Role of hyperextension for the formation of rift systems and its implication for reactivation processes and orogen formation: the example of the Bay of Biscay and Pyrenees

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EXTENDED ABSTRACT

Most of the questions related to plate tectonic theory are related to lithosphere extensional mechanisms leading to continental break up, oceanic crust formation and eventually to the creation of a new plate boundary.

Since the first conceptual models attempting to explain the processes of lithosphere extension have been proposed during the 60ies and 70ies, studies in present-day magma-poor rifted margins and in their fossil analogues preserved in mountain belts unravelled geometries more complex than previously assumed. At first, most studies were focused in proximal margins and intra-continental analogues unravelling tilted block geometries limited by short offset normal faults in areas of weak crustal and lithospheric thinning. In contrast, results from Ocean-Continent-Transition (OCT) highlighted the role of detachment faulting when the crust is already thinned to about 10km thick and for mantle exhumation during the final stage of rifting. The main crustal thinning from an originally equilibrated crust (~30 km) to a thin crust less than 10 km thick is observed along “necking domains” separating the proximal and distal domains. Recent acquisitions of reflection and refraction seismic data in passive margins (Iberian, Norwegian, South Atlantic) first imaged this transition. These observations were then integrated into conceptual and numerical models before necking domain could be identified as fossil remnants preserved in mountain belts (ex: Alps, Pyrenees).

More and more domains of extreme lithosphere thinning are revealed in present-day and fossil analogues of rifted margins. However, the spatial and temporal evolution of these mechanism remains poorly constrained. The aim of this PhD is two-fold. A first objective is to investigate the 3D architecture of “hyperextended” rift systems to characterize the spatial and temporal evolution of mechanisms leading to the extreme crustal and lithospheric thinning. Furthermore, since both present day and fossil analogues preserved in mountain belts can be observed, a second aim is to understand how far rift-related architecture may be considered as inheritance and may control collisional orogen formation.

The Bay of Biscay and Pyrenees (at the boundary between France and Spain) are used as a natural laboratory. This region corresponds to a former rift domain initiated during the Triassic and leading to the formation of hyperextended domains during Early Cretaceous (ex: the Parentis basin). The ultimate stage of continental lithosphere break-up is characterized by the formation of an oceanic domain. Compressional processes initiated in the Late Cretaceous lead to the progressive reactivation and exposure of some of these domains into the Pyrenees (Arzacq-Mauléon, Basque-Cantabrian, Aulus basins). The West to East increasing deformation provides the access to different stages of reactivation.

A new approach to combine offshore-onshore observations

The first part of this PhD thesis presents a new offshore-onshore approach that combine observations derived from marine geophysics and field geology. At present, most studies aiming to characterize rifted continental margin structures were usually either focused offshore relying on indirect geophysical approaches or onshore in deformed analogues offering direct access to geological observations. Marine and onshore examples provide complementary dataset, but the different scale and resolution of observations coming from regions with distinct geodynamic evolutions prevent direct correlations to be done.

The valuable datasets of the Bay of Biscay (seismic, drill-hole data) and the presence of fossil remnants of the same system in the Pyrenees enables the correlation between offshore and onshore observations to define similar structural domains in present-day and fossil rifted continental margins. Gravity inversion and flexural backstripping techniques are combined with seismic interpretations of offshore examples (Western Approach margin and Parentis basin) to estimate accommodation space, crustal and lithosphere thinning and to identify two extensional settings (depending on their implication into thinning processes). Field observations (Mauléon basin) and drill-hole data (Parentis basin) focused on key outcrops enables the description of the nature of sediment and basement and of the structures forming fossil remnants of rifted margins. This qualitative and quantitative analysis of offshore and onshore observations provides the necessary diagnostic elements to determine 5 distinct domains in present-day and fossil rifted continental margins: the proximal, necking, hyperthinned, exhumed mantle and oceanic domains.

The development of this offshore-onshore approach represents the essential requirement to map rift-related paleogeographic domains inherited from the Bay of Biscay opening and partially integrated into the Pyrenean orogen.

Spatial and temporal evolution of hyperextended rift systems

A second important step to propose a coherent paleogeographic restoration of the system is to decipher the compressional history. Results from this PhD thesis suggest that the observed heterogeneous reactivation is directly related to its complex pre-compressional architecture.

