first to present consistent sedimentological data on the area and to suggest that some reefs and mud-mounds of Arudy were in fact large blocks that have slumped into the black shale basin. Lenoble et Canérot (1993) and Canérot (1996), studied the relationship of the buildups to sea levels changes but they did not consider that some of these buildups could be not in their original position or that they could have been gravitationally displaced from their original environment.

Here we will only visit the Bois du Bager quarry, which represent an uppermost Aptian mud-mound slid into the black shale basin of Arudy during the latest development stage of the mud-mound complex.

Access: only the Bois du Bager quarry (or Borde Dela quarry) and mud mound, will be visited in this excursion. Access to the quarry is restricted. Ask Mr Laplace for access authorization well before visiting at the Paloma quarry along the Arudy to Lurbe road. Visiting the quarry, even when authorized, is at your own risk. North of Arudy, follow D 918 towards the West (Lurbe-St-Christau direction). After 1 km, turn left on a gravel road crossing the forest and leading up to the quarry, past a metal gate.

Structural landscape (R. Bourrouilh, 2000): on the southern edge of the N-Pyrenean rift, the stratigraphic sequence (Fig. 2.3a) shows a typical extensional and transgressive sequence: an Upper Jurassic to Lower Cretaceous disconformity is overlain by continental sandstones (Grès de Lacq), then by Lower Barremian lagoonal carbonates. From the Bay of Biscay, the sea invaded the area south of Pau during the Barremian and Ammonitic black shales were deposited. Then an Aptian-Albian carbonate platform developed: near the small village of Arudy, south of the city of Pau, this platform was marked by the growth of Aptian-Albian reefs and mud-mounds (Fig. 2.3a and 2.3b). These build-ups grew in a context of anoxic pull-apart basins. These pull-apart basins were progressively intruded by basic volcanism with floods of basalts, indicating the eastward progression of sea floor spreading of the Bay of Biscay onto the European continent (Plate 1, fig. 1 and 2).

We will focus here on the Bois du Bager mud-mound, situated 2 km south-west of Arudy. It consists of massive grey mudstones containing flat microsolenidae corals. According to L. Beauvais (pers. comm. 1995), microsolenids can develop in quite deep water and can tolerate a certain amount of turbidity in the water.

Those corals have umbrella-shaped morphologies, indicating a low energy environment; they grow in successive layers, indicating the stratification and the upper part of the mound.

The base of the mound shows a locally deformed contact with the underlying black shales (Fig. 2.3b). Along the basal contact, the lower part of the mound shows a slump breccia, with large mud-mound elongated clasts, some of them with microsolenidae, embedded in a black shale matrix. The mound also exhibits two layers of black shale, locally contorted and with load pockets. These layers of black shales can be interpreted as recurrence of mud sedimentation, or they reflect shale intrusions into the mound which slid downwards with the mound itself. Observation of the orientation of the microsolenids show that these two black shales are obliquely oriented, and clearly indicate that the mud-mound has tilted over to an angle of about 90° from its normal position, considering that the concave side of the umbrella-shaped microsolenids faces upwards. The body of the mound consist of massive grey microsolenidae mudstones but it also shows a large karstic erosional cavity which reflects emergence (Canérot, 1996). The cave is filled with a karstic breccia, partly originating from the mound itself. (Fig. 2.3b). The breccia is partly dolomitized and locally impregnated by bitumen.

In order to determine the diagenesis of the mound and the quality of the marmoreal limestone, a new geophysical tool, the electrostatic quadripole was used (Benderitter *et al*, 1997). A precise resistivity map of the diagenesis and of the fracturing was obtained, showing the mud-mound build-up, the karstic cave and its brecciated karstic infill.

This side of the mound is karstified and the karstic cavities are filled with black shales (Fig. 2.3b). This area presents a white to light grey colour and it is largely dolomitized. Study of the dolomitization shows that this reflects the contact of the mound with the phreatic lens (Bourrouilh-Le Jan, 1973, 1975), related to emergence.

Geodynamic interpretation: a provisional scenario for this mound development is proposed hereafter based on geometry, stratification, geopetal structures, karst, filling of the karst, diagenesis (Fig. 2.3b) The following numbers refer to the numbers on Fig. 2.3b:

1. The Bois du Bager mud-mound began to develop as soon as turbiditic black shale deposition input decreased significantly (microsolenids can tolerate a certain amount of turbidity). The mound probably grew near the aphotic-subphotic zone transition, and certainly below storm wave base (no evidence of wave action). Stratigraphy and original position of the mound is shown by the layering of microsolenids (So on Fig. 2.3b).
2. Due both to growth and to the opening of the Arudy Basin, the mound slid and rotated 65° clockwise towards the NE (N 30°E). This rotation produced emergence of the SW flank of the mound; this emergence does not seem to reflect a sea level fall, as proposed by Canérot (1996). Emergence was accompanied by aerial and karstic erosion of the mound. The flank of the mound is eroded by karstic cavities. Karstic erosion penetrated deep into the dead mound, resulting in the formation of a large karstic cavity, which was then filled by a breccia (Fig. 2 .3b). A large part of the breccia is autochtonous, but blocks of overlying shallow water units, some as long as 2 metres, also fell down in the cavity.
3. These clasts are interpreted here as reworked parts of coarse graded supratidal to intertidal storm deposits (Fig. 2.3b). The original deposits, where they originate, are not observable in place, laterally or directly overlying the mound, but the presence of such clasts among the karstic breccia testify that near the emerged mound, supratidal to intertidal storm sediments were deposited. These flat-lying clasts indicate horizontal bedding during the karstification period, which is in agreement with the geometry of the karstification. We can suppose that the storm deposits have been eroded, broken and taken away as clasts, which fell down in the karstic cavity.
4. Renewed tectonic activity led to an anticlockwise 125° rotation of the mound towards the SW (N 210°E). As a result, the mud-mound collapsed and slid down into the deep black shale basin, where it was buried, forming a large olistolith (the mound is covered up by, and embedded into the black shales). The slumping resulted in the formation of a slump breccia (on the base of the mound, 4, Fig. 2.3b), and of the two intrusive black shale layers, parallel to the basal surface of the mound (4, Fig. 2.3b). These black shale layers are deformed and contorted.
5. The dead mound was then buried in the black shales and later in the basin evolution, oil migrated into the karstic cave (5, Fig. 2.3b).
6. The thermal development which matured the oil also seems to have been responsible for exfoliation of the karstic cave wall, above the reworked "pebbles" (6, Fig. 2.3b). Fragments of exfoliated cave wall are observed lying perpendicularly to the

basal clasts. White calcitic cements filling the pores between exfoliated wall fragments may also be of hydrothermal origin.

7. The Pyrenean orogeny folded the whole area, producing a N 120°E subvertical schistosity in the black shales and fracturation of the mound

Hydrocarbons: the Aptian karsts that affect the mud-mounds produced dissolutions which enhanced the permeability (and thus the productivity) of subsurface analogs (Bonnefond field). The Clansayesian reef intervals also represent one of the plays of the Aquitaine basin (Gaujacq diapiric oil shows in the Arzacq Basin, Mimizan North oil field in the Parentis Basin).

The hydrocarbon filling of the Bois du Bager mud-mounds karstic cavities is a good example of how HCs migrate and how they can concentrate into mud-mounds or intrabasinal high-porosity anomalies.

For further information: Canérot and Lenoble (1991), Canérot (1996), Bourrouilh (2000)

Stop 2.4:

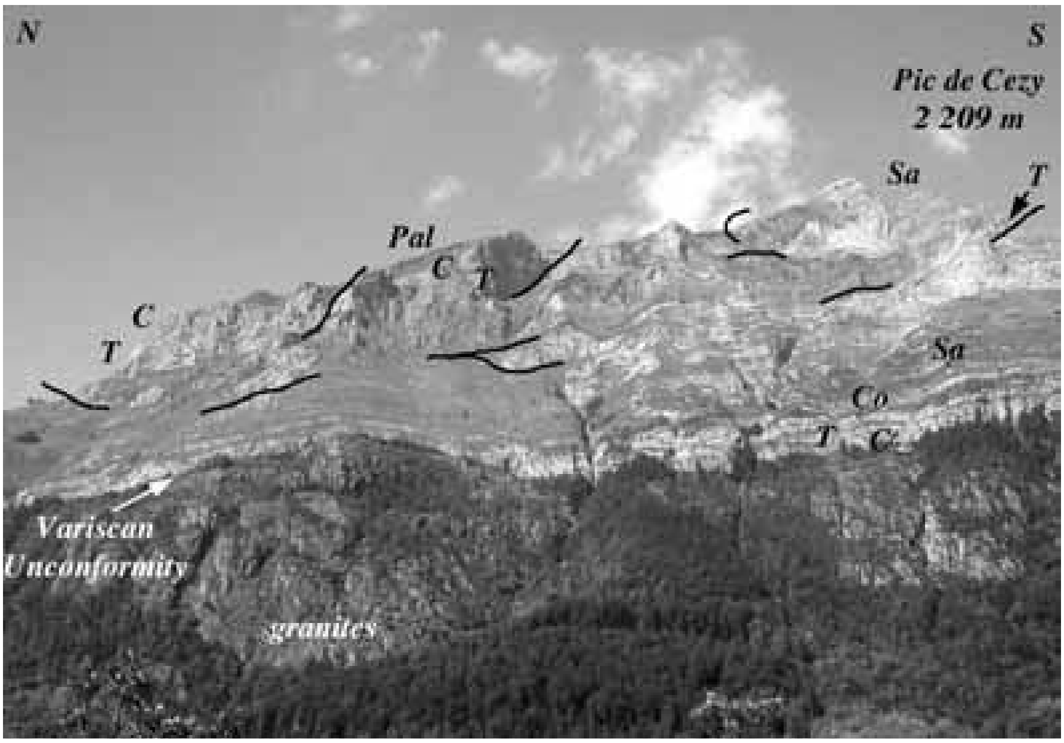
Cezy unconformity and stratigraphy of the Iberian Upper Cretaceous limestones or "calcaires des cañons" or "calcaires des Eaux Chaudes" (fig. 2.4a and 2.4b).

Introduction: crossing the North Pyrenean Fault at Laruns, we now drive along D934 through the Paleozoic basement of the high chain and the Iberian plate. Passing the crossroad with D 918 to Les Eaux Bonnes and Aubisque pass, to the left the road cuts metamorphosed Devonian limestones; to the right, a scenic view shows the ice-age Ossau Valley. When the road begins to turn left, and before entering a tunnel, it is possible to observe, on the right side, the deep sub-glacier valley of the Ossau river, which cuts the Devonian limestones in a deep canyon; meanwhile the base of the glacier is clearly observable above, with its characteristic U-shaped profile.

The road will cross in the forest several south-verging thrusts near Les Eaux Chaudes.

Access: crossing the village of " Les Eaux-Chaudes ", so called for its warm thermal waters related to the North Pyrenean Fault, we will Stop on D 934 to observe the Cezy Upper Cretaceous unconformity (figure 2.4a).

Figure 2.4a - Southverging Thrusts and Folds of Les Eaux Chaudes and panorama of Cézy. Pal= Paleozoic, T= Triassic, C=Cenomanian, T=Turonian, Co=Coniacian, Sa= Santonian



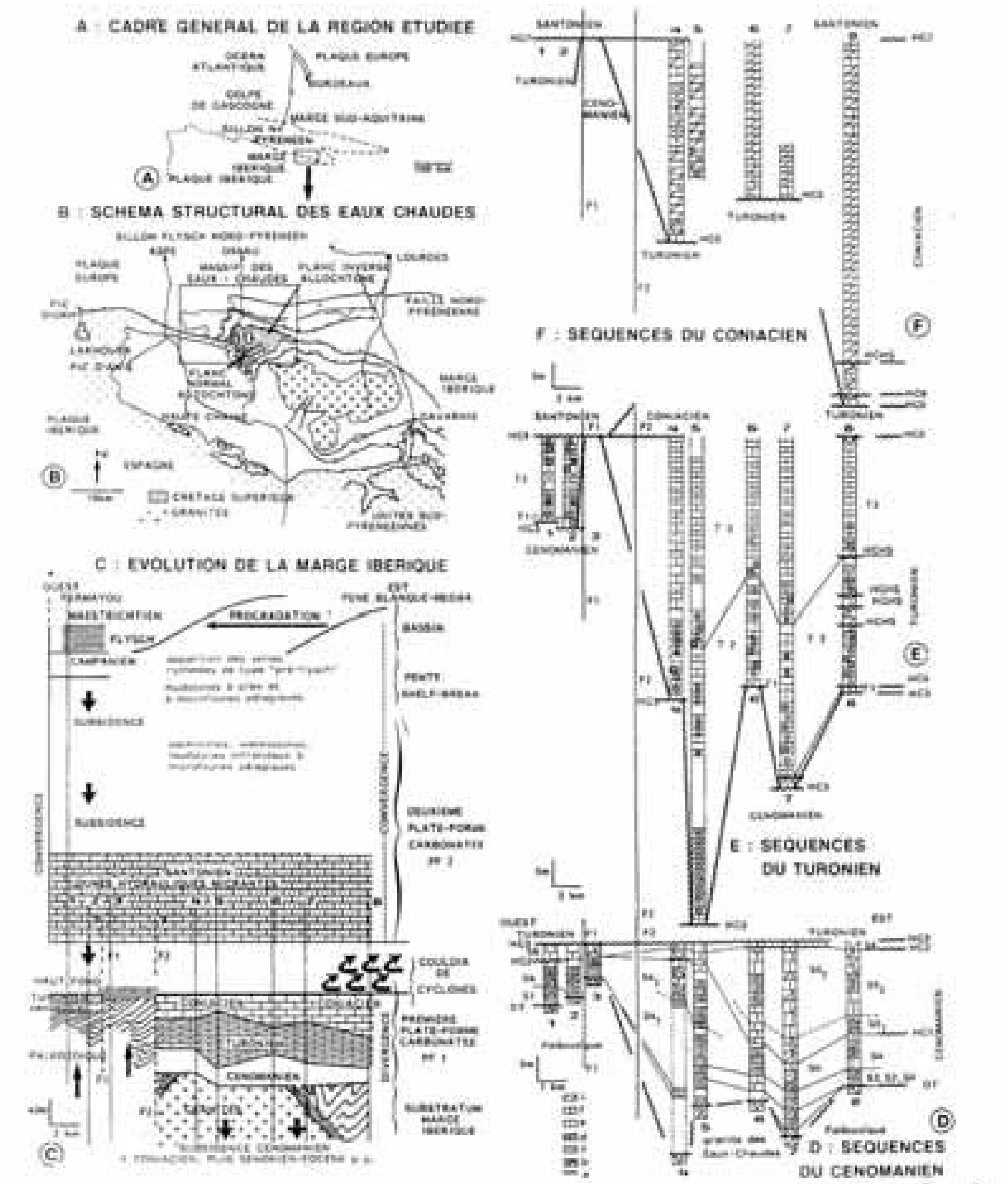


Figure 2.4b - Stratigraphy and geodynamic interpretation of the Iberian margin and of the " Canyon Limestones " or "Calcaires des Eaux Chaudes", at locus typicus. A, B = location and structural sketch ; C= Iberian Margin evolution ; D= Cenomanian ; E= Turonian ; F= Coniacian. From R. Bourrouilh and M. Alhamawi, 1993.

Structural landscape (figure 2.4a): the High Chain is there composed by a huge Paleozoic series, intruded by the Variscan granites of Les Eaux-Chaudes, which metamorphosed the surrounding rocks. Observing the panorama, we see the granites, abruptly eroded and overlain by the transgressive Cenomanian shallow water limestones, followed by Turonian and Santonian (yellow) limestones. This autochthonous

Upper Cretaceous series is overlain by a thrust, composed by a slab of evaporitic Triassic at the base, followed upwards by a thick series of Upper Cretaceous limestones. A second Upper Cretaceous series, still resting on evaporitic Triassic, is thrusted again over the underlying thrust, forming an inverted flank of a south-verging recumbent fold. The inverted Paleozoic basement form the pic de Goupey (2,209m) just above the parking area.

Interpretation: when the Jurassic Aquitaine platform broke up in response to the opening of the Bay of Biscay, the Iberian plate remained emerged in this area up to the Cenomanian (-96 Ma). In large areas such as the Cezy Mountain, the emerged Iberian basement was eroded. Huge erosion accompanied by fluvial transportation led to deposition of thick fluvial and deltaic conglomerates (1 km thick) called the Mendibelza conglomerates. Such fanglomerates deposited on the northern margin of the Iberian plate, from here westwards to the Bay of Biscay. During the Cenomanian, the sea came again in a general transgression. In this period, the Iberian plate was located south of its actual position, in an intertropical latitude. Upper Cretaceous shallow water limestones deposited directly on the Iberian plate basement, from Cenomanian to Santonian, sealing the pre-Cenomanian erosional continental surface of the Iberian plate or resting above the Eaux-Chaudes granites. During the deposition of these Upper Cretaceous shallow water limestones, the Iberian northern margin was also the margin of the deep (-3,000 m deep ?) North-Pyrenean basin which developed north, along a N 110° axis.

Sedimentation of the Upper Cretaceous limestone occurred in a shallow sea (figure 2.4b), in an environment quite similar to the modern Bahamas and, like in the Bahamas, the Eaux Chaudes limestone or " limestone of the cañons " exibits sedimentation features from hurricane trails (Bourrouilh and Alhamawi, 1985). However, as the tectonic compression between Iberian and European plates progressed during the Late Cretaceous, the South Pyrenean and North-Pyrenean foredeep basins developed while the Iberian plate differentially sunk westwards. As a consequence, deep calcareous flysch facies appear here during the Late Santonian-Campanian announcing compression in the West (Bourrouilh and Alhamawi, 1985).

For further information: Ternet (1965), Bourrouilh and Alhamawi (1993).

Stop 2.5:

Artouste: the South-verging Folds and Thrusts of Les Eaux-Chaudes: figure 1.5 and Plate 1, fig.1

Introduction: the Paleozoic basement of the Iberian plate was folded during the Variscan orogeny and then intruded by granites. Directly on this continental basement the Cenomanian transgressive sea deposited shallow marine limestones and this sedimentation continued during a major part of the Late Cretaceous. An incipient compression began here during the Senonian: the future High Chain downwarped, the shallow water limestone facies deepened and was overlain by deep water flysch facies. Tectonic compression migrated westwards and provoked the tectonic inversion of the northern margin of the South Pyrenean basin, and a series of successive basement + cover south-verging thrusts.

Access: from the bank of the Fabrèges dam reservoir we will take the telepherique (cable-car) up to the Artouste ski resort. A short walk will bring us to the top of a Paleozoic crest.

