Pyrenean orogeny and plate kinematics

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[1] The evolution of the Pyrenees, a mountain range between Iberia and Eurasia, has remained the subject of many debates between geologists and geophysicists for a long time. By combining the identification of seafloor spreading anomalies A33o to M0 in the Bay of Biscay with those in the North Atlantic, we have derived a position of a mean pole of rotation for the entire opening of the Bay of Biscay. Four hundred kilometers of shortening took place between the Iberian and Eurasian plates in the Pyrenean domain during the opening of the Bay of Biscay, from chrons M0 to A33o time (118 to 80 Ma). The deep seismic Etude Continentale et Océanique par Réflexion et réfraction Sismique (ECORS) profile shot across the Pyrenees and teleseismic data show the presence of two distinct slabs, which dip to the north. The southern slab is linked to the subduction of the neo-Tethys Ocean, which was created from late Jurassic to early Aptian. Simultaneously, elongated back arc basins formed along the future Pyrenean domain. This slab was active from at least 118 Ma (early Aptian) to 100 Ma (late Albian). The northern slab, active since 85 Ma, is linked to the subduction of the lower continental crust located south of the Pyrenean domain. In the upper crust, normal faults as well as the north Pyrenean fault became reverse faults, and former back arc basins were inverted, giving rise to the uplift of the Pyrenees as a double asymmetrical wedge. *INDEX TERMS:* 8102 Tectonophysics: Continental contractional orogenic belts; 8157 Tectonophysics: Plate motions—past (3040); 8180 Tectonophysics: Tomography; 8105 Tectonophysics: Continental margins and sedimentary basins (1212); 8159 Tectonophysics: Rheology—crust and lithosphere; *KEYWORDS:* Pyrenean evolution, kinematics, seismic data, tomography

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1. Introduction

1.1. The Pyrenees

[2] The Pyrenees is a 400-km-long but relatively narrow (~150 km) N110°E trending continental collisional fold belt located between France (part of Eurasia) in the north and Spain (part of Iberia) in the south (Figure 1). It consists of an asymmetrical double tectonic wedge more developed on its southern side. Because of its position east of the Bay of Biscay, which opened by seafloor spreading due to the rotation of Iberia (IB) relative to Eurasia (EU) during middle and late Cretaceous [e.g., *Boillot and Capdevila*, 1977; *Le Pichon and Sibuet*, 1971; *Olivet et al.*, 1984; *Srivastava et al.*, 1990b], it is now widely accepted that the Pyrenees must have formed the eastward extension of the plate boundary between EU and IB during these times. Its history of development can thus be related to the plate kinematics, not only of the Bay of Biscay but also of the North Atlantic as a whole. Over the past 30 years, several papers have been published relating the derived motions between the plates across this belt at various geological times to its history of development [e.g., *Le Pichon et al.*, 1971; *Olivet*, 1996; *Roest and Srivastava*, 1991; *Sibuet and Collette*, 1991; *Srivastava et al.*, 1990a]. Yet, none of these papers has been able to explain completely the different geological episodes to which geologists regard this belt to have been subjected.

1.2. Geology of the Pyrenees

[3] The Alpine Pyrenees constitute a narrow, asymmetrical, double-wedged range, including a relatively wide southern proforeland basin and a northern retroforeland basin. Though globally cylindrical, folded Mesozoic strata are not cylindrical along the whole chain. The Pyrenean range (Figure 1) is traditionally divided into four longitudinal structural zones striking N110°E, which are from south to north: (1) the south Pyrenean zone (SPZ) consisting

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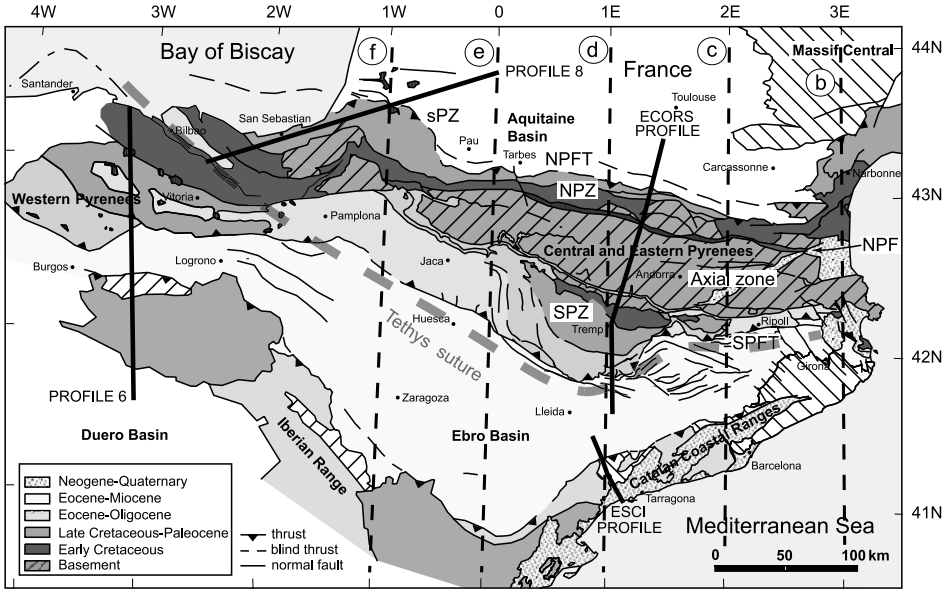

Figure 1. Structural map of the Pyrenees and its main tectonic units (adapted from *Vergés et al.* [1995], reprinted with permission from Elsevier Science). Also shown are the locations of the ECORS [*Choukroune and ECORS Team*, 1989] and ESCI [*Vidal et al.*, 1998] deep seismic reflection profiles, refraction profiles 6 and 8 [*Pedreira et al.*, 2003] (bold lines), and tomographic profiles b to f (dashed lines). The location of the Tethys suture (gray dashed line) is based on the seismic, tomographic, and gravity [*Banda and Daignières*, 1995] data. NPF, north Pyrenean fault; NPFT, north Pyrenean frontal thrust; NPZ, north Pyrenean zone; SPFT, south Pyrenean frontal thrust; SPZ, south Pyrenean units; sPZ, sub-Pyrenean zone.

of Mesozoic and Tertiary terrains and comprising the southern proforeland basin [*Muñoz et al.*, 1986] (a décollement zone emerges at the south Pyrenean frontal thrust (SPFT) north of the undeformed Ebro molassic basin), (2) the axial zone essentially made up of Paleozoic and pre-Paleozoic terrains affected by the Hercynian deformation and metamorphism [*Mattauer and Séguret*, 1966], (3) the north Pyrenean zone (NPZ) comprising Mesozoic and Tertiary strata overlying the Hercynian basement locally exhumed to form the "north Pyrenean massifs," and (4) the sub-Pyrenean zone (sPZ) in which Mesozoic and Tertiary deposits overlie the Paleozoic basement. The axial zone and the north Pyrenean and sub-Pyrenean zones are separated by two major faults which are from south to north, the north Pyrenean fault (NPF) and the north Pyrenean frontal thrust (NPFT). Mantle lherzolite outcrops along the NPF implying that the NPF corresponded in depth to a crustal discontinuity. Currently, it may be allochtonous so that its current base may be displaced from the crustal discontinuity at depth (discussion by, e.g., *Mattauer* [1990]).

1.3. Tectonic Episodes

[4] The presence of several elongated basins, which lie parallel to and at the location of the present-day north Pyrenean zone have been interpreted as pull-apart basins formed from early Aptian to Cenomanian (120–90 Ma) [*Choukroune and ECORS Team*, 1989; *Choukroune and Mattauer*, 1978; *Mattauer*, 1968], when IB was perhaps displaced relative to EU in a left-lateral sense by a strike-slip motion of several hundred kilometers (Figure 2). All the other basins located in the south Pyrenean zone (Organyà basin) but also in southern France (Mauléon and Parentis basins) and northern Spain (Iberian, Catalan, and Basque-Cantabrian basins) were formed in an extensional setting during late Jurassic–early Cretaceous [*Beràstegui et al.*, 1990; *Martín-Chivelet et al.*, 2002; *Vergés and García-Senz*, 2001]. The question of the origin of the so-called pull-apart basins has been often questioned. For example, the approximate continuity of Jurassic facies on each side of the Pyrenees has also been used to suggest a limited amount of early Aptian to Cenomanian strike-slip motion at the emplacement of the Pyrenees [*Souquet and Mediavilla*, 1976]. *Mattauer and Séguret* [1971] also stated that geological data are incompatible with several hundreds kilometers of strike-slip motion during early Cretaceous. All these basins, including those of the north Pyrenean zone, are underlain by thinned continental crust, and apparently no oceanic crust was emplaced during this tensional episode. Shortening started later during late Santonian [e.g., *Garrido and Rios*, 1972], though local occurrences of compressive deformation during middle early Albian (around 106 Ma) are observed in the north Pyrenean area [*Souquet and Peybernès*, 1991]. The maximum shortening occurred during Eocene-Oligocene times [*Fitzgerald et al.*, 1999], simultaneously with the subduction of the Bay of Biscay oceanic domain beneath Iberia [e.g., *Alvarez-Marrón et al.*, 1997; *Sibuet and Le Pichon*, 1971; *Sibuet et al.*, 1971].

1.4. Problem to Solve

[5] The question that needs to be answered is: Are the implied motions between the EU and IB plates as obtained from the plate kinematics of the North Atlantic at various

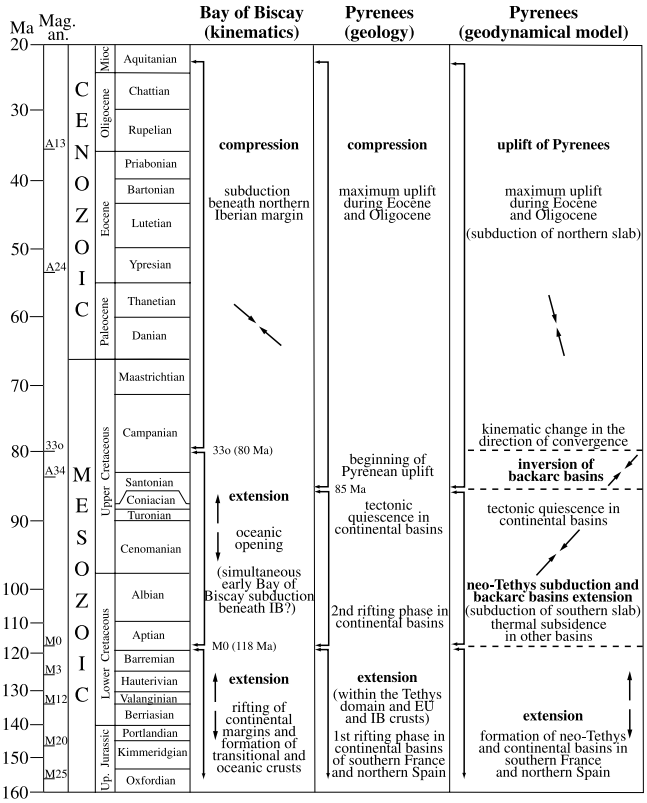

Figure 2. Summary of the principal kinematic, geological, and geodynamical events in the Bay of Biscay and the Pyrenees.

geological times compatible with geological observations? If not, what could be the possible cause? Seafloor spreading in the Bay of Biscay stopped soon after chron A33o (old), as demonstrated by the absence of magnetic anomalies younger than A33o in the Bay of Biscay. In addition, to the west of the Bay of Biscay, younger isochrons are not deflecting into the Bay of Biscay (Figure 3), thus providing strong evidence that seafloor spreading had stopped in the Bay of Biscay soon after chron A33o. This observation, however, does not imply that the motion between IB and EU also stopped at this time. In fact, the plate kinematics of the North Atlantic for chron A32 and younger suggests that IB kept moving separate from EU, but as part of Africa, until chron A6. The IB/EU plate boundary extended west of the Pyrenees along the north Spanish marginal trough and Azores-Biscay rise. At the time of chron A6 the plate boundary between IB and EU shifted to the south, along the present Azores-Gibraltar fracture zone thereby terminating any further relative motion between IB and EU [*Roest and Srivastava*, 1991; *Srivastava et al.*, 1990a]. The relative motion between IB and EU resulted in convergent motions of different magnitudes across the Pyrenees during these times. The magnitude and direction of such derived convergent motions agree with the geological observations since the end of seafloor spreading in the Bay of Biscay (chron A33o, 80 Ma [*Roest and Srivastava*, 1991]). However, considerable differences exist between the kinematic solutions given by various people for times earlier than chron A33o [e.g., *Olivet*, 1996; *Roest and Srivastava*, 1991; *Sibuet and Collette*, 1991] and the geological observations.