The restoration of the domain enables the distinction between two extensional rift systems preserved at different evolutionary stages. They are separated by the “Landes block” representing a continental ribbon of weakly thinned continental crust. The Bay of Biscay spatially records different phases of lithospheric thinning processes ranging from the hyperthinned Parentis basin in its eastern termination to oceanic crust at the junction with the Atlantic Ocean in the Western Bay of Biscay. A second system corresponds to a hyperextended domain with local mantle exhumation (the present-day Arzacq-Mauléon, Basque-Cantabrian basins…) without forming an oceanic domain. Both systems are strongly segmented during hyperextensional processes. Segmentation is inherited from the Variscan to Permo-Triassic structuration preceding hyperextension. The reactivation of some of these crustal and/or lithospheric structures delimits variations in the margin architecture. Major changes are related to switches in the vergence of exhumation systems during the final stage of rifting. The evolution of thinning and hyperextensional processes is slightly diachronous between the two systems. The estimated age of the first oceanic crust and the study of the Parentis basin suggest that crustal and lithospheric thinning occur between the Jurassic and Aptian-Albian time. In contrast, processes of extreme lithospheric thinning continues until Albian to Early Cenomanian in the system preserved in the Pyrenees as indicated by reworking of mantle rocks in Albian to Cenomanian breccias (Mauléon and Aulus basins).

The restoration of the system at the end of hyperextension, prior to convergence underlines the system segmentation (NE-SW in the Pyrenean system). This observation prevents any large Albian to Cenomanian strike-slip movement during and after hyperextensional processes along the North Pyrenean fault as suggested in most models.

Role of extreme thinning for the formation of collisional orogens

The complex architecture inherited from hyperextension will result in a competition between the two systems during convergence and leading to the heterogeneous reactivation of the rift system that provides access to different “genetic” steps of compressional processes. As a result, in the West, the initiation of deformation can be observed, whereas in the east (in the Basque-Cantabrian and Arzacq-Mauléon basins) the reactivation of rift structures can be investigated.

The compilation of offshore and onshore observations across the overall system is used to propose an evolutionary model explaining the progressive reactivation. Serpentinization of exhumed mantle basement represents a major rheological weak layer where reactivation processes may initiate. The former top basement detachment fault roofing the exhumed mantle can be used as a decoupling interface. The propagation of the deformation to the hyperthinned crust initiates the formation of an accretionary prism. In the former hyperextended basins (e.g.: Arzacq-Mauléon), the arrival of the conjugate margin enables the initiation of a proto-subduction of the former hyperthinned domain in a later stage. The new interpretations proposed in this study suggest that the subducted crustal wedges imaged by tomographic and seismic methods may partly correspond to the former exhumed mantle and hyperthinned domains.

The results of this part of the PhD thesis suggest that most of the compressional deformation is accommodated by the exhumed mantle and hyperthinned domains. The study also shows that most of the reactivation is localized at the transition between rift domains, suggesting that most compressional structures are inherited from the former rift system. Results of this study further support the necessity to take into account the 3D architecture of former hyperextended rift systems to unravel the formation of collisional orogens.

Consequences for the nature and evolution of the Iberian-European plate boundary

A final part of the study focuses on the formation of the European-Iberian plate boundary and its possible link with the complex 3D architecture of hyperextended rift system. Regional paleogeographic reconstructions require a left-lateral movement of the Iberian plate relative to the European plate. The timing and total amount of the displacement, as well as the location and nature of the plate boundary are, however, controversial. Based on a synthesis of previous and new observations at the scale of the European and Iberian plate boundary, a model of formation and evolution is proposed showing the repartition of the deformation between the different rift systems.

The main results highlight the polyphased evolution of the Iberian-European plate boundary and its diffuse nature during the displacement of the Iberian plate. Instead of a localized deformation along a strike-slip fault system, the deformation seems to be partitioned between two distinct oblique rift systems (the proto-Bay of Biscay and the Iberian range Mesozoic basins) oriented NW-SE. Between these two rift systems, separated by the Landes High and Ebro blocks, the deformation is diffused and accommodated along several E-W trending basins (e.g.: Arzacq, Mirande), The formation of an oceanic domain at Aptian-Albian time succeeding hyperextensional processes localize extensional deformation in the Bay of Biscay after the counter-clock wise rotation of the Iberian plate. The progressive extension of the oceanic domain is associated with an NNE-SSW to NE-SW extension underlined by the segmentation of the basins preserved in the Pyrenean system.