Stop 2-5a:

in this panorama to the south (Fig. 2.5a), the highest peak of Pic du Midi d'Ossau (2,884m), springing out of the landscape to the West, is constituted by a ring dyke surrounding a large caldera (about 7 km in diameter) of an Upper Carboniferous-Permian volcano. Later, the caldera was truncated by Upper Variscan faults and then, during the Pyrenean orogeny, the northern half of the caldera and of the ring dyke were thrusted to the south; post-glacial erosion as well as recent uplifting of the Chain produced the Pic du Midi d'Ossau (F. Bixel *et al.*, 1985).

The high part of the Ossau Valley clearly shows a glacial morphology, created by the Quaternary Ossau glacier, now occupied by the Fabrèges reservoir.

Stop 2-5b:

crossing the crest, we now go downhill and have a northward panorama on the Eaux Chaudes massif: see front cover, figure 1: over the Paleozoic basement of the High Chain (right), intruded by Upper Variscan granites, the Upper Cretaceous limestones rest unconformably and are overlain by the large south-verging thrust sheet of Les Arcizettes. Further to the north, this thrust sheet is also overlain by another south-verging thrust sheet of Gourzy, which is an overturned fold (Ternet, 1965, Alhamawi, 1985).

For futher information: Ossau peak: F. Bixel *et al.*, 1985, Eaux Chaudes massif: Y. Ternet, 1965, M. Alhamawi, 1985, R.Bourrouilh and M. Alhamawi, 1993.

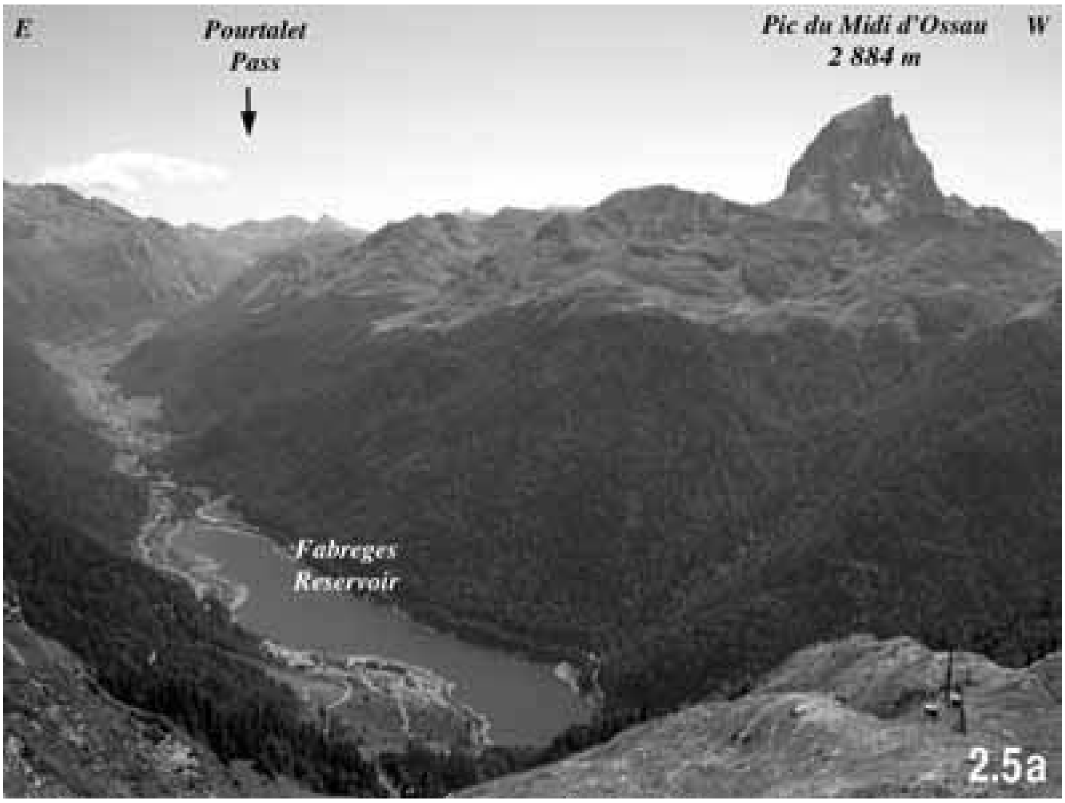


Figure 2.5a - Southverging Thrusting of the Late Paleozoic Ring Dyke of the caldeira of the Pic du Midi d'Ossau.

Stop 2.6:

the Benou Plateau : thrusts and folds in the Ossau Valley: figure 2.6a and 2.6b.

This Stop represents a panoramic view of the Ossau Valley in the North Pyrenean Zone. It is the domain of the Béarn Ranges dominated by E-W structures (Mailh Arrouy, Barescou syncline, Moulle de Jaout anticline …). Nevertheless, the Ossau Valley marks the occurrence of transverse structures (lateral ramp),

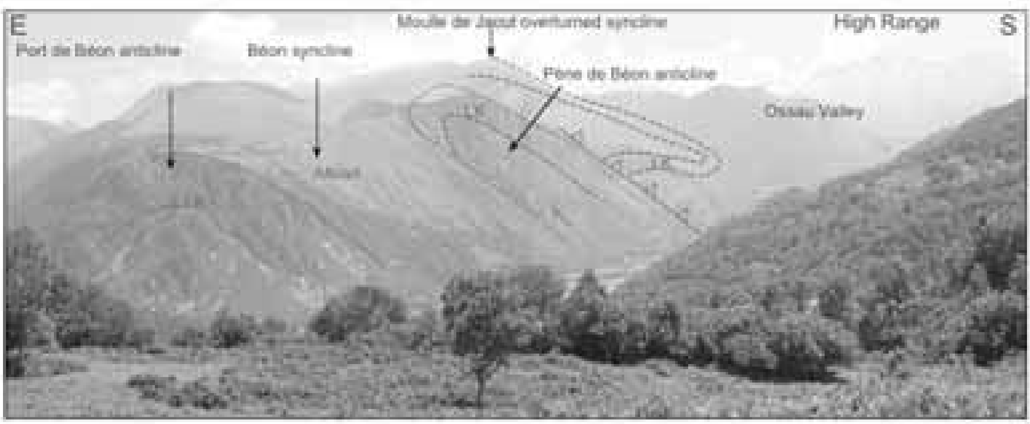


Figure 2.6a - The Benou Plateau : thrusts and folds in the Ossau Valley.

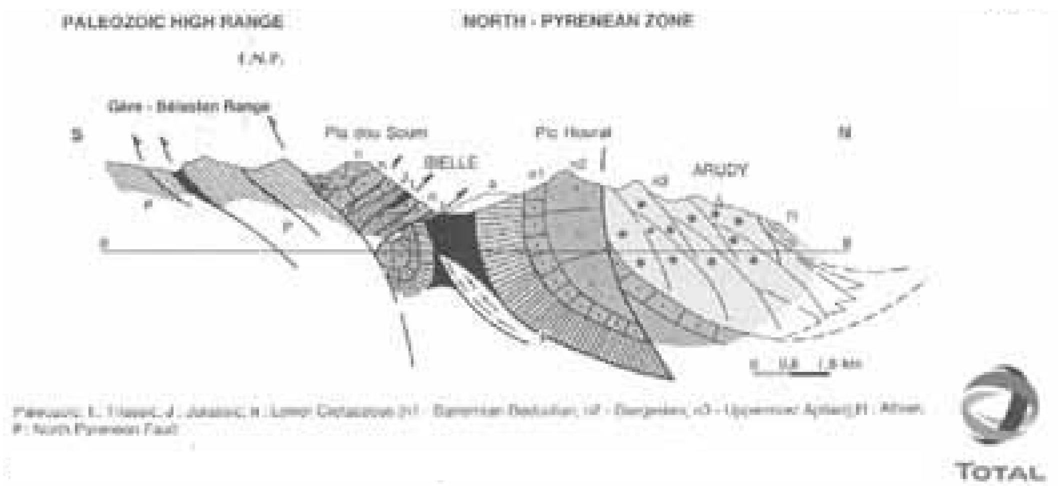


Figure 2.6b - Structural cross-section of the N-Pyrenean Zone in the Bielle area, Ossau Valley, (Lenoble and Canérot, 1992).

related to inherited Cretaceous normal faults.

Access: head towards the eastern end of the Benou plateau above Bielle. Take the trail that heads north, up to the vulture feeding grounds. Park around the high-voltage electric post.

Structural landscape: the glacial topography of the middle Ossau Valley can be observed here from the plateau on the Benou lateral moraine. The view of the southern slope of the valley shows the following, from south to north (Fig. 2.6a and 2.6b):

The Pyrenean High Chain, with its unconformable Upper Cretaceous cover deformed in the complex of the “Eaux Chaudes” recumbent folds (Pic de Ger, especially, see Stop 2.5b).

The North Pyrenean Fault (NPF), striking east-west, which is subvertical and outcrops in the Aste-Béon Triassic strip and in the Louvie pass.

Finally, ***the North Pyrenean Zone,*** detailed by the Jaout Albian-Aptian folded series (Jaout syncline, the Pène de Béon faulted anticline) and the reverse monocline series (Lias to Albian) dipping 70° towards the south, in Rey Mountain. This series is faulted against the Port de Béon Aptian units (Port de Castet fault). This fault zone consists of Triassic ophite horses mixed with Paleozoic horses.

To the east, the Rey Mountain series outcrop again on the crest of the Izeste Woods. The twisted series dips towards the north. The extension of the folded Jaout complex can be found in the five imbricate slices of Pla dou Soum. The latter constitute the detached cover of the Paleozoic sheet of Gère-Bélesten. The entire group is thrust to the northwest over the folded WE, sub-vertical structures of Bergoueits Woods. The flexured thrust fault, called the Ossau transverse thrust, strikes N30°E (Lenoble and Canérot, 1992).

Interpretation: the Ossau thrust clearly cuts the south-verging Benou thrust front (and that of the Mailh Arrouy). Its late activity has a W-E right-lateral strike-slip component. It has been established that the fault re-used an ancient N30° normal fault, active during the Liassic, the Early Kimmeridgian and the Albian. That fault bordered the western edge of the Ossau Horst. Lenoble and Canérot (1992) have also demonstrated that the Pic de Lauriolle diapir (second generation ridge), marked by breccia infill (collapse breccias) to the south of the Lassourde basin, is located exactly at the intersection of that N30° Albian Ossau fault and the Ibech fault (a W-E fault running along the Haute Soule first generation diapir ridge). This again demonstrates the fundamental role that is played by extension-related (transtension) Mesozoic paleogeographic structures in the subsequent development of compressional structures (transpression).

For further information: Lenoble and Canérot, 1992; Casteras, 1974; Henry, 1987.

Stop 2.7:

Mailh Arrouy view from Barescou syncline (Jurassic reservoir and source rocks) : figure 2.7a and 2.7b.

Access: at Escot, take the road that leads along the Barescou Valley to the Ossau Valley through the Marie Blanque pass. Park near the bridge which

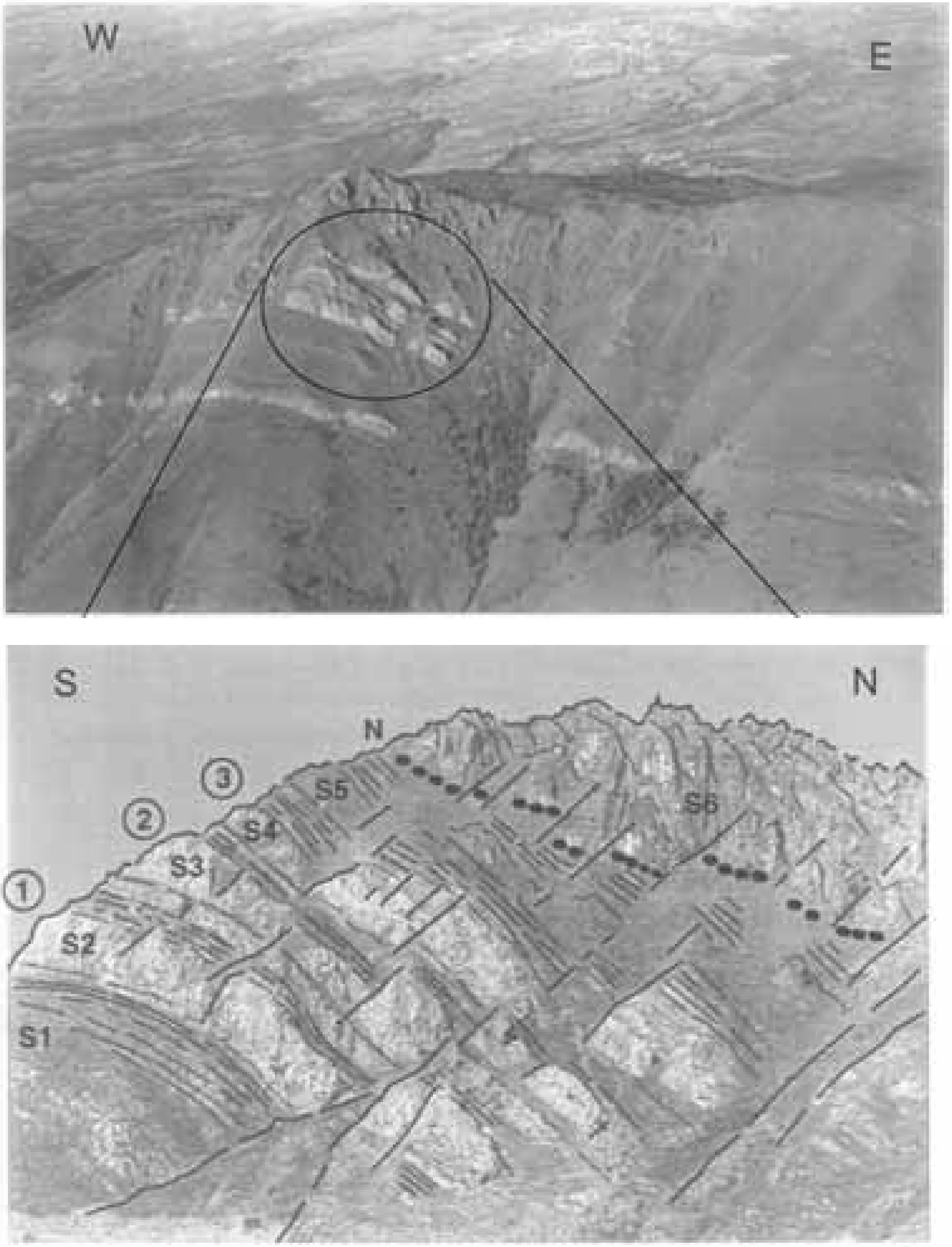


Figure 2.7a and b - Sequence Stratigraphy of the Middle Jurassic of the Mailh Arrouy (Total).

crosses over the Barescou stream.

Structural landscape: The topography of the Mail Arrouy reveals good outcrops of the Jurassic series that are thrusted southwards with a sole of Triassic evaporites and shales onto the Albian flysch of the Barescou syncline (Fig. 2.7a and 2.7b). This series shows, from base to top, the Lower and Middle Liassic limestones, the Upper Liassic marls, the

grey "filament limestones" of the Aalenian, lower dolomites called the "Mailh Arrouy dolomites" made of recrystallized oolitic shoals and lagoonal facies.
The thrust fault dips 20° to 30° northward, at the base of the Lower Liassic limestones, and laterally at the base of the Triassic.
Sequence stratigraphy: the Bathonian limestones were divided into 3 third-order sequences (S2, S3 and S4, shown on Fig. 16 a and 16 b). Observe the sequence limits in the grey scarps of the landscape. Each sequence limit begins with a coarsening upward series (highstand system tract, followed by lowstand system tract) and then a fining upward series (transgressive system tract). The grassy slopes separating the limestone cliffs comprise the maximum flooding surfaces of the sequences. The overlying black dolomites were cut into two complete series (S5 and S6). Note the rapid development of the highstand system tract in S6 (pinnacle of the Mailh Arrouy).
Hydrocarbons: the Mailh Arrouy dolomites are thick and porous (especially the oolitic shoals), and constitute good hydrocarbon reservoirs. They can be compared to the dolomitic levels of the Meillon field. The Liassic marls constitute one of the source rocks in the Aquitaine basin, especially in the eastern sector (Comminges basin – Saint Marcel field).
The geometry of the monocline (and particularly the southern edge of the block) is also similar to the structure of the Meillon gas field.
For further information: Henry, 1987.

Stop 2.8:

Herrere : Pillow-lavas of the Albian rift stage: figure 2.8 (Moen-Maurel *et al*., 1996).
Access: south of Herrère (D 920) between Oloron and Arudy), park at the Courrèges Farm just north of the railroad crossing. Pillow-lava flows are located in the cliff beside the corn crib. A pedestrian path is located to the left of the farm gate.
Stratigraphy : the pillow-lava flows are interbedded within the Albian black marls (Fig. 2.8). They are part of a suite of volcanic textures also displaying basaltic columns and pyroclastic breccias, visible in other parts of the Ogeu Basin. The epivolcanic to mesovolcanic magmatic rocks include episyenites and teschenites (under-saturated syenites) that are exposed a few km to the west, and that are prevalent in the subsurface to the west. Here the submarine volcanic lava tubes dip 20-30 to the south and end in larger round tips to the southwest. Smaller pillows are intercalated between the tubes.

Figure 2.8 - Pillow-lavas flows of the Albian rift stage at Herrer (20 cm long pencil for scale).

Structural setting and hydrocarbons: part of the dip of the tubes was acquired during folding of the Belair Anticline whose N110 axis is located a few hundred metres to the north of the pillow-lava outcrop. The axis of the anticline plunges gently to the west beneath the Gave d'Oloron Thrust. In the 3D seismic, one can see at - 1200 m/SS a similar antiform made of low frequency markers that were drilled at Ledeuix and which produced gas (Moen-Maurel *et al*., 1996). The reservoir consisted of chilled margins of episyenite sills. The source rocks were the Albian black marls that were affected by magmatism. These marls acted as source rock and seal. The Albian marls can be over a km thick, and thus provide the best seal over most of the Aquitaine fields (Lacq, Meillon ...).
Tectonic significance: the southward flow of the tubes indicates that the normal (extensional) fault that favoured the uplift was located to the north of the outcrop. Volcanics and episyenites represent the magmatic expression of the mid-Cretaceous rifting between Iberia and Europe. However there is no sign of oceanic crust development, nor of margin separation. In the North-Pyrenean domain the rift was limited to half-grabens en-echelon basins, and basins controlled by diamond-shaped fault patterns. Rift axes that are located over the Pyrenean axial zone (eroded down to the basement, or likely beneath the thrust

sheets, such as for the North-Pyrenean Fault) would be better candidates for the hundred-km long, left-lateral strike-slip displacement that occurred between Iberia and Europe through the Mid-Cretaceous.