[6] Anomaly A34 is the oldest seafloor spreading anomaly which has been identified so far [*Srivastava et al.*, 1990b] with any confidence, though we do see a number of positive and negative anomalies (underlined in Figure 3) lying on both sides of the bay landward of anomaly A34. To our knowledge, nobody tried to identify them. If these lineations are older seafloor spreading magnetic anomalies, then they would allow the calculation of a better constrained model for the motion across the Pyrenees than has been possible so far. We have therefore reexamined the above question of the difference between the geological observations and the derived motion from plate kinematics for periods older than 80 Ma by combining the recent identification of chron M0 (part of J anomaly) west off Iberia and southeast of Flemish Cap off Newfoundland [*Srivastava et al.*, 2000] with that in the Bay of Biscay. This allows us to determine more accurately the position of the pole of rotation and to compare the predicted motion across the Pyrenees with other geological and geophysical observations made across it. These observations include the deep crustal structures as obtained from a reinterpretation of the deep ECORS seismic profile shot across the Pyrenees and the new tomographic sections as established from teleseismic data across it. Our major conclusion from examination of these data is that during the early Cretaceous the neo-Tethys Ocean subducted beneath Eurasia simultaneously with the opening of elongated back arc basins located at the present-day location of the northern Pyrenees. Later on, these back arc basins located on the thinned Eurasian continental crust were inverted during the Pyrenean orogeny.

2. Position of Iberia Relative to Eurasia at Pre-Chron A33o Time (>80 Ma)

2.1. Magnetic Anomalies in the Bay of Biscay

[7] A large amount of magnetic data exists in the Bay of Biscay [*Verhoef et al.*, 1996] showing the presence of seafloor spreading anomalies (Figure 3). The prominent negative anomaly in the center of the bay has been recognized as anomaly A33o for a long time. Bands of positive anomalies lie on either side of this negative anomaly. The junctions between the negative and the positive trends have been interpreted as anomaly A34 [*Srivastava et al.*, 1990a] being on either side of anomaly A33o. Careful inspection of this large amount of data has allowed us to identify the presence of anomalies M0–M3. In Figure 4 we show selected profiles from this database, oriented in the northwest to northeast direction. For clarity, we have projected the positive and negative anomalies in the E-W direction. Identification of anomalies A34, M0, and M3 are based on the correlation of observed and modeled anomalies as shown in Figure 5. We have calculated a magnetic model along 10°W longitude (parameters in legend of Figure 5). We do not see any seafloor spreading anomalies older than M3 in the bay. The good correlation between observed and calculated anomalies allows us to point out the following points: (1) The large negative anomaly between chrons A34 suggests that spreading in the Bay ceased at chron A33o time as already mentioned [e.g., *Sibuet and Collette*, 1991].

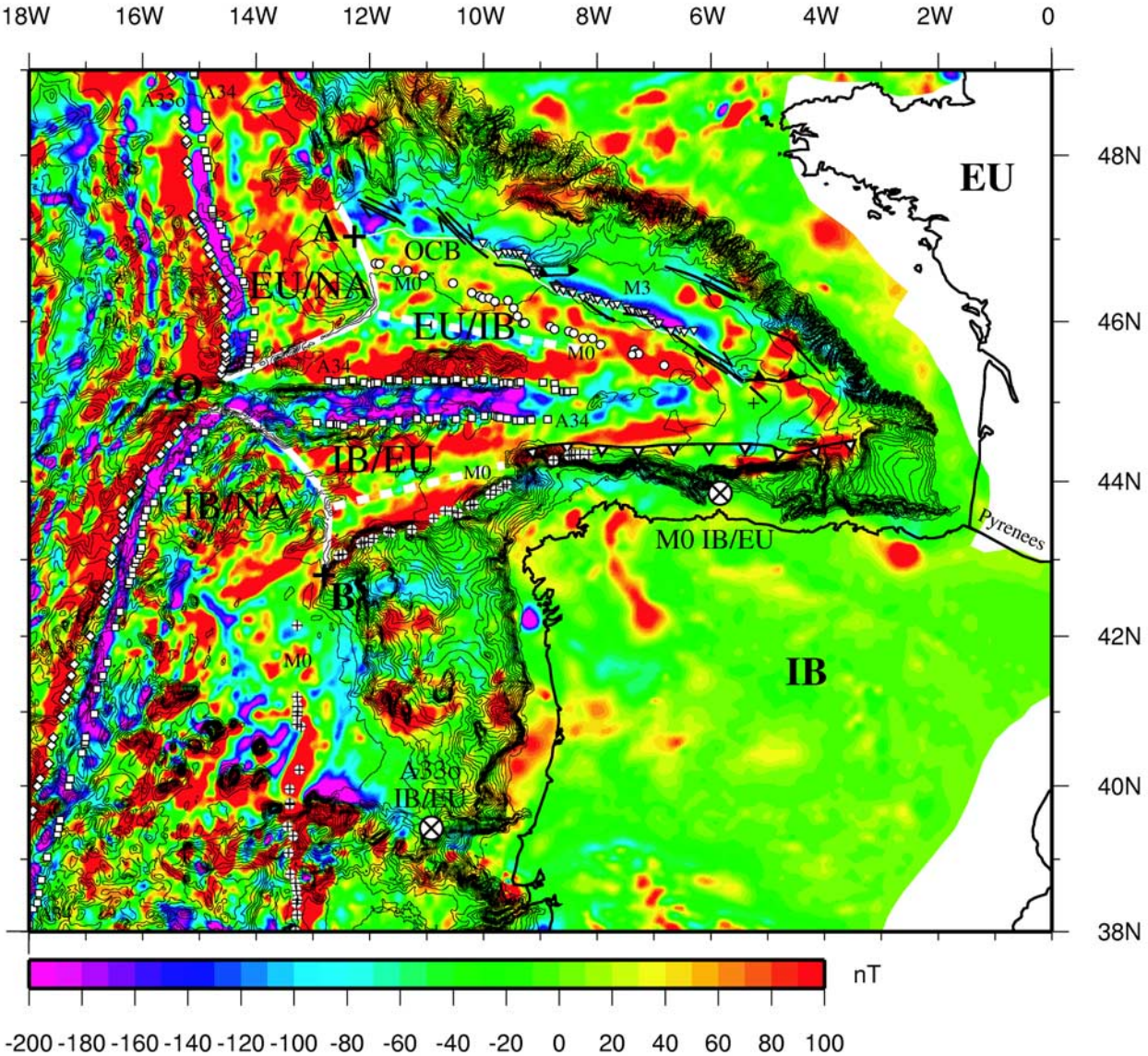

Figure 3. Main magnetic and bathymetric features in the Bay of Biscay and the northeast Atlantic Ocean as extracted from *Verhoef et al.* [1996] and *Sibuet et al.* [2004]. The bathymetric contours are at 200-m intervals. EU, Eurasian; IB, Iberian; and NA, North American plates. Also shown are the fossil triple junction (TJ) trajectories with their conjugate points A and B on the EU and IB plates [*Sibuet et al.*, 2004] in white lines; position of IB/EU A33o pole (39.42°N, 10.91°W) and IB/EU M0 pole (43.85°N, 5.83°W) as large crosses within circles. O, A33o TJ location; white thin line, ocean-continent boundary (OCB). White diamonds are A33o picks, and white squares are A34 picks [*Srivastava et al.*, 1990a; this study], small crosses within circles are M0 picks west of IB [*Srivastava et al.*, 2000] and north of IB (this study), white small dots are northern Bay of Biscay M0 picks (this study), and inverted white triangles are M3 picks (this study). Black lines are features active since 80 Ma in the northern Bay of Biscay transitional domain, and the black line with white triangles is the north Spanish marginal trench also active since at least 80 Ma. Thick dashed lines underline the change in direction of the magnetic lineations in the Bay of Biscay.

Using the *Kent and Gradstein* [1986] timescale, the calculated spreading rates between these anomalies lie in the same range as in the North Atlantic (7.7–13 mm/yr) [*Srivastava et al.*, 1990a], (2) M0 lineations, symmetrical with respect to A34, are identified both in the northern and southern Bay of Biscay, and (3) lineations M0–M3 exist in northern Bay of Biscay with half-spreading rate of 11.7 mm/yr. Though not highlighted in Figure 4, anomalies M1 and M2 can be seen at places in the northern bay. Chron M3 corresponds to the negative anomaly located at the southern boundary of the transitional crust of the Armorican basin, where 7.2–7.5 km/s crustal velocities have been determined [*Thinon*, 1999]. Similar half-spreading rates have been obtained for the M0–M3 period between IB