Additionally, the large scale architecture of the boundary between the European and Iberian plates is indeed complex. Onshore, a transition zone is observed characterized by different rift systems separated by weakly thinned continental ribbons (Ebro block and Landes High). The initiation of sea-floor spreading processes in the Bay of Biscay and the diffuse extensional deformation accommodated in the Pyrenean-Basque-Cantabrian rift system in the meantime is interpreted as tentative propagation of a plate boundary that failed to localize.

The polyphased evolution of this region illustrates the importance of pre-breakup movements for kinematic restorations. Furthermore, the oblique rift episode in the Bay of Biscay may partly represent an analogue to understand the processes related to the formation of transform and segmented margins.

In conclusion, the results of this PhD thesis emphasize the role of pre-rift crustal and lithospheric inheritance for the formation of hyperextended systems on the one hand and on the other hand the role of the architecture of these systems for the formation of collisional orogens. Furthermore, the methodology developed may be applied to other systems for the identification, description and interpretation of present-day rifted margins and their fossil remnants preserved in mountain belts.

RÉSUMÉ

La théorie de la tectonique des plaques soulève de nombreuses interrogations sur les mécanismes d'extension de la lithosphère responsables de la rupture continentale, de la création d'un domaine océanique et d'une nouvelle limite de plaque.

Depuis les années 1960-1970 et les premiers modèles conceptuels pour expliquer les processus d'extension lithosphérique, des études menées dans les marges passives peu magmatiques actuelles et fossiles, préservées dans les chaînes de montagne, ont révélé des géométries de plus en plus complexes. Les premiers travaux se sont focalisés dans les marges proximales et leurs analogues intracontinentaux, révélant des géométries de blocs basculés délimités par des failles normales à faible rejet et associés à un amincissement lithosphérique et crustal limité. Au contraire, les résultats des études sur les Transitions-Océan-Continent (TOC) ont mis en avant le rôle de failles de détachement lorsque la croûte, déjà amincie, avoisine les 10 km d'épaisseur et pour l'exhumation de manteau lors de la phase finale du rifting. L'amincissement crustal majeur qui marque le passage d'une croûte de ~30 km à ±10km entre domaines proximaux et distaux s'observe le long des «domaines de necking». L'acquisition de nouvelles données de sismiques réflexion et réfraction dans les marges passives peu volcaniques (Ibérique, Norvégienne et Sud Atlantique) a permis dans un premier temps d'imager cette transition. Ces observations ont ensuite été intégrées dans des modèles conceptuels et numériques avant que des domaines de necking ne soient identifiés en tant que reliques fossiles préservées dans les chaines de montagne (ex : Alpes et Pyrénées).

De plus en plus d'études mettent en évidence des domaines d'amincissement extrême de la lithosphère à la fois dans les marges actuelles et leurs analogues fossiles. Cependant, l'évolution de ces mécanismes dans le temps et dans l'espace reste mal contrainte. Le but de ma thèse est double : d'une part de caractériser l'architecture 3D de systèmes de rift «hyper-amincis» pour évaluer l'évolution spatiale et temporelle des mécanismes associés l'amincissement extrême de la croûte. D'autre part, puisque ces domaines sont observés en mer mais également réactivés et préservés dans des chaînes de montagne, un second objectif est de mieux comprendre en quelle mesure l'architecture issue du rift peut consister un héritage et contrôler la formation des orogènes de collision.