DAY 3

from Oloron Ste Marie to Saint Jean Pied de Port

Leaving Oloron by the southern road N 134 to Col du Somport and the Spanish border, we enter the Aspe Valley, along which runs the Gave (river) d'Aspe. Along the valley, there also exists a railway linking Oloron and the Spanish railway station of Canfranc. This railway was cut a long time ago and discussions between Spain and France are still taking place, to restore it and the railway link. Entering the valley and the mountain, railway bridges, tunnels are visible on both sides of the N 134.

Stop 3.1:

Pont d'Escot - Ste Suzanne Formation (seal) : figure 3.1a, 3.1b, 3.1c

Access: 15 km south of Oloron, on the N 134, we will Stop in the parking lot, just after passing one of these railway bridges, the Pont d'Escot.

Stratigraphy, structural geology: from south to north, the road cut reveals the Aptian sedimentary evolution of the Sarrance anticline's sub-vertical northern limb. The rapid transition from Bedoulian Ste-Suzanne marls, to Gargasian (Urgonian) limestones (with Toucasia) is nevertheless progressive within a highstand environment. Further up-section, the Urgonian facies is prevalent within the lowstand of the next sequence (Fig. 3.1). The sequence limit probably corresponds to a sedimentary hiatus, which can be observed at the base of the thick beds with abundant Rudistids. The schistosity is particularly well marked within the marly beds. Further south, and upslope, the marls are interbedded with limestones, indicating that the marly series comprises two distinct groups (Fig. 3.1b). This intermediate limestone and dolostone is a regional seismic marker in the foothills and foreland sub-surface.

Hydrocarbons: the Sainte Suzanne marl formation forms the upper or direct seal of many hydrocarbon fields in Aquitaine (Vic-Bilh, Pecorade, Lacq), and Meillon gas field, located 30 km SE.

For further information: Canérot, 1964; Castéras, 1974.

Stop 3.2:

Sarrance roadcut - Jurassic dolomites (reservoir formation).

Access: again driving south on the N 134 the Stop concerns the road-cut, three kilometres north of Sarrance. Park at the picnic grounds. Park on the side of the road-cut, please, put on the warning lights.

Sequence stratigraphy: the road cuts across the Jurassic rocks of the sub-vertical northern limb of the Sarrance anticline (fig. 3.1c). The road-cut reveals some sedimentary key points which have given rise to a proposed sequence stratigraphy (Canérot *et al.*, 1990) for the North Pyrenees dolomite series. This study concerns the stratigraphic interval from the Bathonian to the Upper Callovian in the Bearn Ranges. The set was divided into five third-order sequences, termed DSI to DSV (Lenoble, 1992). Only part of sequences DS III to DS V can be observed here (Fig. 3.2).

<u>Sequence DS III</u> (30 m): finely bedded dolomites (outer shelf), grading into oolitic (offshore bars) and finally reef deposits (colonial corals, fig 3.2b). The series in this first unit represents the highstand system tract. The contact with sequence DS IV is not distinct.

<u>Sequence DS IV</u> (40 m): bioclastic dolomites, thick and then finely bedded (offshore bars), characterizing the lowstand system tract. They are overlain by a striking ferruginous discontinuity, overlain by bioclastic horizon, interpreted as a transgressive surface. The oolitic bars (shoreface) and the overlying striped dolomites (inner lagoon) probably correspond to a highstand system tract. The maximum flooding surface has not been identified.

<u>Sequence DS V</u> (15 m): thick dolomites (shoreface bars) overlain by striped dolomites (inner lagoon) correspond to lowstand and highstand system tract, respectively. The transgressive interval and the maximum flooding surface have not been clearly observed.

Because of intense fracturing, it was not possible to make a precise reconstitution of the described sequence. Stratigraphic determinations (Bathonian for DS III, Callovian for DS IV and DS V) are based on the correlation scheme between the Béarn dolomite series and the Basque series with abundant Ammonites, which was not subjected to epigenetic processes.

Hydrocarbons: the dolomite complex represented in the area was designated as the Mailh Arrouy formation, and considered as the equivalent of the

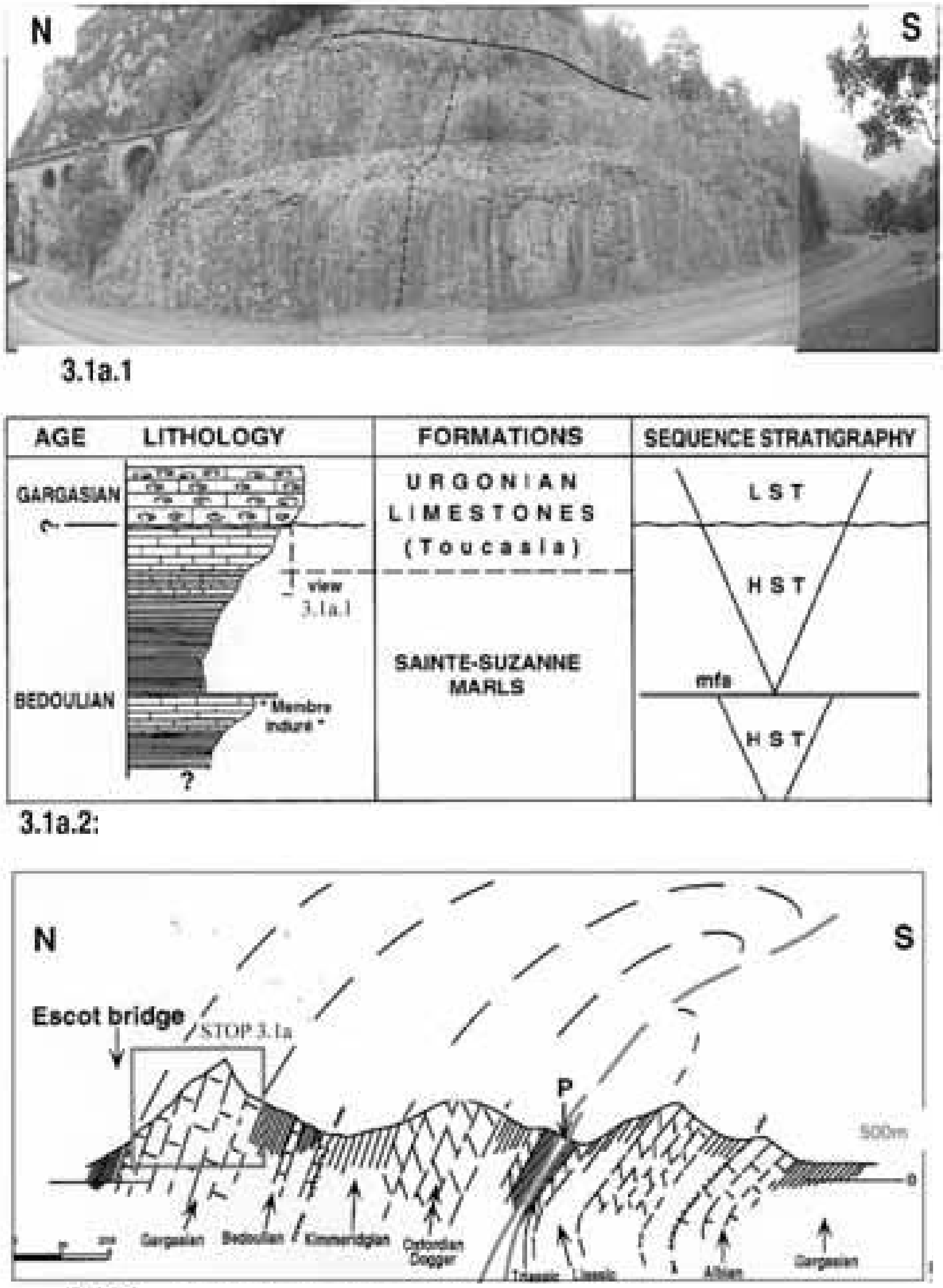


Figure 3.1a - Pont d'Escot : Sainte Suzanne Aptian formation Sequence Stratigraphy .

Meillon gas-bearing dolomitic formation. Further research has revealed that the Meillon formation is probably much younger (Upper Oxfordian to Lower Kimmeridgian) than the Sarrance sequence (or than the Mailh Arrouy sequence).

For further information: Canérot *et al*., 1990; Lenoble, 1992.

Stop 3.3:

Bedous : the southern margin of

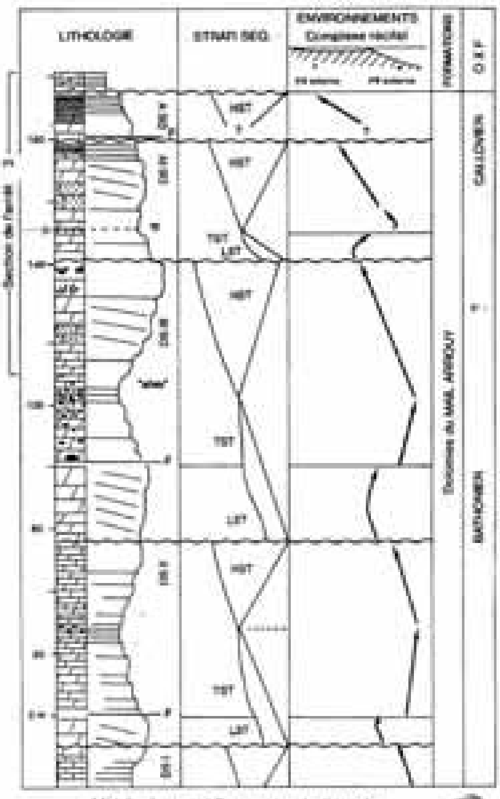


Figure 3.2a - Middle Jurassic Sequence Stratigraphy (Bathonian-Callovian), Northern side of Sarrance anticline, (Canérot and Lenoble, 1993).

(Fig. 3.3). In the Bedous Valley, the Triassic includes Muschelkalk limestones, Keuper shales and ophites in slivers of duplexes that were thrust to the south (visible along the halpin road). These rocks were thrust onto the flysch of the Upper Cretaceous cover in the High Range, around the village of Lées (Fig. 3.3). Further north, the crest of the Layens Mtn shows a thick sub-horizontal Jurassic and Lower Cretaceous series belonging to the normal limb of a recumbent anticline verging to the north. To the east of the viewpoint the Bergon Mtn shows a hanging recumbent syncline with a top-to-the-north vergence.

Geodynamic interpretation: the Bedous area marks the mid-point of the double vergence that

Figure 3.2b - Colonial corals in Sequence DIII.

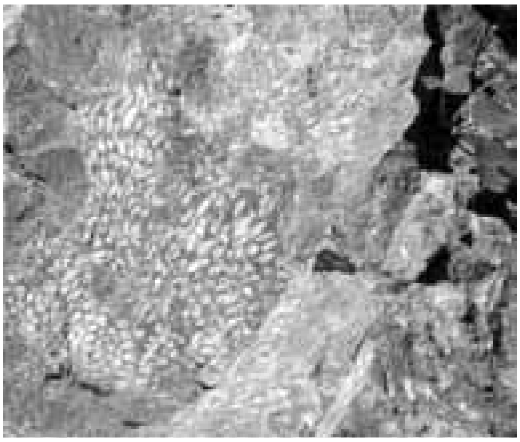


Figure 3.3 - Structure of the High Range at Bedous, Aspe Valley, (Total). Fl= Upper Cretaceous Navarella flysch (Senonian), UK= Upper Cretaceous Canyon Limestones (Cenomanian- Lower Senonian), Tr= Triassic ophites.

Figure 3.4a - Megaturbidite at Osquich Pass.

Figure 3.4b - The folded Megaturbidite in the landscape.

characterizes the Pyrenean mountain range. Some authors traditionally set the North-Pyrenean Fault at this location but this hundred-km long strike-slip fault is likely concealed beneath the south-verging thrust sheets, and would occur further north in depth, probably where some earthquakes still originate nowadays.

To the South, subduction of Iberia beneath Europe was so large in magnitude that only south-verging thrusts and nappes can be observed. To the north, depending on the original dip of the rift faults, inversion produced either north-verging or south-verging folds. The rift structures are located in a retro-arc setting, which limited its inversion and its uplift (and erosion), thus permitting the petroleum system to exist till today.

Stop 3.4:

Osquich Pass: Upper Cretaceous Megaturbidites: figure 3.4a, 3.4b, 3.4c and 3.4d.

Access: from Stop 3.3, drive back to Accous and now drive N 134 back to Oloron. At the entrance to Oloron, drive on D 936 to Navarrenx and Sauveterre de Béarn. 11 km from Oloron, take on the left the D 25 to l'Hôpital St Blaise and Mauléon-Licharre and then the D 918 to Col d'Osquich.

The Stop is located on the road from Mauléon to Larceveau, D 918, about 1km after the Osquich pass, at the first curves in the road, just before the village of St Jean Ibarre.

The Megaturbidite concept was first established by Soler and Puigdefàbregas, 1970; Rupke, 1976, for huge gravity deposits occurring among Eocene deposits of the Southern Pyrenean basin. Later, Johnson and al (1981) studied the South-Pyrenean basin and Labaume *et al*. (1983) studied these deposits in detail. Megaturbidites were later identified among the Cretaceous North Pyrenean sediments and were studied in detail by Bourrouilh, Coumes and Offroy (1984) and Offroy (1984) who interpreted the infilling of the North-Pyrenean cretaceous flysch basin, demonstrating that:

- two large megaturbidites occurred in the basin. One of them extends over more than 90 km, and varies from 60m in thickness, near Pau, to 20m in its western part, near the Atlantic Ocean.
- Megaturbidites originate as evolutionary mass-flow-megaturbidites (figure 3.4d)

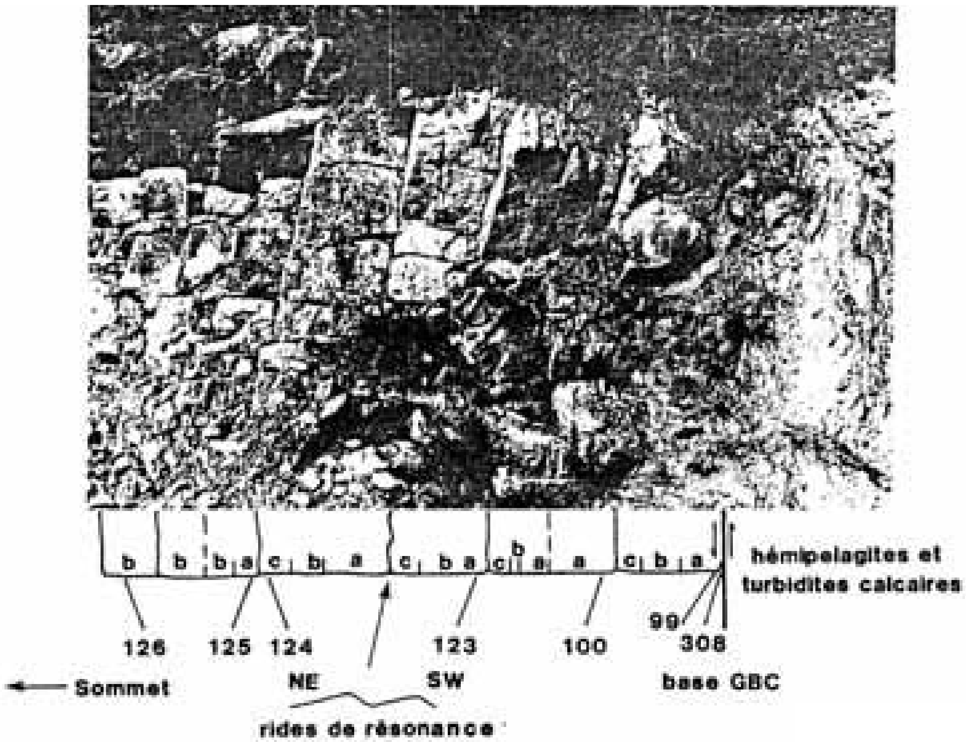


Figure 3.4c - Sedimentological interpretation of the Megaturbidite, from Offroy, 1984.

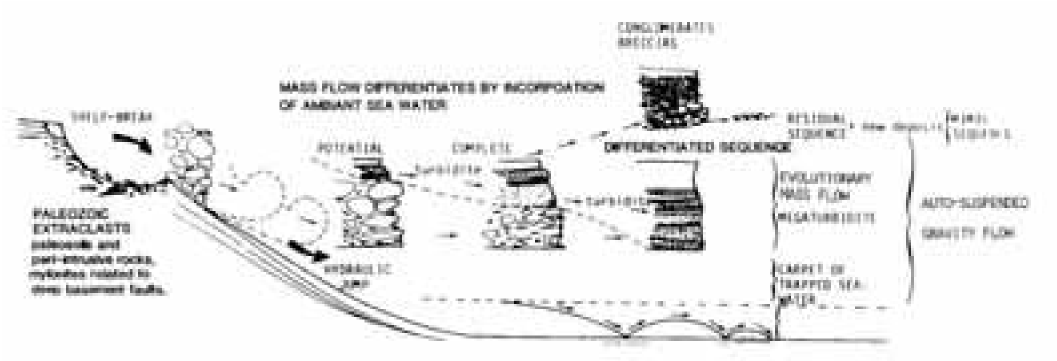


Figure 3.4d - The Evolutionnary-Mass-Flow-Megaturbidite concept, R. Bourrouilh et al., 1984, R. Bourrouilh, 1987.

Structural landscape: the panorama shows a massive turbidite, 40 m thick, which outcrops on the right side of the road, but also in the landscape, on the left side of the road.

Interpretation: studied by Bourrouilh and Offroy (1983), Offroy (1984), this megaturbidite is constituted by successive but amalgamated Bouma's intervals, which demonstrate that the turbiditic flow was internally stratified, each strata having a specific kinetic energy. Vibratory ripples can be also observed in interval surfaces (figure 3.4c).

On the left side of the road, and if the light is oblique enough, we can observe large Pyrenean kink-folds affecting the flysch and particularly the Megaturbidite (figure 3.4b).

For further information:

Bourrouilh and Offroy, 1983, Bourrouilh, Coumes and Offroy (1984), Offroy (1984),.