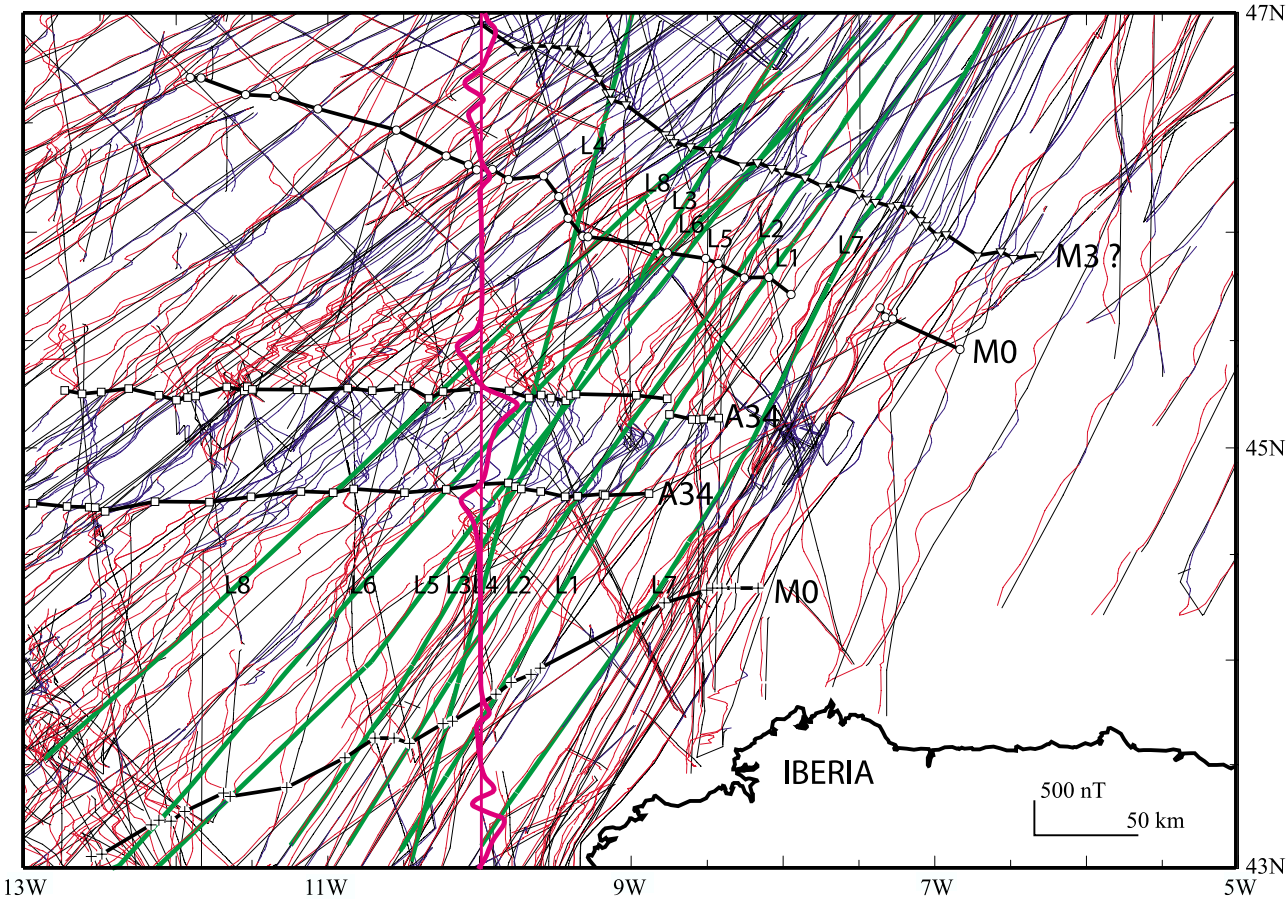

Figure 4. Selected marine magnetic profiles extracted from *Verhoef et al.* [1996] in the Bay of Biscay and projected in the E-W direction. Positive and negative magnetic anomalies are indicated by red and blue, respectively. The synthetic model, which appears in purple along 10°W longitude, has been computed with parameters of Figure 5. A34, M0, and M3 picks are shown as well as the corresponding identified lineations.

and NA (9–15 mm/yr) [*Srivastava et al.*, 2000], which adds credence to our identification of the M0–M3 sequence in northern Bay of Biscay. No corresponding M1–M3 lineations can be identified in the southern Bay of Biscay. If they existed, they have possibly subducted now under Iberia.

[8] The consequences of our new identification of chron M0 in the Bay of Biscay are significant. Seafloor spreading may have started in the Bay of Biscay as early as anomaly M3 (124 Ma, lower Barremian). The presence of transitional crust between chron M3 and the base of the northern Biscay continental slope suggests that a 100-km-wide stripe was previously created as exhumed mantle or at very slow spreading rate. As a main consequence of this observation, rifting on the Bay of Biscay margins ended sometime during lowermost Cretaceous [*Sibuet et al.*, 2004], which is much earlier than generally accepted (Aptian time for *Montadert et al.* [1979]), thus reducing the duration of rifting of the continental margins to a period of no more than 10 Myr. More generally, there is now a large discussion concerning the age of the end of the main rifting phase on continental margins, which seems to be much older than previously thought (e.g., Berriasian instead of late Aptian for the west Iberian Abyssal Plain margin [*Wilson et al.*, 2001]) and might be associated with a transient thermal uplift caused by the occurrence of hotter oceanic lithosphere after the emplacement of transitional crust [*Reston and Morgan*, 2004]. For continental basins the end of the rifting phase might be at least partly hidden by thermal subsidence processes, which follow the rifting period [e.g., *Lin et al.* [2003]. The differential amount of thermal subsidence across a rifted basin can also generate motions along normal faults, which might be artificially attributed to a prolongation of the rifting episode.

2.2. Kinematic Constraints

[9] Even though small differences exist in the proposed positions of the continents relative to one another in different reconstructions of the North Atlantic at chron A33o (80 Ma) [e.g., *Olivet*, 1996; *Sibuet and Collette*, 1991; *Srivastava et al.*, 1990b, 2000], they all seem to be very similar. The small differences arise because of differences in the location of the plate boundaries used and they do not change significantly the over all implied motions across the Pyrenees since chron A33o time as given by *Roest and Srivastava* [1991]. This is not the case, however, for older times such as chron M0 (118 Ma) where large differences exist in the position of IB relative to EU while positions of North America (NA) and Africa (AF)

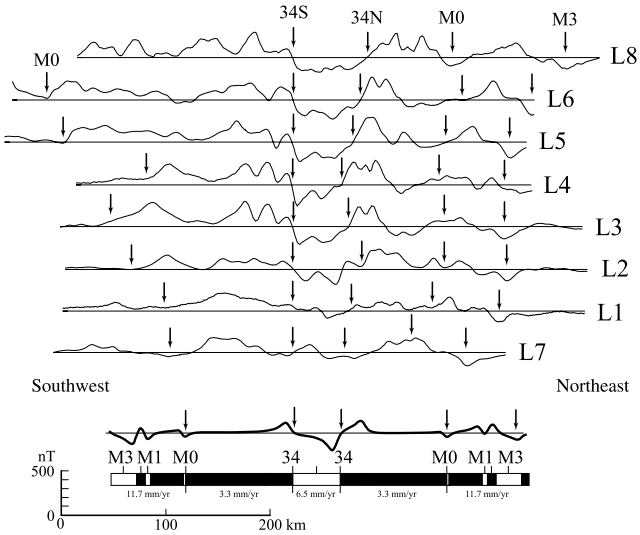

Figure 5. Magnetic anomalies along profiles L1 to L8 located as green track lines in Figure 4. The synthetic model has been computed along 10°W longitude with a basement depth at 6 km and a 2-km-thick magnetized layer. Blocks in black and white correspond to positive and negative magnetization, respectively. Half-spreading rates are also given. Other parameters used in the model calculations are the remnant declination and inclination: $D_r = -43°$ and $I_r = 58°$. The calculated anomaly is displayed without amplitude scaling.

remain roughly the same in such reconstructions. This is largely because of the differences in the constraints used in deriving the poles of rotation for IB in these reconstructions. To illustrate this point, we will use two extreme cases here: the reconstruction as given by *Olivet* [1996] and that given by *Srivastava et al.* [2000] of these plates. The poles of *Srivastava et al.* [1990a, 2000] (Figure 6a) are based on the fit of the anomaly M0 identifications across the North Atlantic and constrained by maintaining the direction of motion between the plates along the Azores-Gibraltar fracture zone. *Olivet*'s [1996] poles of reconstruction for chron M0 (Figure 6b) are derived on the geological assumption that prior to chron A33o, there was largely a strike-slip motion between IB and EU in an east-west direction. Surprisingly, the correspondence of geomorphological features located between Iberia and its adjacent plates is satisfactory in both reconstructions: a good fit between the shape of the northern and southern Bay of Biscay continental margins, an alignment of the Goban Spur/N-E Flemish Cap continental margin trend, either with the trend of the Interior basin located between Iberia and Galicia Bank [*Olivet*, 1996] or with the western Galicia margin [*Sibuet and Collette*, 1991; *Srivastava et al.*, 1990a]. However, the two models correspond to two different modes of opening for the Bay of Biscay during this time: a scissors-type opening of the bay with a pole of rotation located in the southeastern Bay of Biscay corner (Figure 6a, hypothesis 1), giving rise to simultaneous convergence in the Pyrenean domain [*Argand*, 1924; *Carey*, 1958; *Choubert*, 1935; *Masson and Miles*, 1984; *Matthews and Williams*, 1968; *Roest and Srivastava*, 1991; *Schoeffler*, 1965; *Sibuet and Collette*, 1991; *Sibuet and Srivastava*, 1994; *Srivastava et al.*, 1990a, 1990b, 2000] or largely a left-lateral strike-slip motion in the bay and along the north Pyrenean fault with an IB/EU pole of rotation located in northern Europe (Figure 6b, hypothesis 2) [*Choukroune et al.*, 1973; *Le Pichon et al.*, 1971, 1970; *Le Pichon and Sibuet*, 1971;

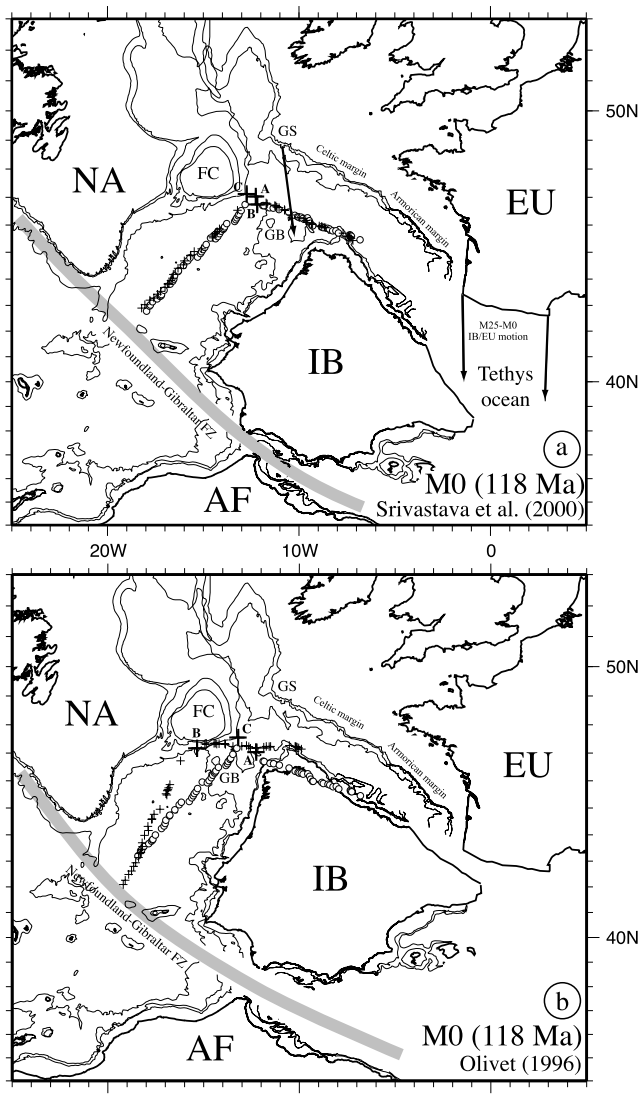

Figure 6. Reconstructions of the North Atlantic between Africa (AF) and British Isles at chron M0 with respect to Eurasia (EU) as given by *Srivastava et al.* [2000] (IB/EU 43.85°N, 5.83°W, −44.76°; AF/EU 43.61°N, 6.93°W, −40.89°; NA/EU 69.67°N, 154.26°E, 23.17°) and *Olivet* [1996] (IB/EU 47.79°N, 0.22°W, −26.81°; AF/EU 44.41°N, 6.13°W, −40.80°; NA/EU 69.73°N, 157.17°E, 22.75°). M0 picks west of IB (crosses) and east of NA (white circles) are from *Srivastava et al.* [2000]; north of IB (crosses) and south of EU (white circles) from this study. A, B, and C are conjugate points that should coincide before the oceanic opening of the North Atlantic and Bay of Biscay. FC, Flemish Cap; GS, Goban Spur; GB, Galicia Bank. Note the discrepancy in the fit of M0 lineations as well as between rotated positions of conjugate points A, B, and C in *Olivet*'s [1996] reconstruction (Figure 6b).