Ce travail de thèse utilise le Golfe de Gascogne et les Pyrénées (à la frontière entre la France et l'Espagne) comme laboratoire naturel. Cette région correspond à un ancien domaine de rift initié au Trias et qui aboutit au Crétacé inférieur à la formation de domaines hyper-amincis (ex : bassin de Parentis). Le stade ultime du processus de rupture de la lithosphère continentale est marqué par la création d'un domaine océanique dans le Golfe de Gascogne. Au cours de la convergence initiée au Crétacé supérieur, certains de ces domaines sont progressivement réactivés et exposés dans les Pyrénées (Bassins d'Arzacq-Mauléon, Basque-Cantabre, Aulus…). La déformation croissante d'Ouest en Est donne l'accès à différentes étapes de réactivation.

Une nouvelle approche pour coupler les observations terre-mer

La première partie de ma thèse est consacré à la présentation d'une nouvelle approche terre-mer couplant des données de géophysique marine et de géologie de terrain provenant d'analogues fossiles préservés dans les chaînes de montagne. La majorité des travaux caractérisant la structure des marges passives continentales se focalise généralement sur l'un ou l'autre type de données. Les exemples à terre et en mer fournissent des informations complémentaires qui sont souvent difficiles à relier entre elles dû aux différences de résolution et d'échelle d'observation et provenant de régions qui ont connu des évolutions géodynamiques différentes.

La richesse des données du Golfe de Gascogne (sismiques, forages) et la présence de reliques fossiles du même système dans les Pyrénées permet de coupler des observations faites à terre et en mer pour définir des domaines structuraux comparables dans les marges passives actuelles et leurs analogues fossiles. L'association d'interprétation de données de sismique réflexion avec des techniques d'inversion gravimétrique et de backstripping flexural sur les exemples en mer (marges des Entrées de la Manche et bassin de Parentis) permet d'estimer l'espace d'accommodation, l'amincissement crustal et lithosphérique ainsi que d'identifier deux types de domaines extensifs (selon leur implication pour les processus d'amincissement). Les données de terrain (bassin de Mauléon) et de forages (bassin de Parentis) focalisées sur des affleurements clefs permettent de décrire la nature des sédiments et du socle ainsi que les structures qui forment les restes fossiles des domaines de marge. Cette analyse qualitative et quantitative d'observations à terre et en mer fournit des éléments de diagnostic essentiels pour identifier 5 domaines distincts dans les marges passives actuelles et fossiles : les domaines proximaux, de necking, hyper-amincis, de manteau exhumé et océaniques.

Le développement de cette approche terre-mer a été un prérequis nécessaire pour cartographier les domaines paléogéographiques issus du système de rift lié à l'ouverture du Golfe de Gascogne et partiellement intégré à l'orogène Pyrénéenne.

Evolution spatiale et temporelle des systèmes de rift hyperamincis

Dans un second temps, déchiffrer la déformation compressive représente une étape essentielle pour accéder à une reconstruction palinspatique cohérente du système. Les résultats issus de cette thèse suggèrent que la réactivation hétérogène observée est directement liée à une architecture pré-compressive complexe.

La restauration du domaine met en avant deux systèmes extensifs distincts préservés à des étapes d'évolution différentes. Ils sont séparés par le bloc des Landes représentant un «ruban» de croûte continentale pas ou peu aminci. Le Golfe de Gascogne enregistre différentes phases d'amincissement spatialement depuis sa terminaison Est dans bassin hyper-aminci de Parentis jusqu'au stade ultime de l'océanisation à l'Ouest à la jonction avec l'océan Atlantique. Un second système correspond à un domaine hyper-aminci avec exhumation locale de manteau (à présent les bassins d'Arzacq-Mauléon, Basque-Cantabre…) sans former de domaine océanique. Les deux systèmes sont fortement segmentés lors des processus d'hyper-extension. La segmentation est contrôlée par la structuration héritée des épisodes varisque à permo-triassique qui précédent l'hyper-extension. La réactivation de certaines de ces structures d'échelle crustale et lithosphérique délimite des variations d'architecture de la marge dont les changements majeurs sont attribués à des changements de vergence de système d'exhumation lors de la phase finale du rifting. L'évolution des processus d'hyper-extension et d'amincissement est légèrement diachrone entre les deux systèmes. L'âge estimé de la première croûte océanique et l'étude du bassin de Parentis indiquent que l'amincissement lithosphérique et crustal se fait entre la fin du Jurassique et l'Apto-Albien. Au contraire, le processus d'amincissement extrême du système préservé dans les Pyrénées se poursuit jusqu'à la fin de Albien et le début du Cénomanien comme attesté par le remaniement de morceaux de manteau dans des brèches Albiennes à Cénomanienne (bassins de Mauléon et d'Aulus).