Stop 3.5:

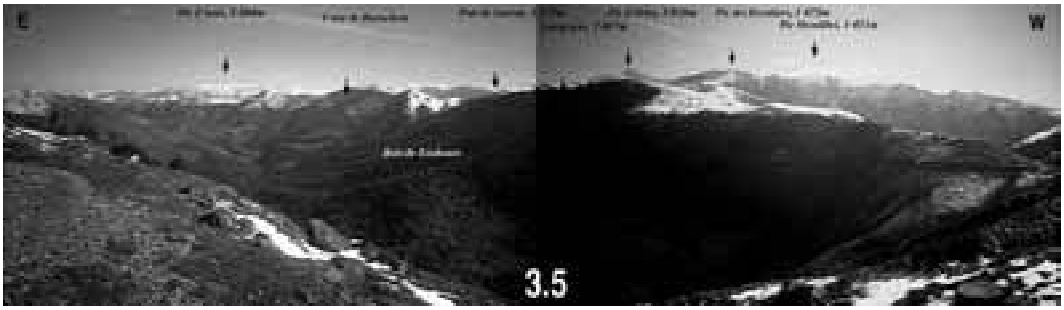
Ahusquy Pass: Structure of the tilted blocks and of the Iberian margin : (figure 3.5).

We drive back to Mauléon by the D 918 road.

Introduction: from Mauléon to the small village of Aussurucq, the road cuts the Upper Cretaceous flysch of the N-Pyrenean trough. From Aussurucq to the Ahusquy pass, the road will cross a little-deformed tilted Mesozoic block of the Iberian margin (Arbailles block), whose northern edge is deformed in a NE verging thrusted anticline. This N110°E anticline shows a complete series, from the Upper Carboniferous, unconformably overlain by a typical German type Triassic, followed by a calcareo-dolomitic Jurassic. The tilted block was part of the southern margin of the N-Pyrenean Lower Cretaceous rift which extended the Bay of Biscay. This history is clearly registered in the erosional surface affecting the Upper Jurassic limestone, locally showing bauxitic remnants. The progressive rotation of the tilted block is demonstrated by the unconformable fan-like dips of the Neocomian continental sands, the Hauterivian-Barremian shallow water limestones and finally by the first large transgressive marine deposits of "the marnes de Ste Suzanne". Then, the shallow marine partly reefal limestones of the Urgonian were deposited. These competent limestones form the frame of the tilted block or the " Arbailles " block.

The Arbailles anticline is followed to the south by a

Figure 3.5 - Panorama from Ahusquy Pass.



large syncline, formed by the Upper Aptian-Albian black shales, in what was the downthrown part of the Arbailles block.

Access: on the way to Mauléon by road D 918, we will take the small road D 147 on the left to Mendy and to Aussurucq and then through the Forêt des Arbailles massif, up to the Ahusquy pass.

Structural landscape and Interpretation: Ahusquy pass is situated on the Upper Aptian-Albian black shales, which form the core of the Apanicé-Bois de Zouhoure syncline.

The southern flank of this syncline is verticalized. The underlying Urgonian limestones, which form the frame of the Arbailles anticline to the north, crops out here again, forming a continuous rocky ridge, reaching an elevation of 1,265 m at the Pic de Behorleguy. The Jurassic and the evaporitic Triassic outcrop on the southern side of the syncline, in a narrow ridge separating it from the Mendibelza conglomerates; this N 110°E zone is generally interpreted as a major Pyrenean Fault, extending to the West Souquet's "Faille de Bigorre" (Souquet et Debroas, 1980). However, this hypothesis does not fit with structural and drill-hole data and could be discussed during the field trip. Immediately south of this fault zone, the thick black Albian-Vraconian conglomerates of the Mendibelza (this name means: black mountain in Basque) range are directly discordant on the Paleozoics. In the background of the panorama, the Upper Cretaceous "calcaires des cañons" or "calcaires des Eaux Chaudes" present a sheared unconformable contact with the Paleozoics of the High Chain (see Stop 2.4a and 2.4b, here above), which culminates at the Pic d'Anie (2,504 m high, to the left) and in front of us, at the Pic d'Orhy, at an elevation of 2,017m.

For further information: see the geological map at 1/50 000 n° XIV-46, Tardets -Sorholus

Stop 3.6:

Lutogagné: Structure of the Lower Cretaceous tilted blocks: figure 3.6.

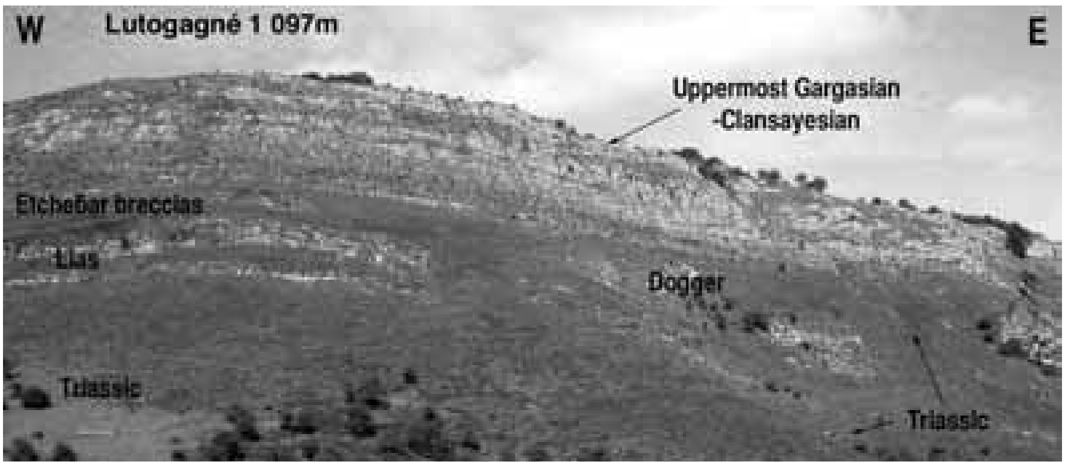
Introduction: this is a general panorama along the strike of the southern flank of the Arbailles, or the Apanicé-Bois de Zouhoure syncline.

Access: leaving the Ahusquy pass, we drive to the Iraty pass and we will progressively cross the Albian black shales of the syncline to reach the Triassic outcrops and the panorama.

Structural landscape: we Stop near the Cayolar (mountain barn) of Arhansus. Looking to the north, we now examine the southern limb of the Arbailles syncline culminating locally in front of us at the Lutogagné peak (1,097m). This southern flank is rotated vertically with the folding of the syncline, and deeply fractured in this area.

Interpretation: according to Canérot and Lenoble (1991), the Jurassic (Lower Lias to Bajocian) is covered by diapiric breccias (Etchebar collapse breccias) of Lower Cretaceous age, and unconformably overlain by an uppermost Gargasian-Clansayesian limestone containing typical Foraminifera. (Peybernès and Garot, 1984). Therefore most of the Jurassic and Barremian carbonates are absent; a reconstruction with subcrop maps indicates the presence in the Barremian to Albian of a diapiric ridge which developed along the downfaulted side of the Arhansus Fault. This ridge was then squashed by the subsequent compression between the upthrown

Figure 3.6 - Lower Cretaceous tilted blocks.



block of the Iberian margin of the North-Pyrenean rift and the Albian syncline which occupied the downfaulted half-graben of the Arbailles Block.

Hydrocarbons: the Clansayesian series regionally marks the vertical transition from the Urgonian platform (Aptian) to the Albian marl basin. This sequence has been observed in the Arzacq and Tarbes basins, where it contains reef structures which have been drilled, and are locally HC bearing and productive (Gaujacq).

For further information: Peybernès and Garot, 1984, Canérot and Lenoble, 1991

Stop 3.7:

Mendibelza mass flow conglomerates: figure 3.7.

Introduction: directly resting unconformably on the Paleozoics, the Mendibelza Albian conglomerates result from the erosion of the Iberian plate. They rework all the Paleozoic series, from at least the Ordovician to the Permian, and also, Triassic evaporites (bipyramidal quartz), Triassic ophites, Jurassic limestones (Digbehi, 1987), as well as subcontemporaneous Lower Cretaceous shallow water and reefal limestones and black shales. They also contain layers with remnants of leaves, broken branches and debris, of the coeval forests which grew up along the coast and the delta bayous. The conglomerates are interpreted as resulting from margin uplift of the North Iberian margin, as the opening of the Bay of Biscay rapidly progressed eastwards. Thus, these fanglomerates mark the passage between the rift and the early plate drift stage. They are a thick accumulation of fluvial to canyon and slope fanglomerates, rapidly slurried down on the growing divergent Iberian margin.

Access: on the road to Iraty pass

Structural landscape: several mass-flow sequences are observable, top on the left.

Interpretation: these slurried beds were emplaced by immature, i.e. proximal mass-flows, probably resulting from fluvial (deltaic) overfloods. Frequently, the top of mass-flow sequences is abruptly washed out by a high energy water current, which took off a part of the former deposited sequences, leaving isolated pebbles outcropping on the sea floor.

For further information: Boirie, 1981, Bourrouilh, Coumes and Offroy, 1984; Digbehi, 1987; Miranda Avilès, 2002.

Figure 3.7a - General view of the Mendibelza massif.

Figure 3.7b - Sedimentology of Mendibelza conglomerates (Total).

Stop 3.8:

Mendibelza conglomerates on Iraty road, figure 3.8.

Introduction: this Stop allows us to observe large olistostromes, with olistolithes of Paleozoic blocks, Lower Cretaceous shallow water limestones, including reefal limestones, and reworking pebbles and gravels of quartz, Ordovician quartzites, Devonian limestones, carboniferous radiolarites and muscovite-rich greywackes, as well as Triassic ophites and bipyramidal quartz.

Access: down Iraty pass, driving to Lecumberry on road D 18, leaving the Saint Sauveur Chapel on the right, the

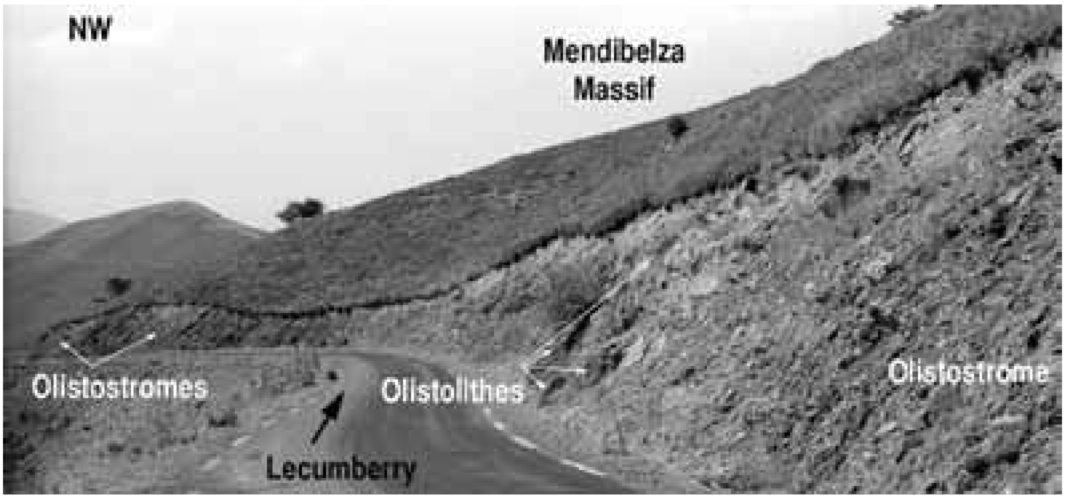


Figure 3.8 - Olistostromes and olistolithes in Mendibelza conglomerates.

Stop begins about 1 km down from the Chapel crossroad, on the first road curve (Fig. 3.8).

Structural landscape: the large olistostrome crops out all along the right side of the road, over several tens of metres, gently resting on channelized mass-flow. Large pebbles and Paleozoic and Lower Cretaceous calcareous fossiliferous olistolithes are clearly observable on the side of the road.

The other side of the road shows the Mendibelza massif and, far in the background the Atlantic coast and the Bay of Biscay.

Interpretation: the on-going growth of transtensional basins provoked inversion of the morphology of their southern margin, revealed both by the infilling of the basins by large amounts of fluvial conglomerates (Mendibelza ones) and by large olistostromes, reworking olistolithes of former deposits but also of the subcontemporaneous reefal carbonates (Boirie, 1981, Souquet and Boirie, 1985, Digbehi, 1987). These transtensional basins are similar to the modern Baja Californian divergent basins of the Gulf of California (Miranda Avilès, 2002).

For further information: Boirie, 1981, Souquet et Boirie, 1985, Digbehi, 1987, Miranda Avilès, 2002.

Stop 3.9:

tilted blocks and European-Iberian margin, 2 km NW of Lecumberry : figure 3.9.

Introduction : this Stop allows us to observe the Arbailles tilted block and the European-Iberian margin.

Access: on road D 18, about 2 km NW from Lecumberry, on the small slope of the road.

Structural landscape and Interpretation:

1. The post-rifting, syn-drifting “ European ” plate is represented here by the western structure of the Arbailles tilted-block, showing the southern limb of the Arbailles (Apanicé) syncline, culminating here at Behorleguy peak (1,265 m).

The Urgonian ridge of the Behorleguy peak is verticalized. The shallow water Urgonian carbonates rest unconformably on bauxitic deposits filling up a previous karst and were mined. This diachronous bauxite extends unevenly from here to Provence, over 800 km away, testifying to complex and polyphased emersions/submersions of the “Durancian isthmus”.

The slope down the Behorleguy peak is formed by an incomplete Jurassic series, the pre-Urgonian erosion reaching irregularly the Callovo-Oxfordian, the Dogger or even the Liassic.

The Lecumberry valley strikes through the soft evaporitic to calcareous, and arenitic Triassic.

The possible fault previously discussed at Lutogagné, must run on the right side of the valley

2. the Iberian plate is constituted here by the large amount of Albian to Vraconian Mendibelza conglomerates, unconformably resting on the Paleozoics. Far in the background, the Mendibelza conglomerates are locally covered by the Upper Cretaceous “calcaires des Cañons”.

For further information: see the geological map at 1/50 000, n° XIII-46, St Jean Pied de Port.

Figure 3.9 - Tilted blocks and European-Iberian margin.

DAY 4

from St Jean Pied de Port to Jaca

The road to Larrau and to the Port de Larrau will drive us through the Mendibelza conglomerates.

Morning:
Stop 4.1a:

Pic d'Orhy South-verging Thrust and Fold structures

Introduction : resting on the Mendibelza conglomerates, a huge series of Upper Cretaceous shallow water carbonates form the main frame of the Pyrenean High Chain. We already observed these limestones the first day, in the Ossau Valley, forming the high peaks of the Eaux Chaudes Massif. These "calcaires des cañons " or "canyon limestones" develop from Cenomanian to Lower Campanian. Then, a carbonate flysch facies up to the Lower Lutetian progressively appears.

Access: from Larrau, road D 26 climbs up to the Port de Larrau, 1,575 m high. We will Stop at 1,400 m elevation, near the thrust surface (fig.4.1a1).

Structural landscape and interpretation: looking westwards we can observe the S-shaped south-verging folds of the Orhy peak, 2,017m high, forming a souththrusted anticline, involving the Upper Cretaceous flysch, followed by an south-verging overturned syncline, formed by the Paleocene-Lower Eocene flysch (fig.4.1a2).

Here we can see one of the highest fold-and-thrust tectonic units.

For further information: see the geological map at 1/50 000, n° XIV-47, Larrau.

Stop 4.1b:

Port de Larrau: Structural Panorama

Access: drive on D 26 up to the Port de Larrau, 1,573 m high.

We are now in the heart of the Pyrenean High Chain, just below Orhy Peak and with the ample panorama of the Iberian plate in front of us.

Structural landscape and Interpretation: the High

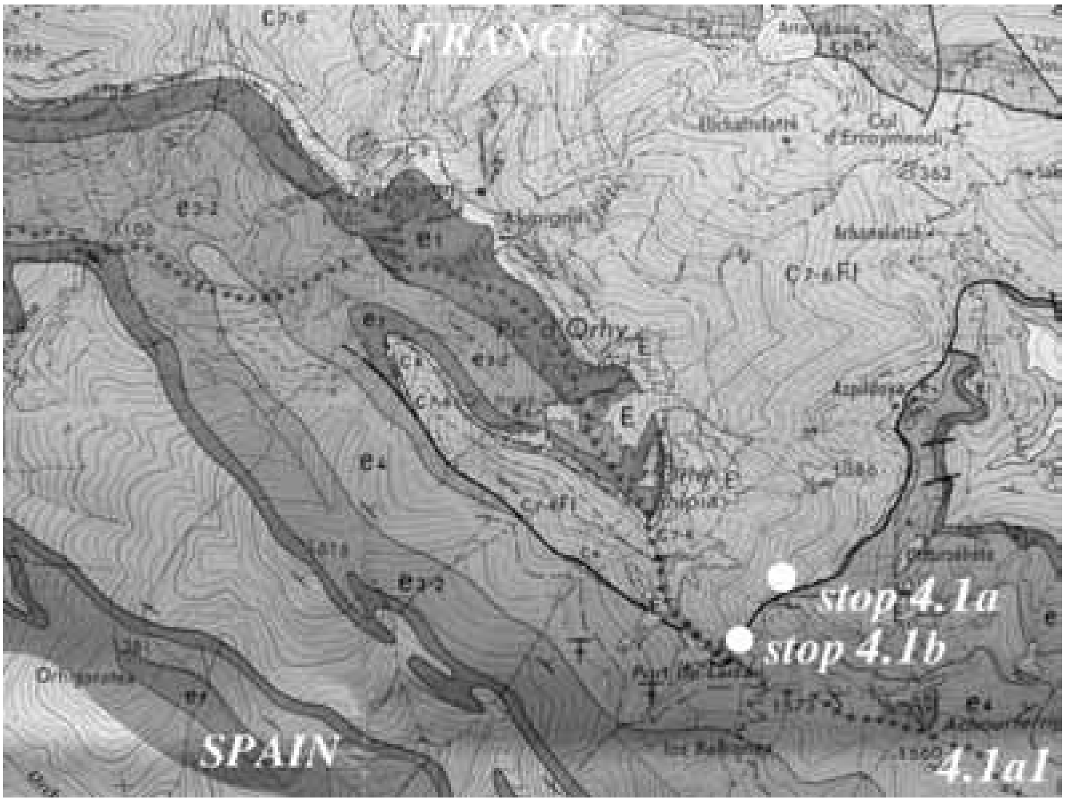


Figure 4.1a1 - Geological map of the Pic d'Orhy area, from BRGM Map, n° XIV-47, Larrau.

Chain now develops west and eastwards, the Upper Cretaceous to Lower Eocene "Calcaires des Cañons" forming the mountain range.