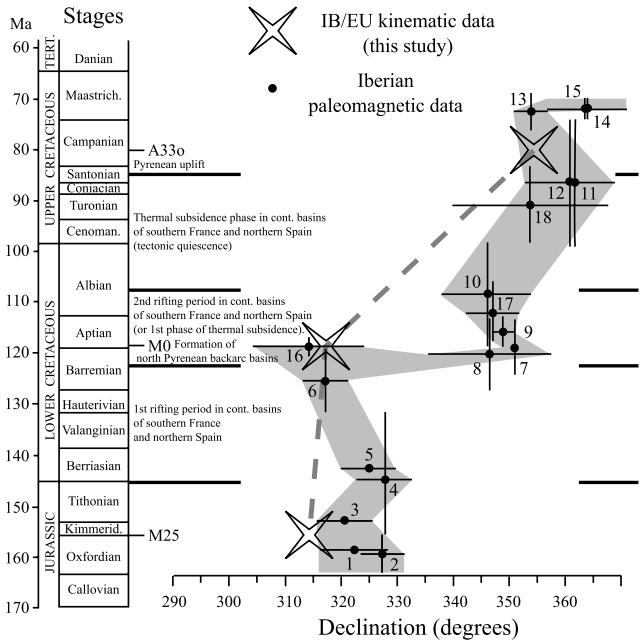

Figure 7. Comparison of rotation of Iberia with respect to Europe as deduced from kinematic parameters at chrons M25 (156.5 Ma), M0 (118 Ma), and A33o (80 Ma) (large transparent stars) and those from the paleomagnetic declinations [*Dinarès-Turell and García-Senz*, 2000] (black dots) versus time, calculated for a common reference site at Tremp (42.2°N, 1°E). The α_{95} error envelope is drawn for the paleomagnetic data. Numbers besides the points identify sources of data mentioned by *Dinarès-Turell and García-Senz* [2000]. Note the good correspondence between the two types of data and, based on paleomagnetic data, the probable existence of several kinematic phases, in particular during the M0–A33o time interval period.

Olivet, 1996; *Olivet et al.*, 1984]. In both cases, the amplitudes of relative motion across the Pyrenees remain similar, i.e., several hundreds kilometers.

[10] In hypothesis 1 the IB/NA position is based on the fit of about 60 individual anomaly M0 identifications [*Srivastava et al.*, 2000] (Figure 6a). Anomaly M0 is characterized by a small minimum located at the younger side of the positive J anomaly. Since there are no clear fracture zone offsets along the M0 magnetic trends, the shape of the magnetic lineations, in particular a small curvature in the middle of each of the two conjugate lineations, and the correspondence between the Newfoundland and Gibraltar (Gloria-Pico) fracture zones are two independent elements which have been used to fit IB/NA M0 picks. In the past, it was assessed that the number of published M0 picks was too small to really constrain the M0 fit [*Olivet*, 1996]. However, there is only a slight difference between the M0 reconstructions using the Newfoundland and Gloria-Pico fracture zones correspondence together with only a few M0 picks [*Roest and Srivastava*, 1991; *Sibuet and Collette*, 1991; *Srivastava et al.*, 1988] and the last *Srivastava et al.* [2000] reconstruction where 60 M0 picks were used.

[11] The fit between the new M0 picks of *Srivastava et al.* [2000] and those identified in the Bay of Biscay are shown in the two reconstructions of Figure 6. In hypothesis 2 [*Olivet*, 1996] (Figure 6b) the IB/NA reconstruction was based on the fit of the broad J anomalies, which are about 50 km wide on each side of the north Atlantic Ocean in the south but very narrow in the north both off Iberia and Newfoundland where they become gradually unrecognizable (Figure 3). If the fit of the M0 magnetic lineations seems to be correct in the south, though longitudinally inappropriate, it deteriorates in the north with a misfit of about 200 km between northern Iberia and Newfoundland. Lineations M3–M20 [*Srivastava et al.*, 2000, Figure 5] are roughly parallel to M0, which suggests that in the reconstructions of *Olivet* [1996], Iberia might be clockwise rotated of about 15 additional degrees with respect to NA (and also with respect to EU). This would mean that the 35°–40° total rotation of Iberia with respect to Eurasia based on paleomagnetic results [e.g., *Galdéano et al.*, 1989; *Juarez et al.*, 1998; *Zijderveld and van der Woo*, 1971] would be accounted for.

[12] A second argument, which gives credence to hypothesis 1, is the difference in the fit of the M0 anomaly picks (Figure 3) between the two sides of the Bay of Biscay in Figure 6. By using the published pole of *Srivastava et al.* [2000] these picks fit remarkably well in Figure 6a but not in Figure 6b, showing that seafloor spreading may have started as early as anomaly M0–M3 time in the Bay of Biscay. This also strengthens the interpretation of *Sibuet and Collette* [1991] for the existence of a triple junction at the mouth of the Bay of Biscay, since its opening, until chron A33o as the three conjugate points lie fairly close to each other in Figure 6a but not in Figure 6b. The location of this triple junction (TJ) at chron A33o marked by letter O in Figure 3 is identified by the intersection of the three rift axes corresponding to magnetic lineation A33o formed between the three plate pairs (EU/NA, EU/IB, and IB/NA) when these are brought together at chron A33o time. TJ trajectories separating oceanic domains formed between these three plate pairs were first identified by *Sibuet and Collette* [1991]. Here, they have been updated (Figure 3) using new swath bathymetric [*Sibuet et al.*, 2004] and multichannel seismic (MCS) data, together with already published magnetic data [*Verhoef et al.*, 1996] for the eastern part of the North Atlantic. For the western side of the North Atlantic we used largely magnetic and MCS data. In Figure 3 the northern TJ branch extends from point A (47.03°N, 12.26°W), while the southern branch starts at point B (42.79°N, 12.80°W) and the western branch, not shown here, starts at point C (46.82°N, 43.03°W). Figure 6 shows location of these points in the two reconstructions. They lie within a circle of 25-km radius in Figure 6a suggesting a few tens kilometers of pre-M0 seafloor spreading in the Bay of Biscay. In contrast, the three points are more than 200 km apart in Figure 6b.

[13] A third argument, which also gives credence to hypothesis 1 comes from late Jurassic and Cretaceous paleomagnetic declination data of the stable Iberia [*Dinarès-Turell and García-Senz*, 2000], which are plotted in Figure 7 together with our kinematic IB/EU rotations at chrons M25, M0, and A33o (both for the same reference site at Tremp 42.2N, 1°E). The three "kinematic" points fall within the α_{95} error envelope of paleomagnetic data, confirming the validity of our kinematic parameters. In addition, paleomagnetic data show a fast Barremian/Aptian age

counterclockwise rotation of about 25°, though error in paleomagnetic data could reach 5°. Another counterclockwise rotation of 13° occurred between Albian and Maastrichtian [*Dinarès-Turell and García-Senz*, 2000]. These two distinct phases are not evidenced in our kinematic work, as no recognizable magnetic anomalies exist in the M0–A34 quiet magnetic zone. However, a change of kinematic phase is underlined in the Bay of Biscay, both by the abrupt change in the direction of magnetic trends (dashed white lines in Figure 3, at about a third of the distance from chron M0 in direction of the Bay of Biscay axis) and by the associated change in direction of the triple junction branches. Though it is not the purpose of this work to establish closely time spaced plate kinematic reconstructions, the numerous paleomagnetic declination data of stable Iberia are not only in agreement with our kinematic model but will be useful in the future to refine the kinematic model.

[14] In conclusion, the kinematic model presented here is based on the identification of new M0 magnetic lineations in the Bay of Biscay and corroborated by two other independent data sets: the Bay of Biscay triple junction trajectories and the paleomagnetic declination data of the stable Iberia. As hypothesis 2, which leads to a M0–A33o transcurrent motion in the Pyrenean domain, has to be rejected, we have to explore an alternate way to explain how tensional basins can be formed in a convergent setting.

2.3. Motion in the Pyrenean Domain as Obtained From Plate Kinematics

[15] Figure 8 illustrates the motions across the Pyrenees over time, as implied from plate kinematics. We have divided these motions in three parts; one from chron M25 to chron M0 (Figure 8a), one from chron M0 to chron A33o (Figure 8b) and the last one from chron A33o to Present (Figure 8c). The M25–M0 (156.5–118 Ma) IB/EU stage motion (3.74°N, 85.68°E, 3.44°) has been estimated by using the *Srivastava and Verhoef* [1992] parameters combined with those of the M0 NA/IB pole of *Srivastava et al.* [2000], even though closures of the EU, NA, and IB plates are too tight. A 350-km N-S tensional motion is calculated between IB and EU plates (Figure 8a). A number of reconstructions for chron A33o exist, and they are all very similar. For illustrative purpose, Figure 8b is based on the IB/EU pole used by *Sibuet et al.* [2004]. We have also shown the locations of four stage pole rotations for chron M0–A33o based on the M0 pole of rotation of *Srivastava et al.* [2000] and those for anomaly A33o poles of *Sibuet et al.* [2004], *Srivastava et al.* [2000], *Sibuet and Collette* [1991], and *Olivet* [1996]. As can be seen, they all lie within a circle of 25-km radius showing the degree of similarity in their positions for chron A33o as mentioned earlier. From chrons M0 to A33o the SW-NE convergence in the Pyrenean domain increases eastward from 250 to 450 km (Figure 8b). West of the Pyrenees, the amount of shortening decreases to zero in the area of the IB/EU M0–A33o stage pole of rotation. Since chron A33o time, the SE-NW convergent motion in the Pyrenean domain gradually decreases from 150 km in the eastern Pyrenees to 100 km in the western Pyrenees (Figure 8c). Simultaneously, the oblique NW-SE subduction of the Bay of Biscay oceanic domain occurred beneath IB in the eastern Bay of Biscay, evolving to strike-slip motion to the west (Figures 8c and 8d).