La restauration du système à la fin de l'hyper-extension, avant l'initiation de la convergence met en avant la segmentation du système (NE-SO dans le système Pyrénéen). Cette observation laisse peu de place pour l'activation d'un large épisode transformant syn à post hyper-extension (albien ou cénomanien), le long de la faille Nord Pyrénéenne, comme proposée dans la majorité des modèles d'évolution du système.

Rôle de l'amincissement extrême pour la formation des orogènes de collision

L'architecture complexe héritée de l'hyper-extension va aboutir à une compétition entre les deux systèmes lors de la convergence et créer une réactivation hétérogène qui permet d'accéder à différentes étapes de la déformation compressive. A l'ouest, on peut observer où s'initie la déformation, tandis qu'à l'est (dans les bassins Basques-Cantabres et Arzacq-Mauléon), il est possible de voir comment les structures de rift sont réactivées.

La compilation d'observations à terre et en mer travers l'ensemble du système permet de proposer un modèle pour expliquer la réactivation progressive. La serpentinisation du manteau va constituer une zone de faiblesse majeure qui peut permettre d'initier la réactivation. L'ancienne faille de détachement au toit du manteau exhumé et serpentinisé peut être utilisée comme surface de décollement. La propagation de la déformation au domaine de croûte hyper-amincie va initier la formation d'un prisme d'accrétion. Dans les anciens bassins hyper-amincis (ex : Arzacq-Mauléon), l'arrivée de la marge conjuguée va permettre dans un stade ultérieur une proto-subduction de l'ancien domaine de croûte hyper-amincie. L'architecture finale de la chaîne s'acquiert lors la collision des anciens domaines conjugués de necking et proximaux qui représentent deux «buttoirs». Les nouvelles interprétations proposées dans ce travail proposent que les domaines «subductés» imagés par la géophysique correspondent en partie aux anciens domaines de manteau exhumés et de croûte hyper-amincie.

Cette partie focalisée sur la réactivation du système montre que l'essentiel de la déformation compressive est accommodée par les domaines de manteau exhumé et de croûte hyper-amincie et les structures de rift associées. Les anciennes limites de domaines constituent les zones majeures de réactivation. Les résultats de ce travail montrent qu'il est nécessaire de prendre en compte l'architecture 3D des anciens domaines rift hyper-amincis dans les modèles de formation des orogènes de collision comme les Pyrénées.

Conséquences pour la nature et l'évolution de la limite de plaque entre Ibérie et Europe

Dans une dernière partie, je me suis intéressée au contexte de formation de la limite de plaque entre Ibérie et Europe et au lien possible avec l'architecture 3D complexe des systèmes de rift hyper-amincis. Les reconstructions paléogéographiques régionales imposent un mouvement senestre de l'Ibérie dont le timing et les implications pour la limite entre Europe et Ibérie sont le sujet de nombreuses controverses. La synthèse de précédentes et nouvelles observations à l'échelle de la limite de plaque Ibérie-Europe m'a permis de proposer un modèle de formation et d'évolution focalisé sur la répartition de la déformation entre les différents systèmes de rift observés.

Les résultats majeurs mettent en avant l'évolution polyphasée de la limite de plaque entre Ibérie et Europe et soulèvent des questions quant à sa nature diffuse lors du mouvement de l'Ibérie. En effet, je propose qu'au lieu d'une déformation localisée le long d'un système de faille, celle-ci soit partitionnée entre deux systèmes de rift obliques distincts (futur Golfe de Gascogne et bassins Intra-Ibériques) d'orientation NW-SE. Entre ces deux rifts, séparés par les blocs des Landes et d'Ebro, la déformation est diffuse entre plusieurs bassins individualisés d'orientation E-W (ex : Arzacq, Mirande). La formation d'un domaine océanique à l'Apto-Albien suivant les processus d'hyper-extension localisent la déformation extensive dans le Golfe de Gascogne, suite à la rotation antihoraire de l'Ibérie. L'extension progressive du domaine océanique est accompagnée d'une extension NNE-SSW à NE-SW dans les bassins du système Pyrénéen, marquée par la segmentation du système.