To the south, the large south Pyrenean basin develops, filled with Eocene flysch of the Hecho Group.

Stop 4.2a:

Igountze Thrust and St-Engrace Thrust sheet.

Access: take the D 26 back to Larrau and then along the Gave de Larrau to take road D113 on the right along the northern slope of the Ste-Engrace valley. Stop 500m east of the Kakoueta canyon road access.

Stratigraphy, structural geology: the Igountze unit consists of the Devonian – Carboniferous basement and of its unconformable Albian cover comprising the Mendibelza conglomerates. This unit is southward thrusted, almost horizontally, over the Ste Engrace thrust sheet (fig.4.2a1 and 4.2a2). The latter is also thrusted (fig.4.2a3) over the so called canyon limestones (Cenomanian to Campanian) and the Navarella flysch (Campanian and Maastrichtian). This flysch joins the unconformable cover overlying the Paleozoic basement, in the High Range further to the south.

Flattened and sheared (boudinage-affected)

Figure 4.1a2 - Aerial view of the Pic d'Orhy, courtesy of Vivien De Feraudy (Total).

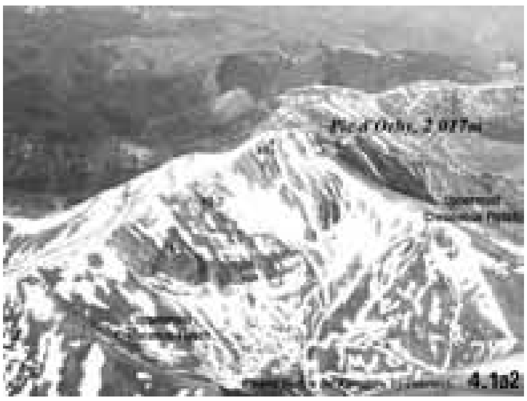


Figure 4.2a1 - Geological map of the Saint-Engrâce area, from BRGM map n°XIV-47, Larrau.

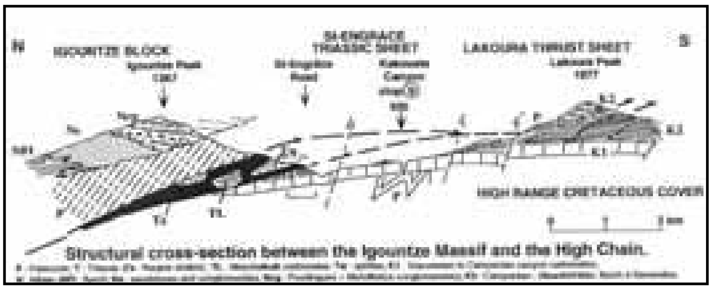


Figure 4.2a2 - Structural cross-section between the Igountze Massif and the High Chain (Total).

sedimentary breccias occur at the base of the flysch. Laterally the internal shearing of the Senonian series beneath the Igountze thrust is characterized by a north-dipping duplex shear belt, located at the limestone-flysch contact. To the south, Lakoura Peak is made of a klippe with a complex structure (Paleozoic covered by the Mendibelza conglomerates, the Canyon limestones and the flysch) which is disconnected in front of the main thrust sheet.

Structural interpretation: the Igountze Thrust reactivated an ancient north-dipping normal fault. In the Early Cretaceous this fault separated the Igountze block from the High Range basement. The diversity of the marine facies in the Mendibelza conglomerates (enormous fanglomerates with Paleozoic fragments from the south, and then puddingstones that reworked the basement and the Permian-Triassic sandstones and siltstones) as well as their sedimentary organization (onlap deposition, gradual thickening towards the south), suggest that these are canyon-cone deposits on a tilted block with increasing drowning to the south (Boirie, 1981). This organization is similar to, but younger than the one in the Arbailles (see former Stops). The Igountze Block therefore corresponds to a slab of the Pyrenean Iberian margin that collapsed during the Albian. The conglomerates that reworked the basement belong to the southernmost block (which was still high) of the High Range. The block collapsed in the Senonian, demonstrating the progressive widening of the Pyrenean sedimentary basin towards the South, during the Late Cretaceous (Souquet *et al*., 1985).

For further information: Ribis,

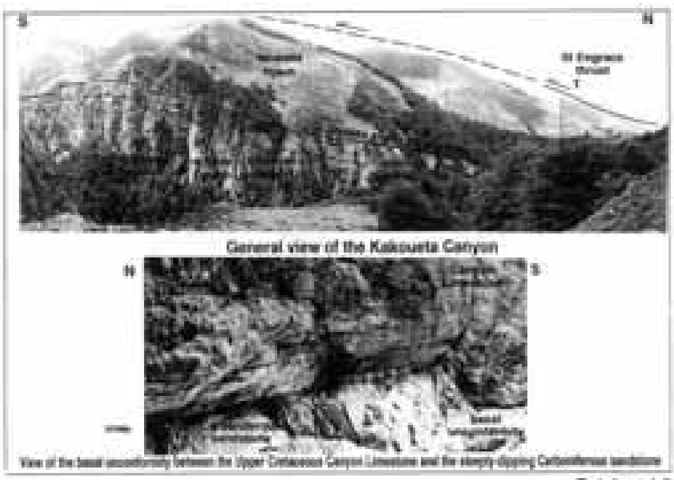


Figure 4.2a3 - Larrau-Lakhoura Thrust sheet (Total).

1965; Ducasse and Velasque 1988, Boirie 1981.

Stop 4.2b:

Intra-Senonian duplex underneath the Igountze and Lakhoura Thrusts.

Access: 500m east of the previous Stop. Park on the south side of the road.

Stratigraphy and structural geology: the Navarella turbiditic flysch (Campanian to Maastrichtian) is mainly composed of alternating marly limestones

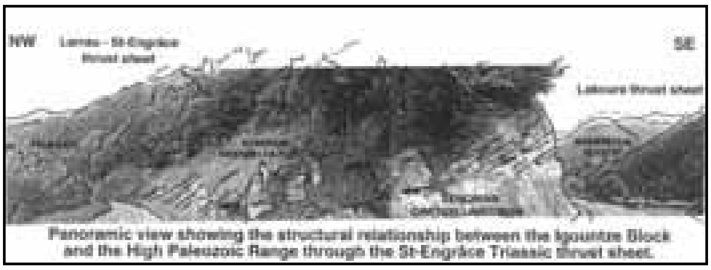


Figure 4.2b - Larrau-Saint Engrâce-Lakhoura Thrust Sheet.

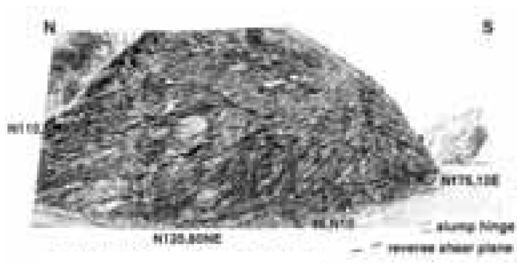


Figure 4.2b2 - Intra-formational antiformal duplex Navarella Flysch; St. Engrâce road

and marly sandstones. The beds are intensely sheared (fig.4.2b). Intraformational duplexes result from the development of reversed parallel south-verging ramps accompanying the motion of the Ste Engrace-Igountze Thrust sheet above, as well as that of the Larra Thrust further to the south.. At this Stop we can observe an antiformal duplex, with a subhorizontal floor shear plane and a roof shear plane folded along a N 120 axis. The sub-NS hinges of gravity slumps are preserved in the internal layers of the duplex.

This intense shear deformation occurred for at least 7 km southward, along a bedding plane slip zone at the top of the Canyon Limestones and underneath the Igountze and Lakhoura Thrusts, to the south of the modern Kakoueta canyon. The intraformational duplexing contributed to the southward allochthonous transport of the Igountze-Ste-Engrace Thrust and the Lakoura Thrust over a minimum of 7 km. The whole sheared Navarella flysch acts as a detachment thrust with a thick sole.

II. A transect of the Southern Pyrenees and Jaca basin

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Aim of the field trip

The aim of this itinerary is to examine thrust tectonics and interaction with sedimentary systems in the Southern Pyrenees. The Southern Pyrenees consist of south-vergent folds and thrusts and contain a very well-preserved record of synorogenic sedimentation. Synorogenic sediments are of uppermost Cretaceous to Lower Miocene age, and constitute the Jaca and Ebro basins (Plate 2, fig.1).

The field itinerary will proceed from north to south from the Axial Zone -a large thrust culmination that forms the orographic axis of the Pyrenees- to the Jaca basin -a Paleogene foredeep now incorporated in the the orogen showing a typical turbidite to molasse succession-, and finally to the southern mountain front of the External Sierras –where the Pyrenees overthrust the present foreland of the Ebro basin. Synthetic geological descriptions of the transect visited can be found in Teixell (1996, 1998).

During the trip we will have the opportunity to study features as 1) detailed kinematics and mechanics of thrusting, 2) interactions between growing thrust faults and folds and proximal alluvial fans, and 3) episodic evolution and progressive deformation of foreland basin deposits.

Afternoon:
Axial zone and Northern Jaca basin

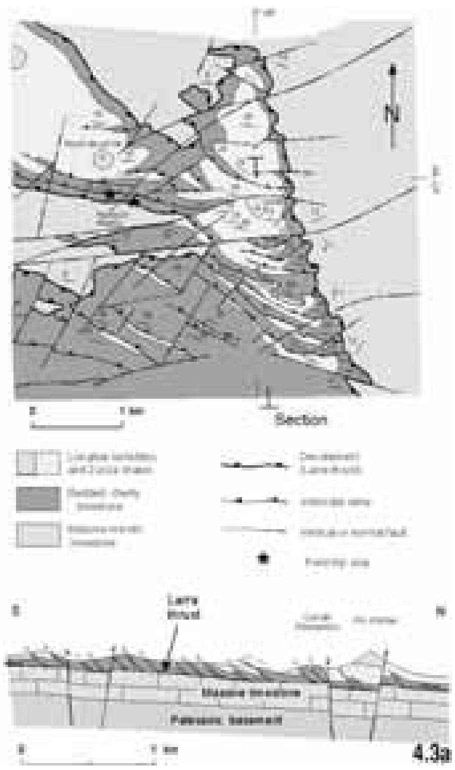
Stop 4.3:

Pierre-Saint-Martin. The Larra thrust system.

Access: follow the main road NA 1370 to the SW to cross the Spain/France border and park some 100 m after the signal, beside a road bend.

Structural landscape: the Upper Cretaceous rocks

that form the Axial Zone cover consist of basal platform limestones ("Calcaires des Cañons", Cenomanian to Santonian) that rest directly on the Paleozoic. Overlying the limestones are open marine shales, and a sandstone/shale turbiditic unit, visible in the landscape (Ribis, 1965). The latter formations are of Campanian-Maastrichtian age, and represent a drowning of the Santonian platform that can be attributed to flexure at the initial stages of the Pyrenean thrust loading.
One special interest of the Cretaceous rocks of the visited region lies in that post-Hercynian rocks have been eroded away in more eastern parts of the Pyrenean Axial Zone (Plate 2, fig. 1), and thus the area provides an unique opportunity to constrain the Alpine deformation in this basement massif.
In this Stop we can observe details of the Larra thrust system, characteristic of the Upper Cretaceous rocks of this northern part of the Axial Zone. The Larra system is a thin duplex composed of numerous south-vergent ramps that root in a bedding-parallel floor thrust (Larra thrust) (Fig. 4.3a). In the study area, the Larra thrust is located within the Calcaires des Cañons, at the boundary between a lower massive limestone member and a upper, well-bedded and chert-bearing limestone member (Teixell, 1990; Teixell *et al*., 2000). The ramps above cut the chert limestones and the Zuriza shales.
From the parking area we can walk westwards for some metres to get a panoramic view of thrust imbrications on a hillside in the landscape.
Back to the road, the cuts located just north of the parking place show a good outcrop of a thrust ramp, with a set of associated minor structures that provide inferences for the kinematics of the thrust system. The thrust ramp, brings the middle part of the chert-bearing limestone member and the shaly upper part of this stratigraphic unit in contact (Fig. 4.3a). Bedding is truncated in a ramp-over-ramp geometry. The fault zone is characterized by two main subparallel fracture surfaces, which enclose an intervening rock slice that is strongly deformed and displays an internal oblique foliation coherent with ramp shear. Within the wall rocks, there are contractional, conjugate en-echelon arrays of calcite veins. As the shortening axes of the arrays have been tilted together with bedding, they indicate an early episode of layer-parallel shortening prior to folding. An incipient foliation, oblique to bedding, is also present. Transport directions (contraction or shear) that can be deduced from the orientation of minor structures range from N-S to NE-SW (fig. 4.3b).



Stop 4.4:

Pierre-Saint-Martin. Fault zone structure of the Larra floor thrust.

Access: walk down the road back to the roadside outcrops by the signal post of the Spain/France border.

Structural landscape: the Larra thrust fits the definition of a décollement: a bedding-parallel slip plane without repetition of stratigraphy. The Larra thrust dips parallel to footwall bedding, at gentle angles to the N, due to post-thrust tilting and normal faulting during the development of the Axial Zone antiform (Gavarnie thrust stage).
The Larra thrust appears in outcrop as a meter-thick band of strongly foliated rock that, at first sight, resembles a mylonitic marble. However, closer inspection shows that its foliated aspect is due to a densely packed stack of bedding-parallel calcite veins. A detailed profile of the fault zone at this locality is presented in figure 4.4. Bedding-parallel veins are

Figure 4.3a - Detailed geological map and cross-section of a segment of the Larra thrust system in the northern Axial Zone (Pierre-Saint-Martin area) (after Teixell, 1990). The Refugio de Belagua is located some 5 km to the SW.

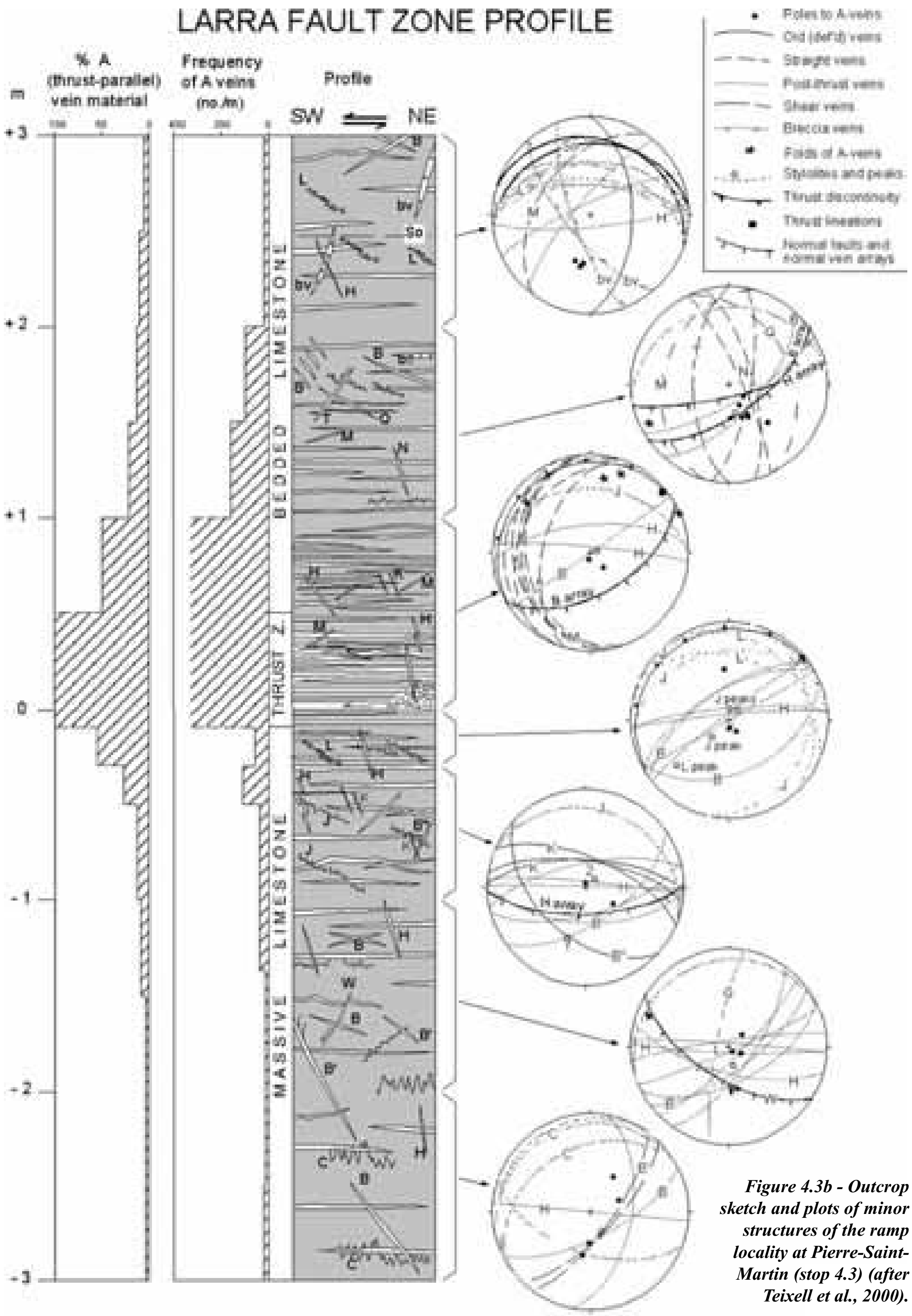


Figure 4.3b - Outcrop sketch and plots of minor structures of the ramp locality at Pierre-Saint-Martin (stop 4.3) (after Teixell et al., 2000).

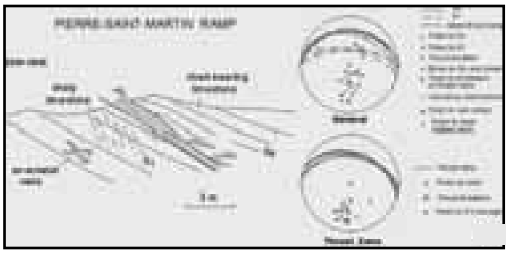


Figure 4.4 - Structural profile of the fault zone of the Larra floor thrust at Pierre-Saint-Martin (stop 4.3). Positions and observed cross-cutting relations of chief structures are shown. Thrust-parallel veins are labelled 'A'. Other letters refer to different sets of veins, stylolites and other minor structures, as indicated in the legend. Orientations are on the right; percentage and frequency of A veins on the left (after Teixell et al., 2000).

abundant for at least 3 m in both hangingwall and footwall, and their frequency increases towards the fault zone center, where the rock appears formed by almost 100% vein material. From a microstructural point of view, veins are filled with blocky calcite crystals with variable internal deformation by twinning and recrystallization. No fibers have been observed. Cross-veins oblique to bedding are also present, although less abundant. In fact, there is a complex assemblage of diverse mesostructures that, apart from the veins include vein breccias, stylolites, folds and minor faults. On the basis of aspect, orientation and relationships to one another, the mesostructures have been correlated across the profile and grouped in sets (Fig. 4.4).