[16] Figure 8d shows the areas of intense deformation (dark pink) since chron A33o time. The NW subduction of the Bay of Biscay transitional and oceanic domains beneath IB in the eastern Bay evolves to strike-slip motion close to B [*Alvarez-Marrón et al.*, 1997; *Sibuet and Le Pichon*, 1971]. Between B and the Azores-Biscay rise, the deformation is located south of the triple junction trajectory [*Roest and Srivastava*, 1991] but also in the area of the triple junction where NE-SW extension is suggested by the presence of E-W negative magnetic anomalies breaking the N-S A34, A33o, and younger lineations between 15° and 16°W longitude (Figure 3). Two tectonic events (light pink) have been identified during late Cretaceous and Eocene [*Thinon*, 2001] in the northern bay near the ocean-continent boundary (OCB) and at the base of the continental slope but also in the western Bay of Biscay between A34 lineations [*Sibuet et al.*, 1993] (Figure 8d). The post-A33o King's trough microplate proposed by *Srivastava et al.* [1990b] might extend in the Bay. Its northern boundary would lie between A34 lineations extending perhaps to Cantabria seamount (45°N, 8°W) and farther east to the Gascony seamount (45.5°N, 5.5°W). Its southern boundary might follow the north Spanish marginal trough [*Alvarez-Marrón et al.*, 1997; *Sibuet and Le Pichon*, 1971; *Sibuet et al.*, 1971] up to the longitude of Cape Finisterre and then to the west the base of the Iberian margin, just north of the Bay of Biscay M0 southern lineation.

[17] The proposed kinematic reconstructions raise the fundamental question concerning the formation of the north Pyrenean Aptian-Albian tensional basins in a convergent setting. The easiest way to solve this problem is to suggest that these basins were formed as back arc basins associated with the subduction of the IB plate beneath EU starting at M0 time. The transitional and oceanic domain formed between chrons M25 and M0 would have first subducted, followed during late Cretaceous by the subduction of the adjacent continental IB plate and the associated uplift of the Pyrenees. This simple model does not leave any trace of the subduction of the previous transitional and oceanic domain as this domain would be located now at great depth beneath the Pyrenees. However, the reinterpretation of the deep seismic ECORS profile and new tomographic data presented below suggests the existence of two slabs. The southern one would correspond to the subduction of the former transitional and oceanic domain, and the northern one would correspond to the subduction of the continental IB crust. However, the suture of the southern slab, which should be located beneath the northernmost part of the Ebro basin, is not recognized [e.g., *Vergés et al.*, 1995].

3. Deep Seismic and Tomographic Data Across the Pyrenees

3.1. New Interpretation of the Deep Seismic ECORS Profile

[18] The French-Spanish deep seismic traverse of the Pyrenees (ECORS) was carried out from 1985 to 1994 [*Damotte*, 1998]. In the first published line drawings [e.g., *Choukroune and ECORS Team*, 1989; *Roure et al.*, 1989], a north dipping seismic feature was identified in the south-

ernmost portion of the profile, beneath the Ebro and Tremp basins (Figure 9a). The top of the series of strong crustal reflections extends from 6 to 18 s two-way travel time (s twtt) and is almost flat at the southern extremity of the ECORS profile [*Choukroune and ECORS Team*, 1989; *Roure et al.*, 1989]. However, the meaning of this north dipping feature was never discussed nor mentioned in all the several tens of scientific papers using either the original ECORS seismic data or these line drawings. The likely reason is that this feature is located too far (100 km) from the axial zone, without any clear relationship with the Pyrenean orogeny, and too close from the southern end of the ECORS profile. In 1992, the ESCI deep seismic profile (located in Figure 1) was shot from about 20 km southwest of the ECORS profile in direction of the Mediterranean Sea (Figure 9). It shows flat lying lower crustal reflectors at midcrustal level (6 s twtt) [*Vidal et al.*, 1998]. We interpret this north dipping feature as a décollement zone located at the top of a subducting plate (Figure 9b).

[19] From 7 s twtt underneath the northern part of the Ebro basin to 15 s twtt beneath the axial zone and more than 20 s twtt beneath the Arize (Figure 9a), the sequence of layered reflectors has been interpreted as the subducting continental Iberian lower crust [e.g., *Beaumont et al.*, 2000; *Choukroune and ECORS Team*, 1989; *Roure and Choukroune*, 1998; *Roure et al.*, 1989]. Above these lower

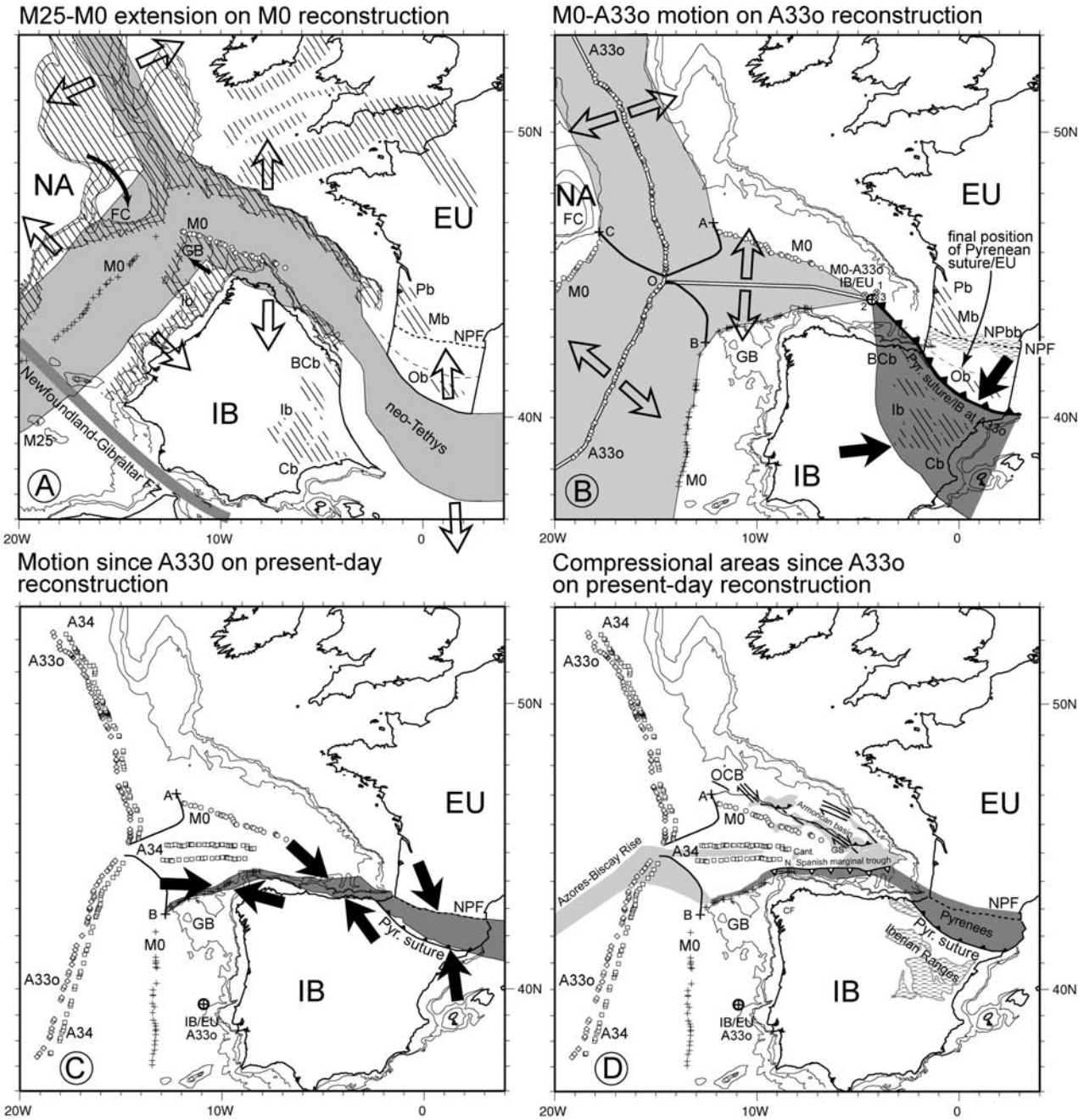

Figure 8

crustal reflectors, between the axial zone and the Tremp basin, a series of overlapping lenses might correspond either to overthrust upper crustal material or to the duplication of small slabs of lower crust during the Eocene-Oligocene Pyrenean shortening [*Vergés and García-Senz*, 2001; *Vergés et al.*, 1995]. They have been previously interpreted as stacked portions of the Iberian upper crust beneath the axial zone [*Roure and Choukroune*, 1998; *Roure et al.*, 1989], which induce the observed thickening of the upper crust beneath the Pyrenees. We suggest that these stacked slices might extend farther south, beneath the Tremp basin (Figure 9b). Thus the contact between the lower crustal layered crust and the base of the upper crustal slices is a major décollement zone as already proposed [e.g., *Roure and Choukroune*, 1998; *Beaumont et al.*, 2000]. Its southward prolongation might correspond to the upper/lower crustal limit [*Beaumont et al.*, 2000], which merges with the southern décollement zone at 6.5 s twtt beneath the northern part of the Ebro basin.

[20] Figure 9b shows the interpretation of the ECORS and ESCI profiles, modified after the *Roure and Choukroune* [1998] sketch south of the axial zone. Slopes, depths, and location of features have been spatially restored by using the seismic migration and modeling results by ray tracing of *Matheron and Bareyt* [1998]. The series of reflectors dipping 40° northward and located in the southernmost portion of the ECORS profile are assumed to be remnants of the subducted neo-Tethys slab, the Tethys suture being located underneath the northern part of the Ebro basin. To the north, the lower part of the continental subducting crust dips toward and beneath the axial zone.

[21] Numerous parameters allowing estimates of shortening are not fully constrained (geometry of ramps beneath the Nogueras units, volume of eroded material in the axial zone, geologic significance of reflectors beneath the north Pyrenean fault, and amount of shortening on north Pyrenean thrusts [*Beaumont et al.*, 2000; *Roure and Choukroune*, 1998]). Though erosion probably has been underestimated, shortening values across the different Pyrenean units suggest 150 ± 50 km of convergence [*Roure and Choukroune*, 1998]. Geodynamical models using the restored crustal structure along the ECORS profile give some indication of the role of inherited crustal structures during orogenesis [*Beaumont et al.*, 2000]. The main contribution of the *Beaumont et al.* [2000] model is that the reactivation of a midcrustal weak detachment and the subduction of the lower crust and upper mantle govern the early contractional style of the orogenesis. From reasonable mass balance estimates, *Beaumont et al.* [2000] suggest 165 km of convergence. Constraints derived from geodynamical models and the geological interpretation of the ECORS profile show that the amount of shortening due to the subduction of the northern slab (150 or 165 km) is slightly larger than our plate kinematic estimate (130 km of convergence along the ECORS profile since chron A33o, 80 Ma). A few tens of kilometers may have already subducted along the northern slab at the end of the M0–A33o kinematic episode, since the beginning of the Pyrenean orogeny in late Santonian (85 Ma [e.g., *Garrido and Rios*, 1972]).

3.2. Tomographic Sections Across the Pyrenees

[22] A global model of seismic travel time tomography has been established [*Bijwaard et al.*, 1998] by using a spatial gridding of the Earth's mantle and lithosphere with irregular cells whose size depends of the amount of rays cutting through the cells. In the Pyrenean area, lateral heterogeneities at scales of 0.5° are resolved in the upper mantle allowing us to map subducting slabs (Figure 10). The data used here are a reprocessed version of the International Seismological Center and U.S. Geological Survey's National Earthquake Information Center data sets. Spike tests were performed to assess the resolution of

Figure 8. (a) Reconstruction of continents at chron M0, where EU is supposed to be fixed (parameters in caption of Figure 6a). Gray indicates M25–M0 (late Jurassic-Barremian) amount of extension calculated from parameters of *Srivastava and Verhoef* [1992] for M25 and *Srivastava et al.* [2000] for M0. Large transparent arrows show the M25–M0 directions of plate motions. Hatchures underline areas where extension occurred during that period (rifting on IB, NA, and EU continental margins and shelves). Small black arrows indicate rotation of Galicia Bank (GB) and Flemish Cap (FC) during this period. BCb, Basque-Cantabrian basins; Cb, Catalan basins; Ib, Iberian basins; Mb, Mauléon basin; NPF, North Pyrenean fault; Ob, Organyà basin; Pb, Parentis basin. (b) M0–A33o IB and NA plate motions with respect to EU on A33o plate kinematic reconstruction (parameters A33o IB/EU 39.42°N, 10.91°W, −7.6°; A33o NA/EU 65.53°N, 148.83°E, 18.90° [*Sibuet et al.*, 2004]). Also shown are the positions of the M0–A33o poles as calculated from the M0 pole of rotation of *Srivastava et al.* [2000] and of the IB/EU stage poles of *Sibuet et al.* [2004] (large white circle with cross inside, 44.35°N, 4.30°W, −37.21°), *Srivastava et al.* [1990a] (44.59°N, 4.08°W, −35.93°), *Sibuet and Collette* [1991] (44.49°N, 4.67°W, −37.70°), and *Olivet* [1996] (44.39°N, 4.11°W, −36.76°) (small white circles with crosses inside labeled as 1, 2, and 3, respectively). Light gray indicates amount of extension calculated from the above parameters, and dark gray indicates amount of convergence in the Pyrenean domain. Large arrows show extensional (transparent) and convergent (black) M0–A33o directions of plate motions. Hatchures underline areas where late Aptian-Albian thermal subsidence (this paper) or continental extension [*Martín-Chivelet et al.*, 2002] still occurred in southern France and northern Spain. We suggest that the north Pyrenean basins formed as back arc basins (NPbb, North Pyrenean back arc basins, horizontal dashed lines) linked to the subduction of the neo-Tethys Ocean. (c) IB/EU plate motion since chron A33o (39.42°N, 10.91°W, −7.6°) on the present-day map. Dark gray indicates resulting amount of convergence. Black arrows show the direction of IB/EU convergence. (d) Location of Pyrenean compressive deformation since chron A33o on the present-day map (bathymetric contours at 0.2, 2, and 4 km). Dark gray indicates areas of intense deformation, and light gray indicates areas of minor deformation [*Thinon*, 2001]. Cant, Cantabria seamount; CF, Cape Finisterre; GS, Gascony seamount. Note that the Iberian Ranges formed during Miocene [*Alonso-Zarza et al.*, 2002; *Muñoz and Casas*, 1997], mostly after the Pyrenean orogeny ceased.

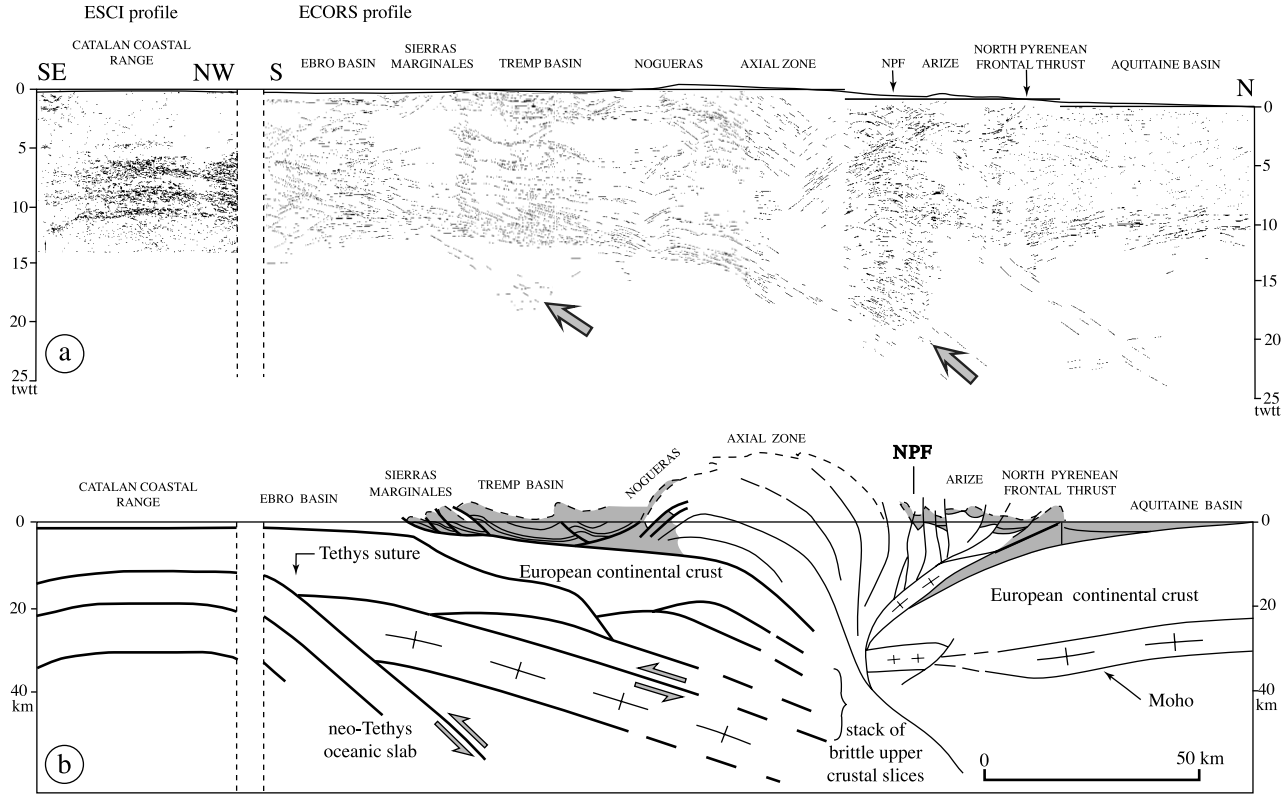

Figure 9. (a) Line drawing of the unmigrated portion of the ECORS profile shot across the Pyrenees from the Aquitaine basin to and the Ebro basin [*Choukroune and ECORS Team*, 1989] and the ESCI profile up to the Mediterranean coast [*Vidal et al.*, 1998]. (b) Interpretation (on depth section) without vertical exaggeration based on Figure 9a. The interpretation of *Roure and Choukroune* [1998] for the northern portion of the ECORS profile is adopted here, while our interpretation only concerns the southern portion of the profile.

tomographic inversions in the studied area. The first 30 km are not resolved because of station corrections. We have made six 1000-km-long N-S cross sections of the Pyrenees located in Figures 1 and 8 down to 700 km into the mantle. Velocity contrasts are ±1.5% for all cross sections. In the Pyrenees and southern France, high-velocity bodies are observed down to a maximum depth of 350 km. Except for the two extreme profiles located across the western Mediterranean Sea and on the eastern border of the Bay of Biscay, two elongated adjacent high-velocity bodies are observed on profile d and generally on the adjacent profiles. They all dip northward. The images of these bodies are similar to those of oceanic slabs evidenced everywhere around the world [e.g., *Bijwaard et al.*, 1998].

[23] Contours of high-velocity bodies underlined in Figure 10 have been plotted on the same graph with all individual cross sections aligned on the NPF in Figure 11. On Figure 11 the simplified cross section of the ECORS profile has been added. Interestingly, there is a good correspondence between the northern and southern décollement zones identified on the ECORS profile and the upper limits of the northern and southern high-velocity bodies. In addition, geometries of the tops of high-velocity bodies lie in the prolongation of the two décollement zones evidenced on the ECORS profile. Thus the two sets of seismic and tomographic data are in agreement, suggesting the existence of two distinct slabs dipping northward.

[24] Though the contours of the two slabs are poorly constrained in their deepest parts (Figure 11), the northern slab seems to dip about 30° northward and is about 100 km thick, and its length is at least 250 km, perhaps 450 km as suggested in Figure 10d. The southern slab dips at least 60° northward and is about 100 km thick, and its length is at least 200 km, perhaps up to 400 km. It seems to be detached and/or overturned in its deepest portion. Among the possible explanations of such a geometry: (1) If the absolute motion of the Iberian plate had a northward component, the overturning effect could be due to the anchoring of the lower portion of the slab in the upper mantle, and (2) if the southern slab was inactive since middle Cretaceous, it was sheared off and sunken into the mantle.

[25] Using local and teleseismic data, a detailed tomographic model with limited geographical extent was previously established beneath the Pyrenees, down to a depth of 200 km [*Souriau and Granet*, 1995]. Their cross section established along the ECORS profile only extends 70 km and 130 km north and south of the NPF, respectively. *Souriau and Granet* [1995] identified a vertical, low-velocity body located beneath the axial zone and down to 100 km, which is not evidenced in our tomographic