The Larra thrust has the singularity of being located within a mechanically strong limestone unit. Hence rheological weakness and water-generating capacity of the host rock were not primary causes for detachment. Teixell *et al*. (2000) proposed that the weakening was induced by stress and fluid channeling. Fluid channeling at a slight lithologic change within the limestone caused episodic brittle weakening and localized yield. Yield initially took the form of bedding-parallel extensional fractures.

Being constituted by interfaces between solid and fluid, the open extensional fractures ("water sills") were surfaces of zero shear strength that could accommodate appreciable increments of slip during short time spans. Even though twinning and recrystallization are conspicuous at the thin section scale, these were not key mechanisms in movement of the Larra thrust, which can be regionally estimated to be in the order of 5 km. Displacement was mostly accommodated by slip on the water sills, leaving little signature.

A remaining problem lies in that bedding-parallel veins are compatible with subhorizontal maximum stress in a thrust regime, but as major floor thrust displacement was parallel to bedding, there must have been some sort of stress reorientation to impart a shear stress on the fracture. This may have occurred by differential elastic contraction (Teixell *et al*., 2000). Hence, we infer that movement on the Larra thrust was achieved by a cyclic repetition of episodes of 1) bedding-parallel stress and crack dilation, and 2) stress reorientation and bedding-parallel shear (leading to décollement slip).

Stop 4.5:

Refugio de Belagua. General view and stratigraphy of the Cretaceous cover of the Axial Zone, figure 4.5.

Access: driving south on NA 1370 up to the mountain hut of the Refugio de Belagua (signal).

Structural landscape: from the Refugio there is a good overview of the Axial Zone and the overlying Lakora klippe. A cross-section of the area is shown in figure 4.5. Upper Cretaceous limestones that are close to the Refugio contain shallow-water forams.

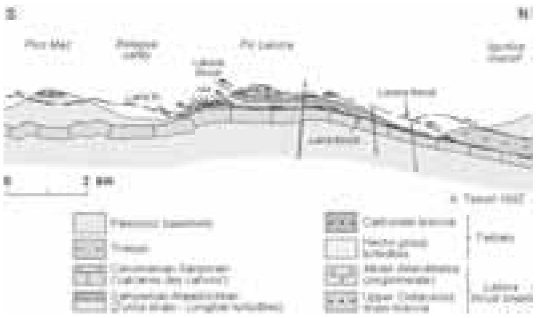


Figure 4.5 - Geological cross-section of the Belagua-Lakora area, located in the northern Axial Zone. Indicated as «hut» is the position of the Refugio de Belagua (stop 4.5).

Overlying the limestones are strongly cleaved shales, outcropping along the road to the Refugio.

The large-scale structure that can be deduced in the landscape is a broad antiform. Minor, south-directed thrust faults deforming the Cretaceous formations in the vicinity of the Refugio belong to the Larra thrust system, formed earlier (Teixell, 1990). Thrusts cause numerous limestone-shale repetitions, and show fault-related anticlines that bear an associated north-dipping slaty cleavage, especially in shaly rocks.

Stop 4.6:

Roncal valley. The Eocene flysch of the Southern Pyrenees (Hecho group).

Access: drive down the Roncal valley and stop along the main road some 2 km south of the village of Roncal.

The roadcuts at this stop provide a good outcrop of the Eocene turbidites of the Hecho group. Turbidites consist of silici-clastic sandstone/shale decimetric-scale cycles, in which the pelitic component is largely dominant. They represent relatively distal, basin plain facies within the Hecho group, and they have been dated as Lutetian in age (Mutti, 1984; Labaume *et al*., 1985).

Structural landscape: the turbiditic beds describe a synclinal fold at the outcrop scale, to which an axial planar cleavage is associated. This fold illustrates the characteristic deformation style of the Hecho group, which originally accumulated in a foredeep in front of the growing orogen, and was later deformed and incorporated in the mountain chain (e.g. Plate 2, fig. 1).

Above this outcrop, in the landscape, we can see a carbonate megabreccia bed, also characteristic of the Hecho group. Clasts within the megabreccia derive from carbonate platforms that flanked laterally the turbiditic trough. The position of these platforms and the geodynamic significance of the megabreccias have been the subject of diverse interpretations. Some authors have proposed a northern provenance of the resedimented material, from hypothetical platforms lying on top of the active thrust margin of the basin (Séguret *et al*., 1984; Labaume *et al*., 1985), wheras other authors have proposed a southern source area, from platforms existing in the distal (foreland) margin of the basin (Barnolas and Teixell, 1994).

The megabreccia bed that we observe is numbered 5 in the sequence of 8 main units like this in the northern Jaca basin. It shows a large mushroom-like feature interpreted to be a giant water-escape structure by Labaume *et al*. (1983).

DAY 5

from Jaca to Labuerda.

Morning

Southern Jaca basin and External Sierras

Stop 5.1:

San Juan de la Peña. Oligocene conglomerates of the Jaca basin.

Access: drive from Jaca to the west along C134 and then take the road to the left that drives to the San Juan de la Peña monastery. On the way to the External Sierras, we stop at San Juan de la Peña for an overview of the Campodarbe formation that forms the youngest infill of the Jaca basin (Puigdefàbregas, 1975).

Structural landscape: the Campodarbe beds describe the Guarga synclinorium (the present-day axis of the Jaca basin), whose position in the general cross-section can be observed in Plate 2, fig. 1. Fluvial sandstone channels and shales grade upwards to layered and massive alluvial fan conglomerates, which constitute the San Juan de la Peña massif. The age of these rocks is Upper Eocene to Lower Oligocene. The conglomerates show growth strata (progressive unconformities) within synclinal folds. The San Juan de la Peña massif is best known by its monastery of the X-XII centuries, picturesquely sheltered by the conglomeratic cliffs.

Stop 5.2:

Embalse de la Peña. Stratigraphy of the External Sierras

Access: follow N240 to the south to stop in the proximity of the reservoir Embalse de la Peña. Proceed along the roadcuts to the south of the reservoir and the bridge.

Structural landscape: the External Sierras are the foothills that constitute the South Pyrenean thrust front (Plate 2, fig. 1). Emergent thrusts bring Mesozoic rocks to the surface, starting with Triassic shales and evaporites (Keuper facies) that, when present, form the regional décollement level of the Pyrenean cover thrusts (Séguret, 1972).

The stratigraphic succession of the External Sierras is sketched in Fig. 5.2. We can recognize the diverse units in the gorge of the río Gállego, which provides one of the best and most visited sections across the Sierras.

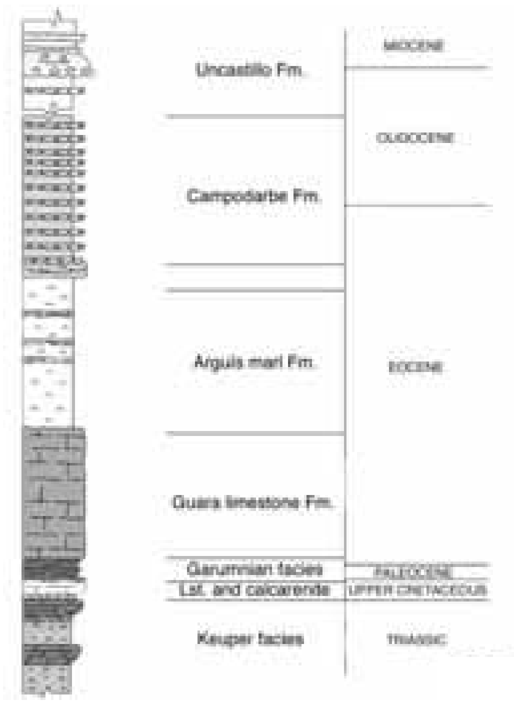


Figure 5.2 - Stratigraphic log of the Mesozoic and Tertiary of the External Sierras (taken from Millán et al., 1995).

Above the Keuper, we find rudist-bearing of Upper Cretaceous age. Although there may be hiatuses of poorly constrained extent, the overlying red beds and lacustrine limestones (Garumnian facies) are of uppermost Cretaceous to lowest Tertiary age. The few hundred m of platform limestone of the Guara formation (middle Eocene) are the basin-margin counterparts of the Hecho turbidites. The Arguís and Campodarbe beds are similar to those within the Jaca basin, but much reduced in thickness.

Stop 5.3:

Riglos. Thrust front and proximal alluvial fan interactions.

Access: continue down the N240 and take the road to Riglos on the left (HU 310). Cross the village and park just below the conglomerate pillars of the Mallos. From the parking place walk up along the trail between the two prominent Mallos up to the Cerro Leonar.

The Mallos is the local name of a group of picturesque stacks that form a well-known touristic attraction of the area. They are built by Upper Oligocene-Lower Miocene conglomerates attached to the southern face of External Sierras.

The conglomerates are composed of clasts derived from the Mesozoic and Tertiary of the Sierras, and formed within small alluvial cones that towards the south pass rapidly to fluvial sandstones and shales (Puigdefàbregas, 1975; Hirst and Nichols, 1986). They accumulated synchronically with the main emergence of the External Sierras, and they represent the last synorogenic foreland basin deposits of this segment of the Pyrenees (see front cover, fig.2 and Fig. 5.3, A and B).

The Mallos show beautiful signs of syntectonic sedimentation. As we will clearly see by the itinerary up the Cerro Leonar, the conglomerates onlap a pronounced paleorelief, sealing some of the older thrust faults, but are in turn overridden by higher-level thrust sheets. This kind of relationship is common in the area, providing evidence for thrust sequences (Millán *et al.*, 1995).

From the top of the itinerary we have a good panoramic view of the entire External Sierras overlooking the Gállego gorge. The structure of the Sierras is certainly very complex, with refolded thrusts and large overturned successions (see front cover, fig. 2). It puzzled geologists for decades, and was not resolved until the work by Puigdefàbregas and Soler (1973). They showed the existence of a large south-vergent thrust fault rising from the axial plane of a tight anticline, a thrust that is itself folded in an antiformal arch (see front cover, fig. 2, Plate 2, fig.1, and fig. 5.3, A and B), so its leading edge, isolated by later erosion, appears unrooted and completely overturned.

The antiform that arches the thrust has a strong easterly plunge, so it develops a fold closure located beneath our feet, just above the Gállego river. The leading edge of the thrust is sealed by the conglomerates of the Mallos, that in turn show progressive unconformities (growth strata) associated to the later antiform (Teixell and García Sansegundo, 1995). Minor out-of-sequence thrusts, as those observed along the way up, also show syntectonic relationships with the conglomerates.

Afternoon

III. The Ainsa Basin

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Introduction

The Ainsa basin is located in the footwall of the Cotiella-Montsec thrust sheet (Seguret, 1972) (Figs. 5.4b, 5.5a). This cover thrust sheet consists of Mesozoic and syntectonic Paleogene rocks that have been detached from the Variscan basement on Lower Triassic evaporites. During the Early Eocene southward displacement of the Cotiella-Montsec, the Tremp piggy-back basin developed on its hangingwall (Fig. 5.5a). The basin contains platformal and continental terrigenous facies that grade westward across an oblique lateral ramp of the Cotiella-Montsec thrust sheet (in the region of the La Foradada fault, Fig. 5.5a) into the marine Ainsa turbidite basin (Nijman and Nio, 1975, Mutti *et al*., 1988).

The Ainsa basin is deformed by the Peña Montañesa thrust system in the northeast, by the north-trending Mediano anticline in the south, and by the Boltaña anticline to the west (Fig. 5.4b). The structure of the eastern part of the Ainsa basin is dominated by a northwest-trending imbricate thrust system and related folds. Two main thrusts in the footwall of the Cotiella thrust sheet, the Atiart and Los Molinos thrusts, detach the Ainsa turbiditic basin from underlying Paleocene and Lower Eocene carbonate platform rocks. The Ainsa basin evolved to a piggy-back condition as thrusting propagating forelandwards and the detachment Mediano anticline and the fault-propagation Boltaña fold developed. These folds developed during the filling of the Ainsa basin synchronously with their dextral rotation in the footwall of the Cotiella thrust sheet.

The basin originated as a foredeep ahead of the innermost of these thrusts in the Early Eocene (Fig. 5.5a) and evolved into a piggy-back setting as the thrust front propagated towards the foreland. The

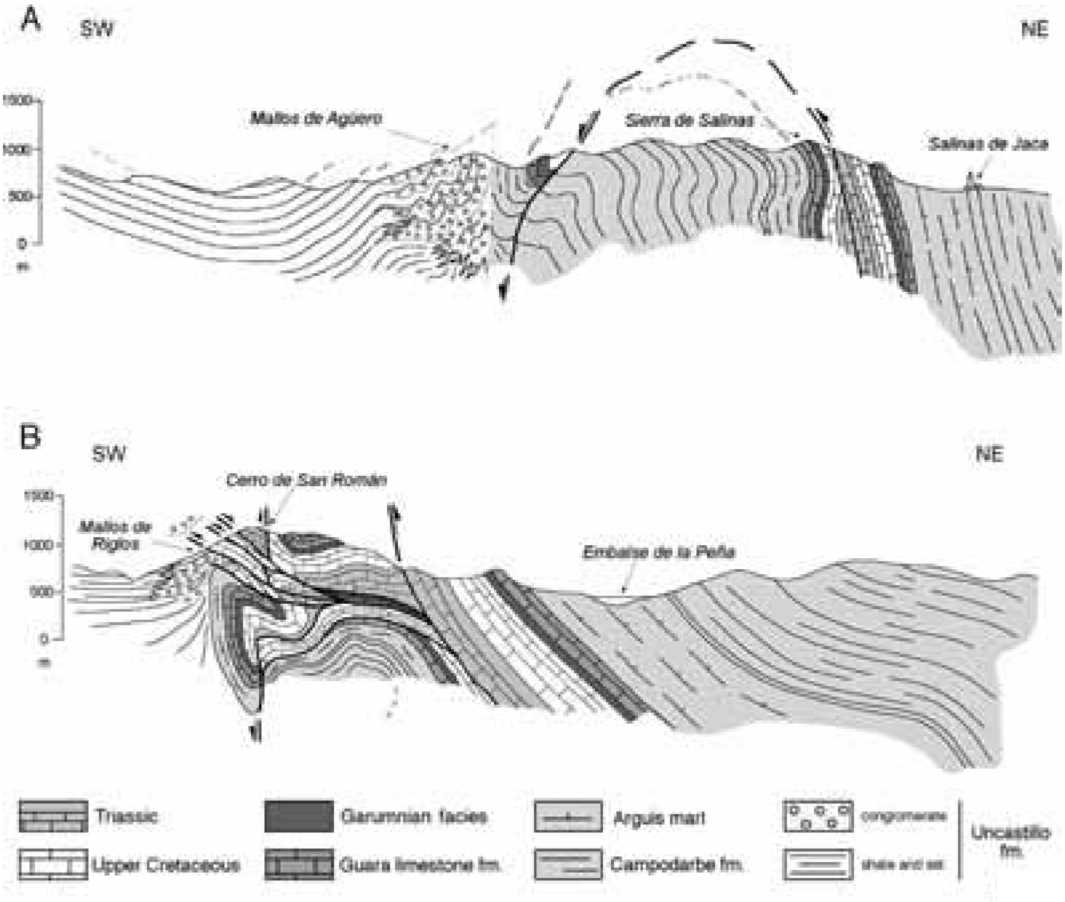


Figure 5.3 - Cross-sections through the External Sierras (after J. García Sansegundo, in Teixell & García Sansegundo, 1995).

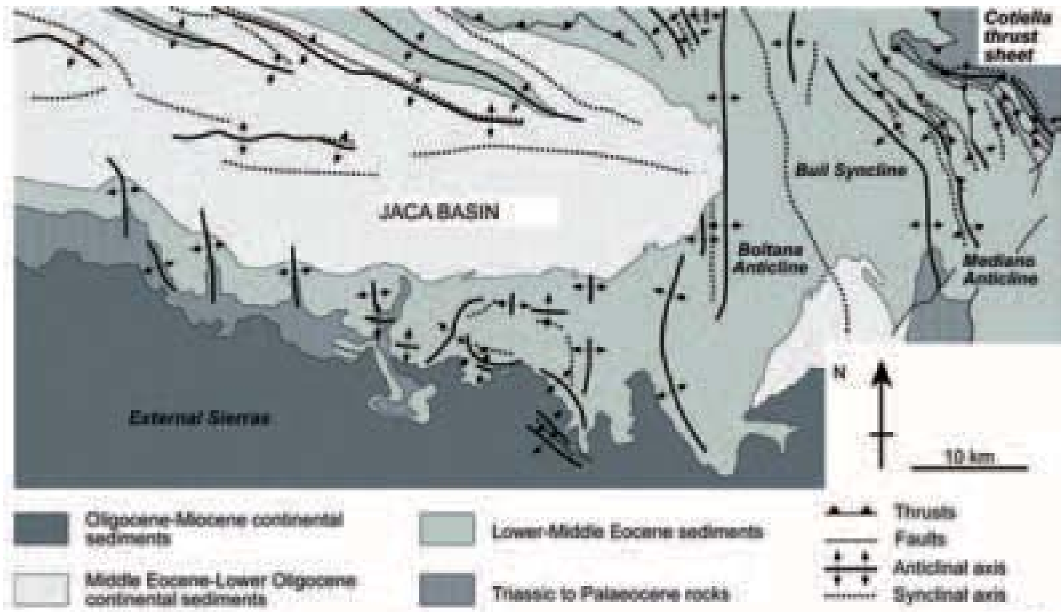


Figure 5.4a - Geological sketch map of the External Sierras, Jaca Basin and the Ainsa Basin.

foredeep filled mostly with a slope complex which is up to 4,000 m thick, 40 km long and 30 km wide. The slope complex had a fluviodeltaic source in the southwest and evolved to the NW into a basin-floor complex (Figs. 5.5a and 5.5b). The basin-floor complex in the Jaca basin consists of the classically termed "outer-fan" sandstone lobes, which are replaced downcurrent by basinal sandstone and mudstone interbeddings. The individual sandstone-dominated lobe elements are several tens of metres thick, more than 15 km wide and up to 80 km long.