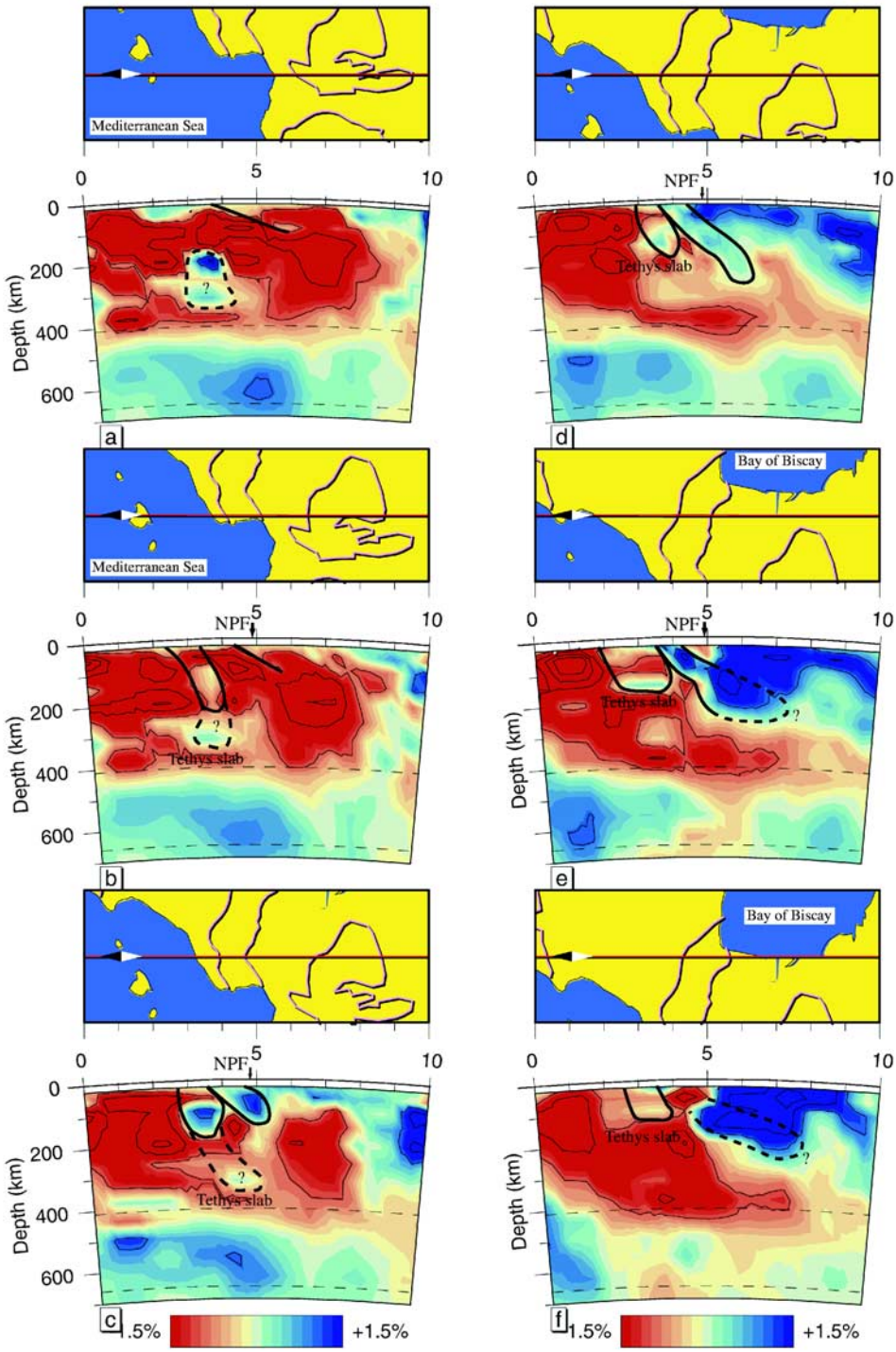

Figure 10. N–S Tomographic sections showing *P* wave velocity anomalies under the Pyrenees. Locations of these sections are shown both in Figure 1 and above each cross section. Velocity anomalies are displayed in percentage (±1.5%) with respect to the average *P* wave velocity at depth. Blue denotes high-velocity zones, and red denotes low-velocity zones. NPF, north Pyrenean fault.

data (Figures 8c and 8d). However, the other features outlined in their model are coincident with those of Figure 10. In particular, the upper limits of their high-velocity zones, though only defined in a small zone, correspond to the roofs of the two slabs already mapped (Figure 11).

[26] On the ECORS profile and tomographic profile d, we have located the Tethys suture beneath the northern part of the Ebro basin where the neo-Tethys slab becomes horizontal (Figure 9b). Such a vertical contact between two different continental crusts is not observed on the ECORS and ESCI seismic profiles, perhaps because vertical features are

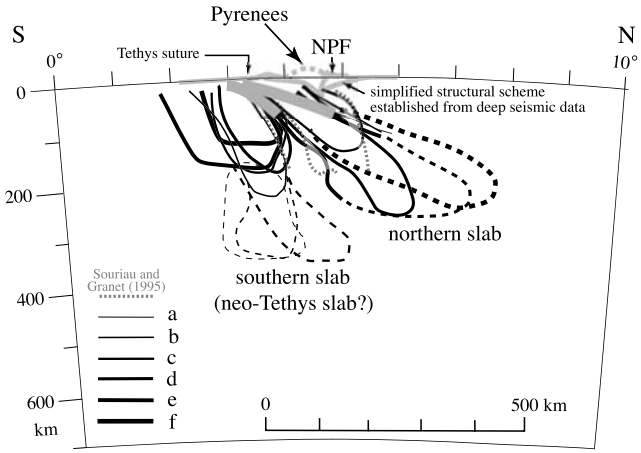

Figure 11. Tomographic sections of Figure 10 aligned on the north Pyrenean fault (NPF). Gray indicates simplified interpretation of the ECORS profile. Note the good correspondence between the positions of slabs identified on seismic and tomographic sections (including *Souriau and Granet*'s [1995] local tomographic model). Dashed lines indicate the ambiguous deep extent of slabs.

never imaged by seismic reflection data. The location of the Tethys suture can be laterally extended on tomographic profiles b to f (Figures 1 and 8). In addition, seismic refraction profiles 6 and 8 (located in Figure 1) provide further evidence on the existence of a slab [*Pedreira et al.*, 2003] whose position and steepness are similar to the neo-Tethys slab imaged in Figure 9b. On the Bouguer anomaly map of the Pyrenees [*Banda and Daignières*, 1995] the proposed location of the Tethys suture coincides with a continuous gravity gradient, which extends farther northwestward on the north Iberian continental shelf. We assume that the trend of the gravity gradient corresponds to the location of the Tethys suture (Figure 1). However, the Tethys suture being defined by low-resolution data, the precision in its location is only several tens of kilometers.

4. Presentation of a Model of Formation of the Pyrenees

[27] The ECORS seismic profile and the tomographic sections bring two independent, complementary constraints concerning the deep structure and suggest the subduction of a former ocean, before the Pyrenan uplift. Such a subduction could have triggered the formation of the north Pyrenean Aptian-Albian basins as back arc basins instead of pull-apart basins formed in an IB/EU transcurrent system. Another relevant point for this study is that plate reconstructions are constrained by the identification of recognizable magnetic anomalies (as chron M0, for example), whose occurrences do not necessarily correspond to tectonic changes. Reciprocally, boundaries between tectonic phases, as the late Santonian (85 Ma) inversion of early Cretaceous basins, which occurred during the Magnetic Quiet Zone, do not correspond to a given plate reconstruction. This point is illustrated in Figure 7 where time intervals characterized by constant angular velocities of the IB/EU declination data might correspond to major tectonic phases in the Pyrenean domain. Finally, both the IB/EU convergent or transform kinematic motions derived from hypothesis 1 or 2 fail to explain the Aptian-Albian rifting in the Iberian basins located in northern Spain [e.g., *Martín-Chivelet et al.*, 2002]. However, we think that it is extremely difficult to separate a rifting from a postrifting phase, even if tectonic subsidence curves are available across a basin. It is a new open emerging research field, as mentioned in section 2.1.

[28] From chrons M25 to M0 (late Jurassic to early Aptian), 350 km of N-S extension occurred between the IB and EU plates (Figure 8a). This extension corresponds to the formation of the northern and southern Bay of Biscay continental margins but also to the formation of both the deep Armorican basin transitional crust and northern M3–M0 Bay of Biscay oceanic crust. We do not exclude that the missing Iberian counterpart of both the Armorican basin transitional crust and M3–M0 Bay of Biscay oceanic crust could have subducted beneath Iberia. Between IB and EU, the extension probably occurred within a 100- to 200-km-wide preexisting Tethys Ocean and gave rise to the neo-Tethys Ocean (Figure 8a). During the late Jurassic to early Aptian period, the extension did not occur only within the former Tethys Ocean but also on the adjacent plates: in the Paris basin, in the Celtic Sea and English Channel [*Ziegler*, 1988], south of the Ebro Variscan massif [*Vergés and García-Senz*, 2001] in the Iberia, Catalan, and Basque-Cantabrian basins [*Casas*, 1993; *Salas et al.*, 2001], in the Organyà basin (southern Pyrenees) [*Beràstegui et al.*, 1990; *Vergés and García-Senz*, 2001], and north of the Pyrenees, in the Parentis, Mauléon, and Arzacq basins [*Bois et al.*, 1997; *Masse*, 1997; *Winnock*, 1971] (Figure 8a).

[29] From chrons M0 to A33o (early Aptian to Campanian), oceanic crust formed in the Bay of Biscay and North Atlantic, but the convergence between the IB and EU plates was consumed in the Pyrenean domain sensu lato (Figure 8b). In our hypothesis, during the Aptian-Albian times, the north Pyrenean basins formed as back arc basins corresponding to the subduction of the southern slab (Tethys and neo-Tethys) beneath EU. Simultaneously, we saw that extension continued in the basins of northern Spain and southern France. Normal faulting in these basins might be due to the differential thermal subsidence occurring within these basins. From late Albian to Santonian, all the basins were characterized by a relative tectonic quiescence [*Martín-Chivelet et al.*, 2002] possibly associated with the decrease of thermal subsidence or to a change in the kinematic parameters (Figure 7). Late Albian–early Cenomanian alkaline magmatic activity and metamorphism are recorded along the north Pyrenean fault zone [*Albarède and Michard-Vitrac*, 1978; *Henry et al.*, 1998; *Montigny et al.*, 1986]. Mantle-derived lherzolitic bodies were emplaced in the floor of the central and western north Pyrenean Albian-Cenomanian basins [*Fabriès et al.*, 1998]. Crustal thinning in the north Pyrenean back arc basins might explain the tectonic emplacement of such ultramafic bodies into the crust, especially along crustal features as the NPF. Since late Santonian (85 Ma), crustal shortening occurred in the Pyrenean area and corresponds to the IB continental subduction beneath EU. In fact, if the first signs of compression in the Pyrenees are dated 106 Ma [*Souquet and Peybernès*, 1991] and are probably linked to some local

closure of the neo-Tethys Ocean, we suggest that the entire closure of the neo-Tethys Ocean was completed 85 Myr ago, when the contraction was generalized along the whole Pyrenean chain.

[30] The major kinematic change at chron A33o (80 Ma, Campanian) corresponds to the end of the Bay of Biscay opening and to a rearrangement of plate boundaries. It implies a global change in the direction of convergence within the Pyrenees from NE-SW to NNW-SSE (Figure 8c). Such a change in the direction of compressive motions at 80 Ma is not observed in the Pyrenees. From kinematic data the post-80 Ma amount of convergence along the ECORS profile is 130 km, less than the 150 km [*Roure et al.*, 1989] or 165 km [*Beaumont et al.*, 2000] of convergence determined along the ECORS profile from balanced cross sections and subsidence data. From a kinematic point of view these observations suggest that the beginning of the orogenic Pyrenean phase associated with the subduction of the northern slab might have started earlier than 80 Ma, like 85 Ma, as suggested in the preceding sections. To summarize, we propose the following scenario for the geodynamic evolution of the Pyrenees (Figure 12).

4.1. Late Jurassic–Early Aptian (156.5–118 Ma): Extension Within the Former Tethys Ocean and Pyrenean Domain (Figure 12a)

[31] From Late Jurassic (M25, 156.5 Ma) to early Aptian (M0, 118 Ma), N-S extension (Figure 8b) occurred within the already existing narrow Tethys Ocean, in the north Pyrenean zone (NPZ) [*Souquet*, 1988] reactivating there older NW-SE faults, in the south Pyrenean zone (Organyà basin), in southern France (Parentis, Mauléon and Arzacq basins), and in northern Spain (Iberia, Catalan, and Basque-Cantabrian basins).

4.2. Early Aptian–Early Albian (118–100 Ma): Neo-Tethys Subduction and Back Arc Basin Extension (Figure 12b)

[32] From 118 to 80 Ma, kinematic data suggest a NE-SW subduction of the neo-Tethys Ocean. The roof of the neo-Tethys slab is the décollement surface observed in the southern part of the ECORS profile. Early Aptian to Cenomanian (120–90 Ma) back arc basins developed about 150–200 km north of the trench (Figure 8b), on the EU plate, at the emplacement of the present-day northern Pyrenees. NW-SE trending features were used as normal faults during the formation of back arc basins. Thus NW-SE trending back arc basins were formed in the Pyrenean domain as narrow basins roughly aligned in the direction of the future Pyrenean chain. Beneath back arc basins, brittle tensional deformation occurred within the upper EU crust simultaneously with lower crustal thinning. At the end of back arc basin formation, ultramafic rocks were emplaced through the NPF crustal feature. Later, compressive motions will preferentially affect such weak zones. During this period, thermal subsidence occurred in the other basins located in southern France and northern Spain.

4.3. Late Santonian (85 Ma): Beginning of Continental Subduction and Back Arc Basin Inversion (Figure 12c)

[33] There is no reason to expect that the two margins of the neo-Tethys Ocean were simultaneously colliding along their whole length. Thus the remainder of the neo-Tethys Ocean progressively disappeared until its complete closure. We suggest that the final suture between the two continental margins was completed 85 Myr ago. On passive continental margins subject to subsequent compression, normal faults located in the ocean-continent transition zone are commonly reactivated as reverse faults [e.g., *Whitmarsh et al.*, 1993]. By comparison, we suspect that back arc basins normal faults were reactivated as reverse faults simultaneously with the occurrence of the underlying decoupling at the lower/upper crust boundary. The NPF would have perhaps initially acted as the southern limit of the chain (S. Brusset and P. Souquet, personal communication, 2003). At 85 Ma (late Santonian) the Pyrenean chain was in an embryonic stage. With the northern shift of the subduction the portion of EU crust located between the Pyrenees and the Ebro basin belongs now to the IB plate.

4.4. Late Santonian–Early Miocene (85–25 Ma): Continental Subduction and Wedging (Figure 12d)

[34] Late Santonian (85 Ma) might correspond to some kinematic change in the Bay of Biscay as suggested by the directions of magnetic lineations, which change between M0 and A34 lineations (Figure 3). Since 85 Ma, after the total closure of the neo-Tethys Ocean, compression inverted previous Pyrenean back arc basins. The opening of the Bay of Biscay ended at 80 Ma. Though compression in the Pyrenees seems to continue along the same lines, there is a major change in the kinematics of Iberia at this date with a shift of the IB/EU pole of rotation from the S-E corner of the Bay of Biscay to west of Portugal and a change in stress direction from SW-NE to SSE-NNW in the Pyrenees (Figure 8). The maximum uplift was observed during middle Eocene–middle Oligocene [e.g., *Fitzgerald et al.*, 1999; *Mattauer*, 1968; *Roure and Choukroune*, 1998]. In the Iberian and Catalan ranges, compression occurred later from middle Eocene to late Miocene [*Muñoz and Casas*, 1997] with a peak of deformation in late Oligocene [*Casas et al.*, 2000].

4.5. Geological Constraints

[35] The presented model shows that the Tethys suture is located beneath the northern part of the Ebro basin and might correspond to a contact between two continental crusts. However, no significant topographic feature affects the basement of the Ebro basin, even though seismic profiles and industry wells are scarce [*Jurado*, 1988]. From a geological point of view, one possibility would be to move 25 km northward the proposed Tethys suture, beneath the southern part of the south Pyrenean thrust system where a vertical offset appears in the basement [*Vergés et al.*, 1995]. On the basis of new data presented here, we favor the double slab model, even if we are concerned with geological data, which do not display clear ground truth evidences [e.g., *Vergés et al.*, 1995]. However, we think that there is no way to pass round the kinematic constraints established in this study. If the existence of the southern slab based on the ECORS and tomographic data is rejected on the basis of geological proofs, the only alternative is a single subduction starting at chron M0 time with the subduction of the neo-Tethys

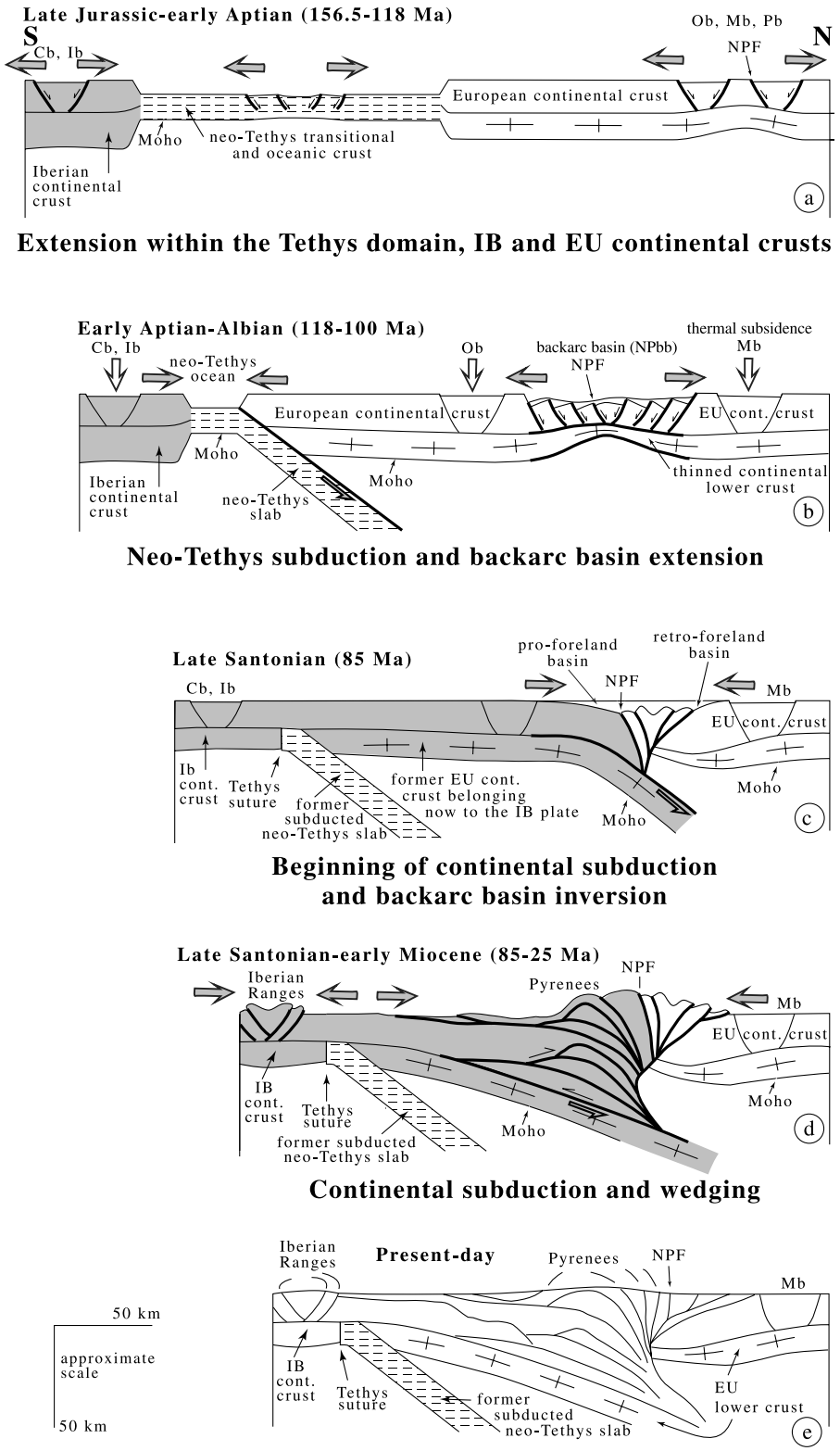

Figure 12. Schematic diagrams showing the evolution of the Pyrenees, incorporating the interpretation of the ECORS and tomographic profiles, combined with constraints from the kinematic reconstructions of the Bay of Biscay and northeast Atlantic. Gray, IB crust; white, EU crust. Thick lines, active faults. Abbreviations are as in Figure 8.

followed 85 Ma by the subduction of the adjacent IB continental crust.

5. Conclusions

[36] The main conclusions of this study are as follows:

[37] 1. The estimate of the M25–M0 (156.5–118 Ma) IB/EU kinematic motion of 350 km of N-S extension gave rise to the rifting of continental margins of the Bay of Biscay and the formation of transitional and oceanic crusts. About 350 km of N-S extension is also distributed within the preexisting narrow Tethys Ocean (neo-Tethys Ocean) located south of the present-day Pyrenees and within the numerous lower Cretaceous adjacent continental basins on the EU and IB plates.

[38] 2. The M0 IB/EU pole of rotation is constrained by the magnetic anomalies identified in the Bay of Biscay and the North Atlantic, by the geometry of a triple junction which had remained at the mouth of the Bay of Biscay during this period, and by paleomagnetic declination data of the stable Iberia. This results not only in a well-constrained position of IB relative to EU at chron M0 but also give us more accurate estimates for the relative motions which took place across their plate boundaries and thus for the formation of the Pyrenees. The analysis shows that simultaneous seafloor spreading occurred in the Bay of Biscay and the North Atlantic at least from chrons M0 to A33o, creating a triple junction west of the Bay of Biscay, which remained there until spreading stopped in the bay at chron A33o. This simultaneous spreading may have even started as early as chron M3 (124 Ma). The absence of this anomaly north of IB, probably as a result of later subduction, makes this assertion only tenuous. The entire oceanic opening of the Bay of Biscay can be explained by an IB/EU rotation around a pole located in the S-E corner of the Bay of Biscay, which resulted in simultaneous convergence in the Pyrenean area.

[39] 3. The deep seismic ECORS and ESCI profiles display a series of north dipping reflectors located about 100 km south of the Pyrenees, visible down to a depth of 18 s twtt. On N-S tomographic sections of the Pyrenees established from teleseismic data, the southern dipping reflectors correspond to the top of a 200- to 400-km-long high-velocity body dipping to the north, in the direction of Pyrenees. We interpret this feature as the roof of a southern slab corresponding to the subduction of the neo-Tethys Ocean. During such a subduction process, elongated back arc basins opened within the Pyrenean area. The complete suture of the neo-Tethys Ocean occurred around 85 Ma.

[40] 4. The roof of the northern 250- to 450-km-long slab identified both on the ECORS profile and on tomographic sections corresponds to a décollement zone located at midcrustal level. This subduction process started at 85 Ma with the inversion of back arc basins. Normal faulting due to differential thermal subsidence across basins still occurred in southern France and northern Spain. Since that time, the Pyrenees fully developed as a double thrust wedge with a peak of deformation during Eocene-Oligocene.

[41] 5. Though still very general, the presented model of the formation of Pyrenees tries to reconcile the geology of Pyrenees and Iberian ranges, tomographic data, and the kinematics of the Bay of Biscay and North Atlantic Ocean. However, if the existence of the southern slab based on the ECORS and tomographic data and of the Tethys suture is disputed, the other alternative is to favor a single subduction of the neo-Tethys Ocean followed by the subduction of the adjacent continental IB plate, simultaneously with the surrection of Pyrenees.

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