The Ainsa Slope Complex consists primarily of mudstones where coarser-grained lithosomes are embedded. The latter correspond to turbidite systems with outcrops allowing the reconstruction of their 3D anatomy. The turbidite systems are generally thinner than 300 m, few kilometres wide, with a preserved length reaching up to 20 km, and mostly formed by channels and associated overbank elements. The outcrops allow insights on channel to overbank transitions and one system preserved significant changes downslope. Most of the facies involved indicate deposition from a broad variety of turbidity currents; however, in particular positions of turbidite systems, packages of turbidites are faulted and

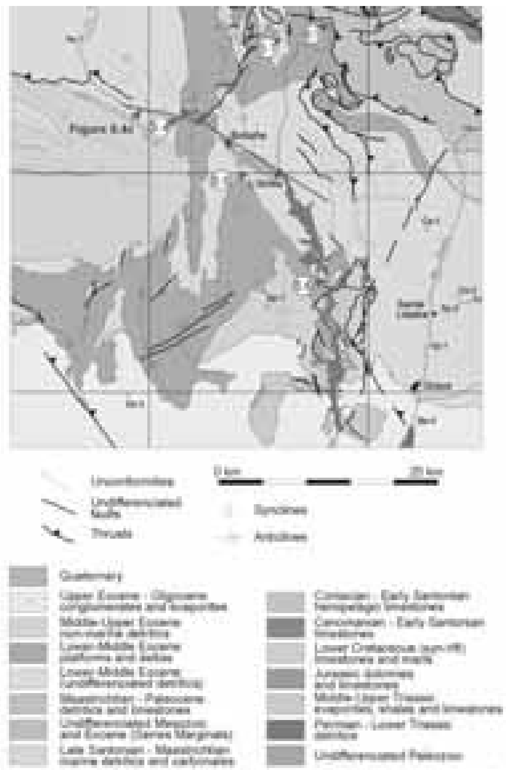


Figure 5.4b - Geological sketch map of the Eastern part of the Ainsa Basin and of the Boltaña Anticline.

deformed by slumping at diverse scales or transformed into debris flow deposits. The resedimentation process interacted with turbidite deposition and introduced a degree of architectural complexity.

The interplay between tectonics and sedimentation in the Ainsa basin had different scales. Below is a summary of these relationships, ordered from basin scale to the scale of individual turbidite channel elements.

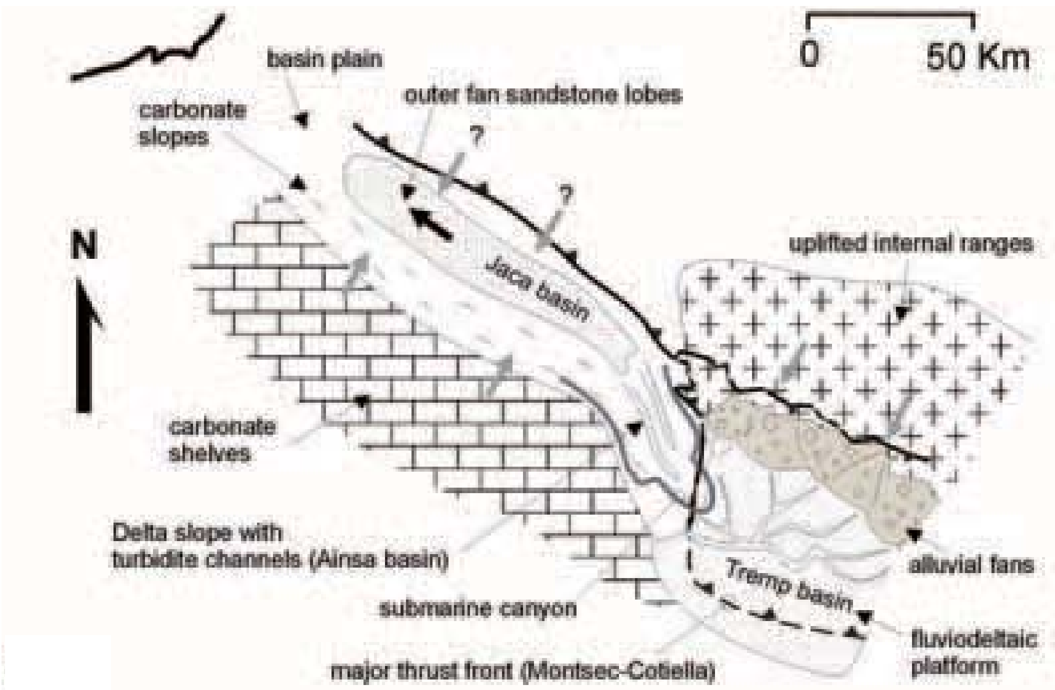
The propagation of the thrust front caused the Ainsa basin fill to be segmented as four major depositional cycles which are delineated by widespread angular unconformities and are stepped towards the foreland (Fig. 5.5b). These unconformities are submarine truncations, up to several hundreds of meters deep, which laid parallel to the elongation of the basin and were carved by mass wasting. The potential for mass wasting was created by forelimb rotation along the active basin margin and by flexural downwarping in the outer basin margin. These "cañons" acted as containers for further slope deposition.

Within a single depositional cycle, an overall decreasing rate of tectonic deformation caused an overall decrease in slope gradient, in turn responsible for contrasts on facies, architectural style and external geometry among stratigraphically consecutive turbidite systems (Fig. 5.5b).

Pulses of increased deformation caused angular unconfomities at the base of turbidite systems. These evolved overall towards abandonment, trough cycles of channel-complex development and abandonment. However, their internal architecture and external geometry varies between end-members: 1) vertically-stacked and symmetrical in cross-section and 2) complexly juxtaposed and markedly asymmetrical. The more complex patterns arose in zones left piggy-back between growing anticlines. The overall straight downslope platform of individual turbidite channels was in some cases modified by anticline-related topographies.

Objectives: to examine the tectono-sedimentation relationships of the Ainsa basin, growthg folds and the surrounding thrust systems.

Figure 5.5a - Palaeogeography of the South Pyrenean foreland basin at Early Lutetian. Turbiditic systems of the Ainsa basin had feeder fluvial systems in the Tremp basin. Clast composition is dominated by limestone clasts derived from Paleocene and Mesozoic rocks, but it also includes granites, volcanic rocks and others, which correlates to the exposed rocks in the hinterland, as it is also indicated by alluvial fans fringing the Northeastern Tremp Basin. From Arbués et al. 1998.



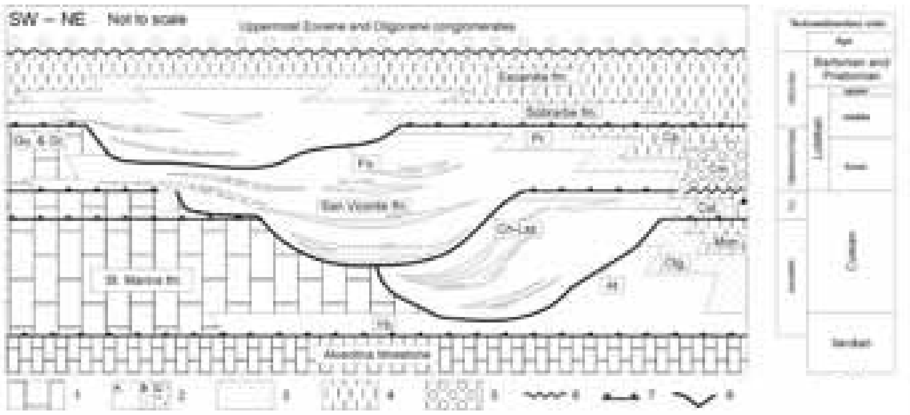


Figure 5.5b - Synthetic stratigraphy of the Ainsa Basin.
<u>Main depositional facies</u> :1 : carbonate shelf ; 2 : delta and carbonate slope ; 2A : mudstones, 2B : silici-clastic turbidites ; 2C : resedimented carbonates ; 3 : delta ;4 : alluvial plain ; 5 : alluvial plain. <u>Discontinuities</u> : 6 : subaerial unconformity ; 7 : condensed section ; 8 : surface of submarine gravitational erosion. <u>Code for local names of stratigraphic units</u> : Yb : Yerba marls (^) ; Gu & Gr :Guara limestones (+) and Grustan mb (*) ; Cgl : Castilgaleu fm ; Mon : Montilobat fm. ; Cst : Castissent fm. ; Cm : Campanué conglomerates (*) ; Cp : Capella fm. ; Pr : Perarrua fm. . <u>Code for surfaces of subamrine gravitational erosion</u> : At : Atiart ; Ch-Lsz : Charo-Lascorz ; Fo : Formigales. <u>Other codes</u> : Fu : Fuendecampo tectosedimentary unit. From Arbués et al. 1988.

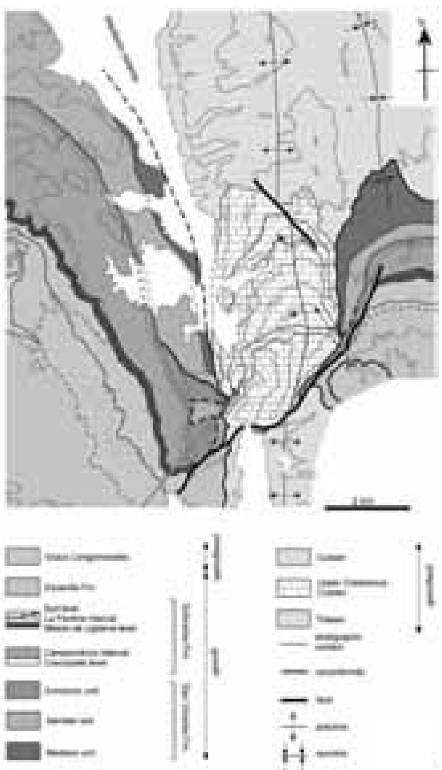


Figure 5.6a - Geological map of the Mediano anticline, Ainsa basin (after Poblet et al. 1988).

Stop 5.4:
Boltaña anticline
Access:
From Broto, drive east to Ainsa. Stop in the intersection leading to the village of Janovas.
The Boltaña anticline is considered the western boundary of the Ainsa basin, however it grew during the latest stages of the turbidite filling of the Ainsa basin and mostly during the Middle to Late Eocene and 5.4c). Most of the turbidite systems of the Ainsa basin continued basinward into the Jaca basin without a topographic barrier at the present location of the Boltaña anticline (Fig 5.5a).
This Stop shows the steep, west-dipping limb of Eocene limestones of the Boltaña anticline.

Stop 5.5:
Tozal de Guaso. Panoramic view of the Ainsa basin.
Access :
From Boltaña drive east to Ainsa. Take the road to Guaso and the Guara park after crossing the Ara river. Follow this road to an intersection, turn right and follow the road to the village of Guaso. Take the road to the church at the top of the hill.
Structural landscape
From this viewpoint the Ainsa basin and the surrounding structures are visible: the Boltaña

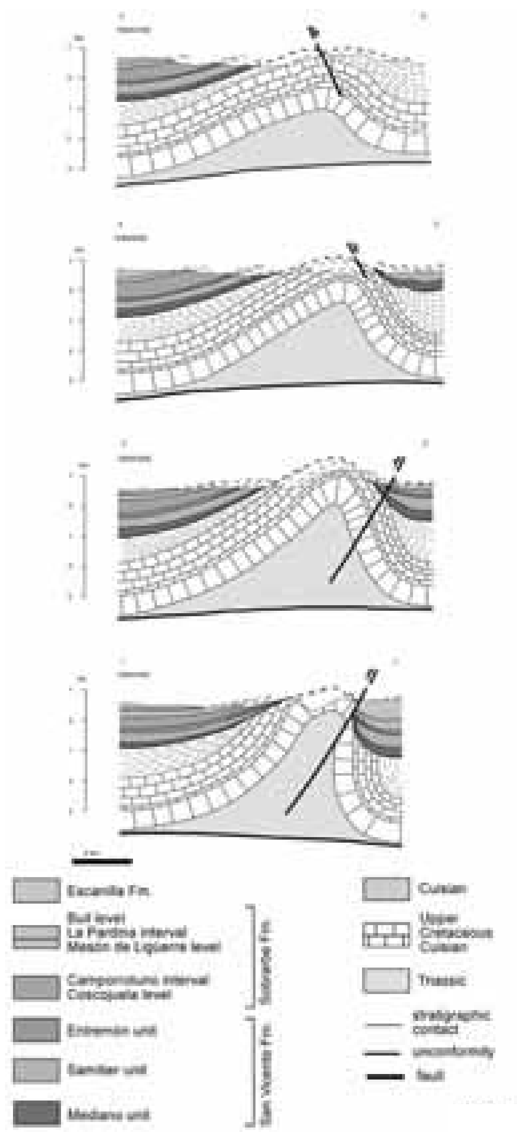


Figure 5.6b - Mediano anticline evolution.

anticline to the west, the Mediano anticline to the east and the Monte Perdido-Gavarnie, Peña Montañesa and Cotiella thrust sheets to the north-northeast.
This viewpoint is an excellent spot to discuss the main geological features of the area, summarize the structures observed these two days.

Stop 5.6:

Ermita de San Emiterio and San Celedonio.

Access: drive to Ainsa and take the C-138 towards Barbastro. Continue southwards past the now submerged village of Mediano. Turn left onto a small side road that leads to the village of Samitier and park the vehicles near the small square and fountain in the centre of the village.
From the parking spot in the village take the farm track that leads off to the left and up the hill to the Ermita de San Emiterio and San Celedonio. Several good view Stops can be found along the way overlooking Mediano lake and the western limb of the Mediano anticline.
Climb to the old church at the Ermita de San Emiterio and San Celedonio. From here there are excellent views of the Mediano anticline, the onlapping Ainsa turbidites, and the Buil syncline to the left. In the distance (if the weather is clear) the main ranges of the Pyrenean axial zone can be seen.

Structural landscape

The Mediano anticline is a north-plunging, north-trending detachment anticline developed above the Triassic evaporites (Fig. 5.6a). It is cored by massive to thick-bedded Eocene limestones (Fig. 5.6b). Eocene (Lutetian) turbidites onlap the anticline in a syn-folding growth sequence (Figs. 5.6a, 5.6b and 5.6c). The geometry, facies distribution (from alluvial to deep-water silici-clastics and carbonate rocks) plus the available magnetostratigraphy and biostratigraphy pose restrictions to the kinematic evolution of the anticline (Fig. 5.6c).

Stop 5.7:

Cotiella and Peña Montañesa thrust sheets.

Access: drive back to Ainsa and Labuerda and continue north along the main road (N-138) until the village of Escalona. At this locality turn left to the Añisclo valley and immediately take the small road to the right to the village of Puertolas. Continue up the mountain and Stop just past the small village of Santa Maria before the village of Puertolas.

Structural landscape

From this viewpoint the large klippes at Peña Montañesa and the Cotiella thrust sheets can be viewed. The flat-lying floor thrusts of these thrust sheets bring Mesozoic and Paleocene limestones on top of the Lower Eocene turbidites and marls of the Ainsa basin. The Atiart thrust (floor of the Peña Montañesa thrust sheet) merges hindwards with the Cotiella thrust.

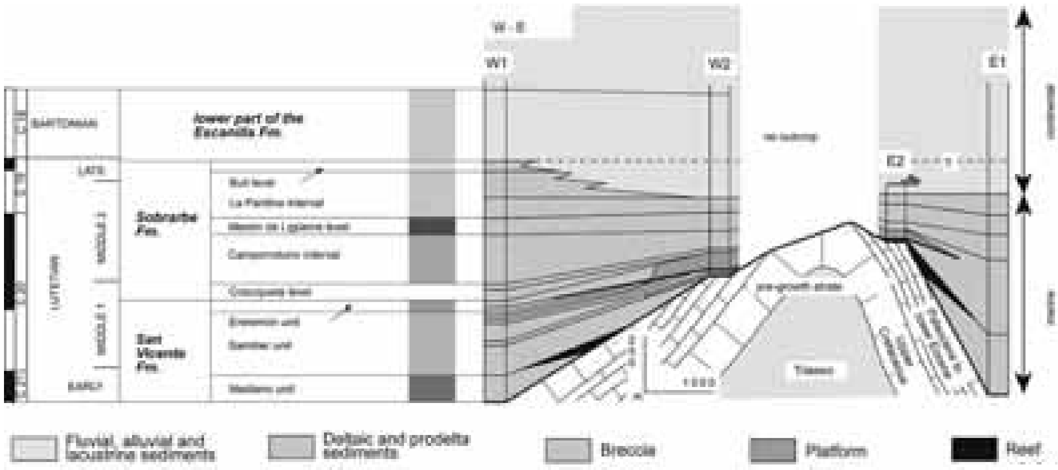


Figure 5.6c - Synsedimentary growth and kinematics of the Mediano anticline.

DAY 6

from Labuerda to Pau.

IV. The Alpine structure of the Basement of the Axial Zone and the Monte-Perdido-Gavarnie and Cotiella Thrust Sheets

By Josep Anton Muñoz

Introduction

The Cotiella thrust sheet (Fig.6.1a) is the northeastern part of the Montsec thrust sheet, the most extended unit of the South Pyrenean Central Unit (Seguret, 1972) or the so-called Mesozoic Upper Thrust Sheets (Muñoz *et al*., 1986). It consists of several structural units characterized by a very thick Mesozoic succession (mostly Upper Cretaceous), detached above the Triassic evaporites (Fig.6.1b). The floor thrust of the Cotiella thrust sheet is folded by the emplacement of the lower structural units (Monte Perdido-Gavarnie and basement thrust sheets). It dips subhorizontaly for most of its cartographic extension and dips southwards in the north above the lower thrust sheets. The Cotiella floor thrust climbs up southwards in the footwall from the Paleocene limestones into the Lower Eocene turbidites of the Ainsa basin. The Cotiella thrust sheet was displaced southwards a minimum of 20 km during the Early Eocene.

Below the Cotiella, the Monte Perdido-Gavarnie thrust sheet shows a distinct stratigraphy. It involves the Hercynian basement and the Mesozoic succession is more reduced and much thinner than the Cotiella. The basement consists of Silurian, Devonian and Carboniferous slightly to non-metamorphosed shales and limestones affected by Hercynian thrusts and folds. The Upper Cretaceous limestones, marls and sandstones rest unconformably on top of the Hercynian basement. They form a succession several hundred meters thick (800m aprox.) in contrast with the several km thick Upper Cretaceous series of the overlying Cotiella thrust sheet. The Paleocene consists of limestones and dolomites and is overlain by the Lower Eocene marls and turbidites.

The Monte Perdido-Gavarnie thrust sheet is structurally in continuation with the Ainsa basin (i.e underneath). The thrusts related with the Boltaña and Mediano anticlines are the southern continuation of the Gavarnie thrust which clims up section southwards into the hangingwall from the Silurian-Devonian rocks to the bottom of the Mesozoic succession. Above the basement, this thrust sheet shows a pile of small thrust imbricates repeating the Upper Cretaceous and Paleocene formations. These units show an increasing displacement westwards and have been described as the Monte Perdido thrust sheet (Seguret, 1972). The Monte Perdido thrusts merge upwards into the lower part of the Lower Eocene turbiditc succession of the Ainsa basin.

Below the Gavarnie thrust, basement thrust sheets with a different stratigraphy are exposed. They consist of metamorphic Lower and Upper Paleozoic rocks intruded by Upper-Hercynian granitoids.The cover of these lowermost thrust sheets only consists of Triassic red beds, unconformably overlying the

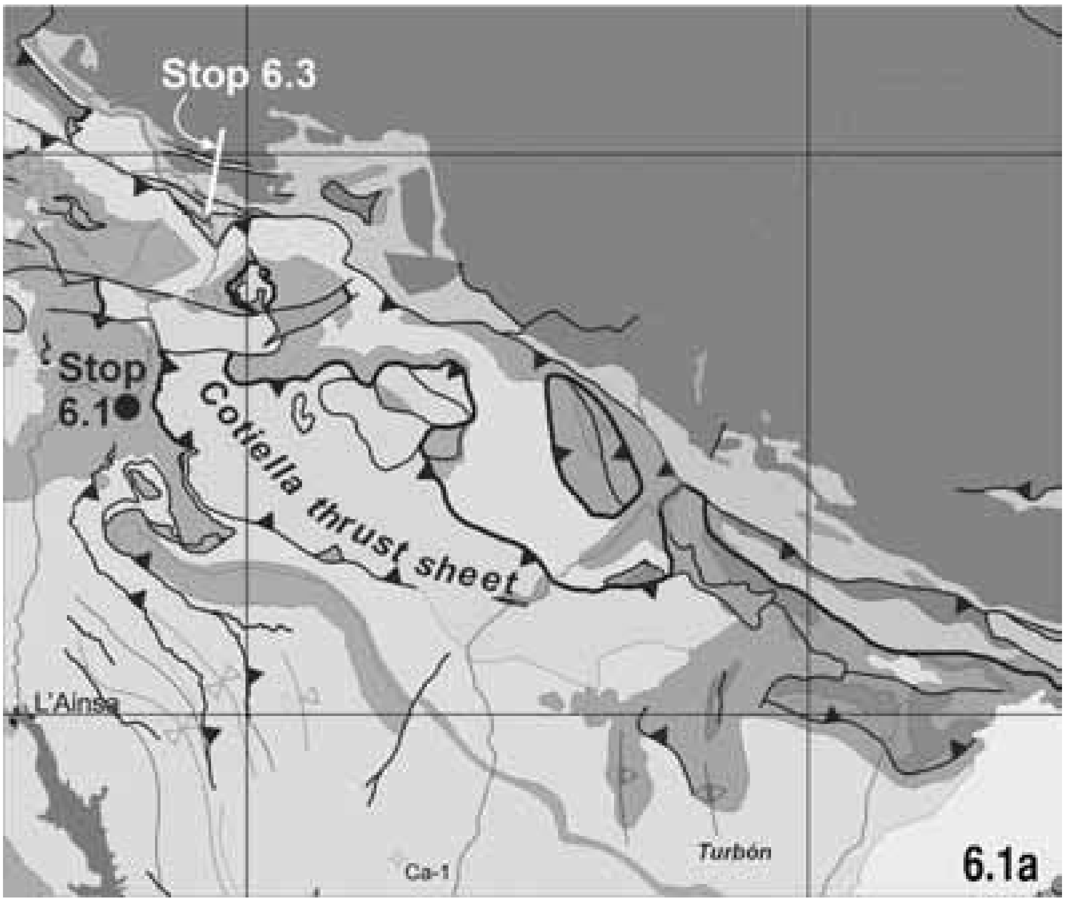


Figure 6.1a - Geological sketch map of the Cotiella thrust sheet and location of stops.

basement, Keuper above, and a thin bed (few tens of meters) of Upper Cretaceous limestones on top. The basement and the cover are both tightly folded and affected by thrusts.

Objectives: to examine the Alpine thrust structure of the basement units of the Pyrenean Axial Zone as well as the structure of the main cover thrust sheets above the basement.

Stop 6.1:

Cotiella thrust and structure of its footwall.

Access: from Labuerda drive north to Bielsa (N-138) until the village of La Fortunada. Pass the village for 1 km before the road tunnel of Las Devotas.

Structural landscape

The purpose of this Stop is to continue a general cross section of the area as well as to observe some structural features of the Cotiella thrust sheet and its footwall (Fig.6.1a, 6.1b, 6.1c, 6.1d). On the other side of the valley the Cotiella thrust is visible at the bottom of the Upper Cretaceous limestone cliffs (1,200 meters high). It lies subhorizontaly above the lowermost Eocene marls. Northwards of this locality the thrust as well as the Upper Cretaceous-Eocene beds in the footwall (Monte Perdido-Gavarnie thrust sheet) are tilted to the south 25-30º and truncated by a pair of E-W trending conjugate extensional faults (Fig. 6.1c, 6.1d).

The Eocene marls are folded and affected by a prominent N-S, NNW-SSE trending cleavage at a high angle with the general E-W, ESE-WNW general trend of the beds. This geometry is the result of a late tilting of the deformed marls and the Cotiella thrust sheet by thrusting of underlying basement thrust sheets.

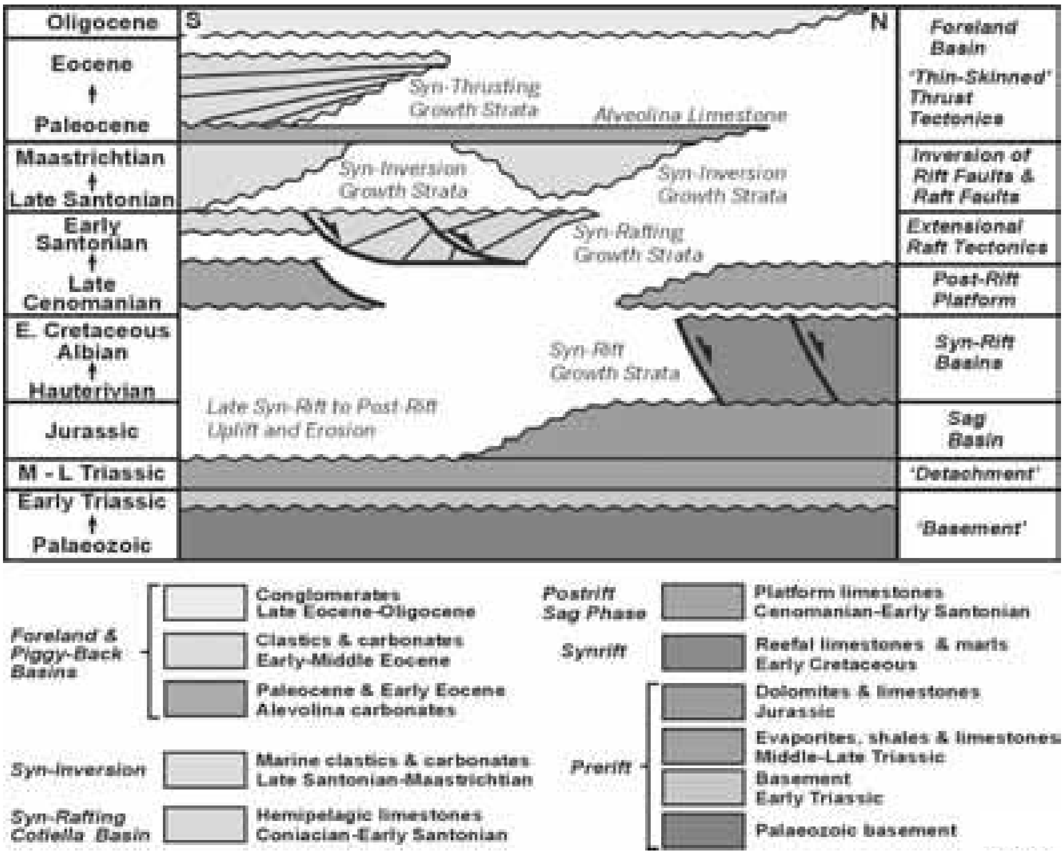


Figure 6.1b - The Syn-Rafting Cotiella basin replaced in the tectono-stratigraphic evolution of the South-Pyrenean basin.

Stop 6.2:

Cross-section of the Monte Perdido and Gavarnie thrust sheets. La Larri tectonic window.

Access

Continue north to Bielsa (N-138) and turn left to Pineta-Parador Nacional Monte Perdido. Take the track immediately before the Parador and park once the river is crossed. Walk for 1 hour.

Structural landscape

This is one of the most spectacular geological scenes of the Pyrenees and a key locality for

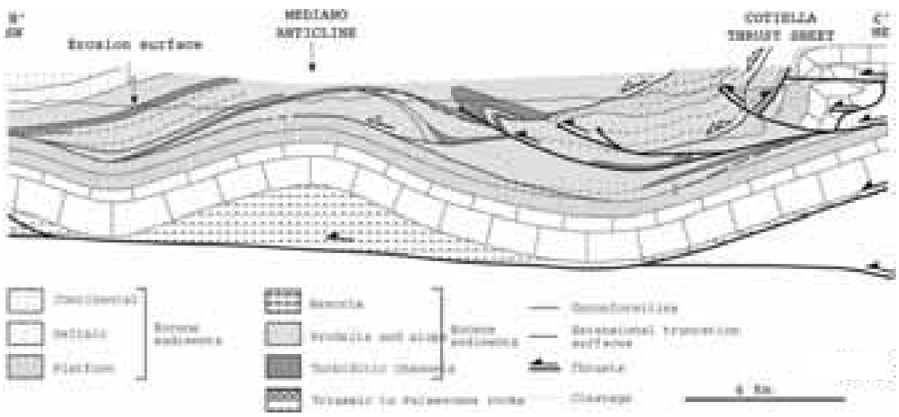


Figure 6.1c - Structure of the Cotiella Thrust Sheet and of its footwall.

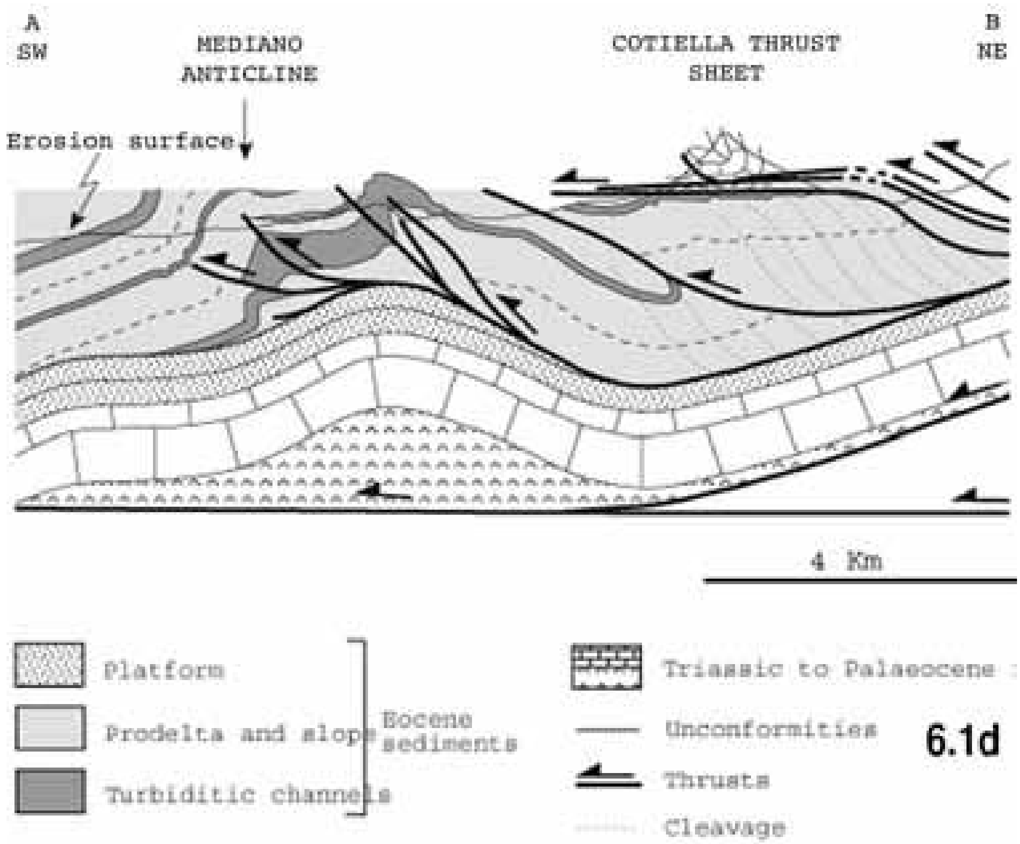


Figure 6.1d - Kinematics of the Cotiella Thrust sheet and the Mediano anticline.

understanding the basement involvement into the Alpine thrust structure and the structural relationships between the cover and the basement thrust sheets.

The course will follow the Pineta glacier cirque across the lower part of the Upper Cretaceous succession of the Gavarnie thrust sheet. The trail climbs up into the lateral glacier valley of La Larri where the Triassic red beds in the footwall of the Gavarnie thrust outcrop.

Climb the grassy slopes southwards of that valley towards La Estiba in order to get a view of the

Figure 6.2a - The Gavarnie Thrust at La Estiba.

northern ridge. From this viewpoint, the Devonian basement of the Gavarnie thrust sheet below the Upper Cretaceous rocks is visible. It rests on top of the Upper Cretaceous limestones-Triassic red beds and the metamorphic basement outcropping at the bottom of the valley. This fault contact corresponds with the Gavarnie thrust. Note its antiformal attitude, the folded unconformity above the Hercynian basement in its hangingwall and the stratigraphic differences between the hangingwall and the footwall (Fig. 6.2a, and Front cover, fig.3).

Stop 6.3:

Structure of the granite basement in the footwall of the Gavarnie thrust. Bielsa-Parzán cross-section.

Access

Drive back to Bielsa and reach main road N-138 leading to the French border.

Structural landscape

Several observations will be made along the roadcuts between the villages of Bielsa and Parzan in order to characterize the fold and thrust structure of the Upper Cretaceous, Triassic red beds and granitic basement of the outcropping lower thrust sheets.

DAY 7

End of the field trip and return of the participants to Florence

Acknowledgements

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Itinerary of the field trip B16

FIELD TRIP MAP

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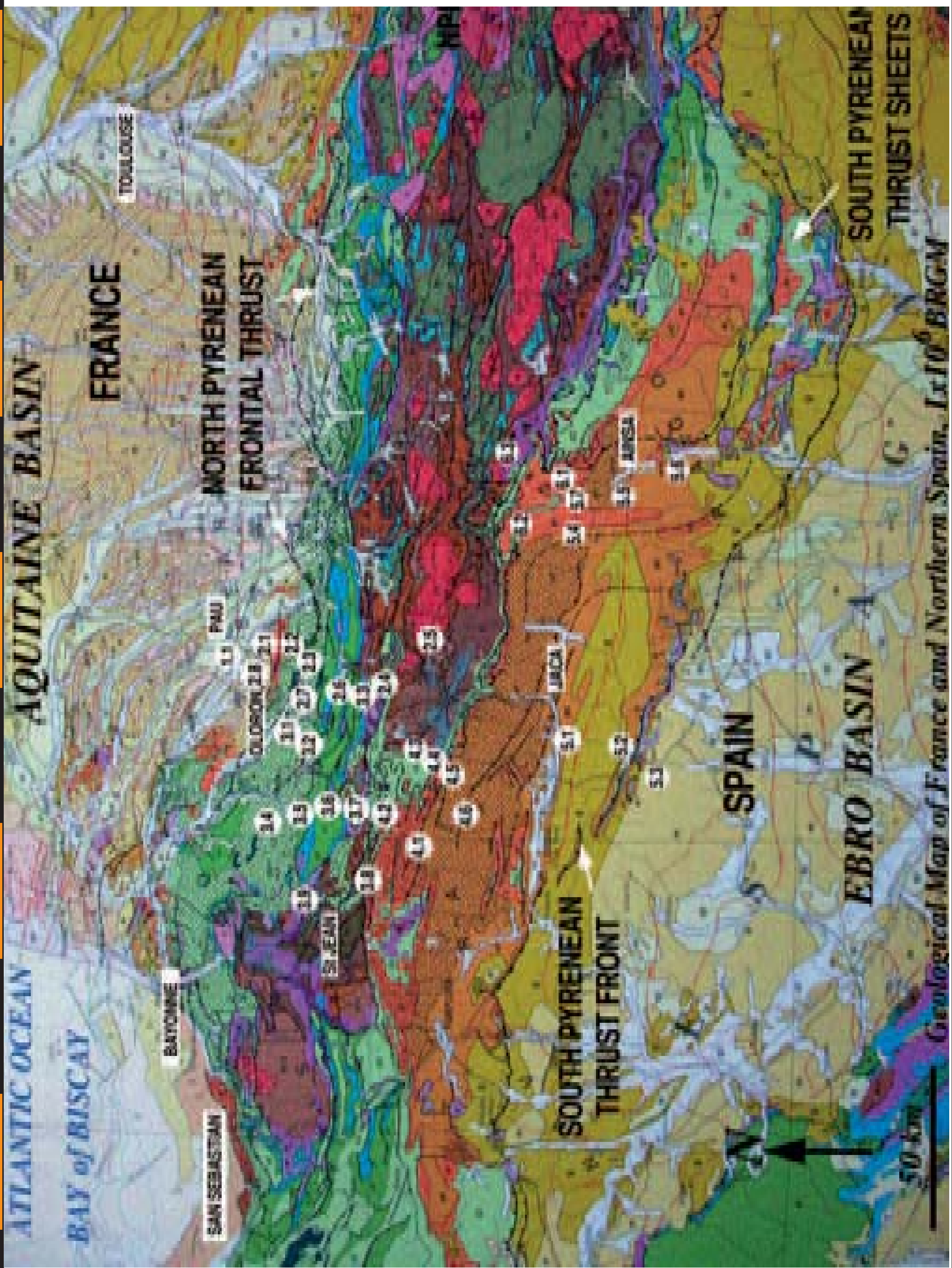

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