L'architecture grande échelle de la limite entre Ibérie et Europe est complexe. A terre, une zone de transition s'observe caractérisée par différents systèmes rift séparés par des «rubans» de croûte continentale non amincis (blocs des Landes et d'Ebro). L'initiation du processus d'accrétion océanique dans le Golfe de Gascogne se traduit par une déformation extensive diffuse dans le système de rift Pyrénéen-Basque-Cantabre. Cette distribution de la déformation est interprétée comme une tentative avortée de propagation et de localisation d'une nouvelle limite de plaque.

Ainsi l'évolution polyphasée de cette région illustre d'une part l'importance des mouvements pré-rupture continentale pour les reconstructions cinématiques. D'autre part, l'épisode de rifting oblique du Golfe de Gascogne peut en partie représenter un analogue pour comprendre les processus de formation des marges segmentées ou transformantes.

En conclusion, les résultats de cette thèse mettent en avant le rôle central de l'héritage crustal et lithosphérique pré-rift pour la formation des systèmes hyper-amincis mais également le rôle de l'architecture de ces systèmes pour la formation des orogènes de collision. De plus, la méthodologie développée dans cette thèse peut s'appliquer dans d'autres systèmes pour l'identification, la description et l'interprétation de marges passives actuelles et leurs reliques fossiles préservées dans les chaînes de montagnes.

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…Une page se tourne et un nouveau chapitre commence…

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Au cours d'une thèse, on ne s'attend pas toujours à ce que l'on va découvrir… Inutile que je livre ici ma pensée, tu le sais déjà… A ma moitié…

SOMMAIRE

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INTRODUCTION

Depuis l'introduction de la théorie de la tectonique des plaques à la fin des années 60, les hypothèses et concepts visant à expliquer les mécanismes d'extension de la lithosphère ont beaucoup évolué. Ces avancées ont été possibles notamment grâce aux interactions et analogies entre des observations faites en mer sur les marges actuelles et à terre sur les marges fossiles préservées dans les chaînes de montagnes. La compréhension des processus de rifting et drifting reste un enjeu majeur en sciences de la Terre, à l'interface de nombreuses disciplines telles que la géophysique marine et la géologie de terrain comme cela est proposé dans ce mémoire de thèse.

La suite de ce chapitre décrit succinctement l'évolution des connaissances et concepts sur les marges passives peu magmatiques avant d'aborder les thématiques à l'origine de ce mémoire de thèse.

1. Evolution des connaissances sur les marges passives peu magmatiques : apport des analogues fossiles

Les marges passives peuvent être classées en marges passives magmatiques et marges peu magmatiques (Reston 2009) en fonction de la quantité de magma observée (fig. 1). La distinction entre ces deux types est en réalité plus complexe comme le soulignent Reston & Manatschal (2011). En effet, les apports magmatiques peuvent varier latéralement (e.g. marge Ouest du Groenland, Chalmers & Pulvertaft 2001). De plus, certaines marges subissent une évolution polyphasée : un rifting peu magmatique est suivi d'un processus de rupture lithosphérique fortement magmatique (Vøring : Osmundsen & Ebbing 2008 ; Rockall Trough : Reston 2009). L'inverse est également possible (Inde-Seychelles : Armitage *et al.* 2010). Les marges transformantes représentent une catégorie particulière (fig. 1) et sont associées à des variations abruptes d'épaisseur crustale qui résultent probablement d'une déformation extrêmement localisée (Reston & Manatschal 2011). Elles partagent néanmoins certains traits caractéristiques des marges passives peu magmatiques : de faibles quantités de magma et des domaines de manteau serpentinisé très localisés (Sage *et al.* 2000). In fine, la plupart des marges passives sont hybrides et peuvent combiner des caractéristiques propres à différents types de marges passives. La classification en marges passives magmatiques, marges peu magmatiques et marges transformantes représente bien évidement des pôles extrêmes (Reston 2009 ; Reston & Manatschal 2011). La suite de ce manuscrit est essentiellement focalisée sur les «*marges passives peu magmatiques*».

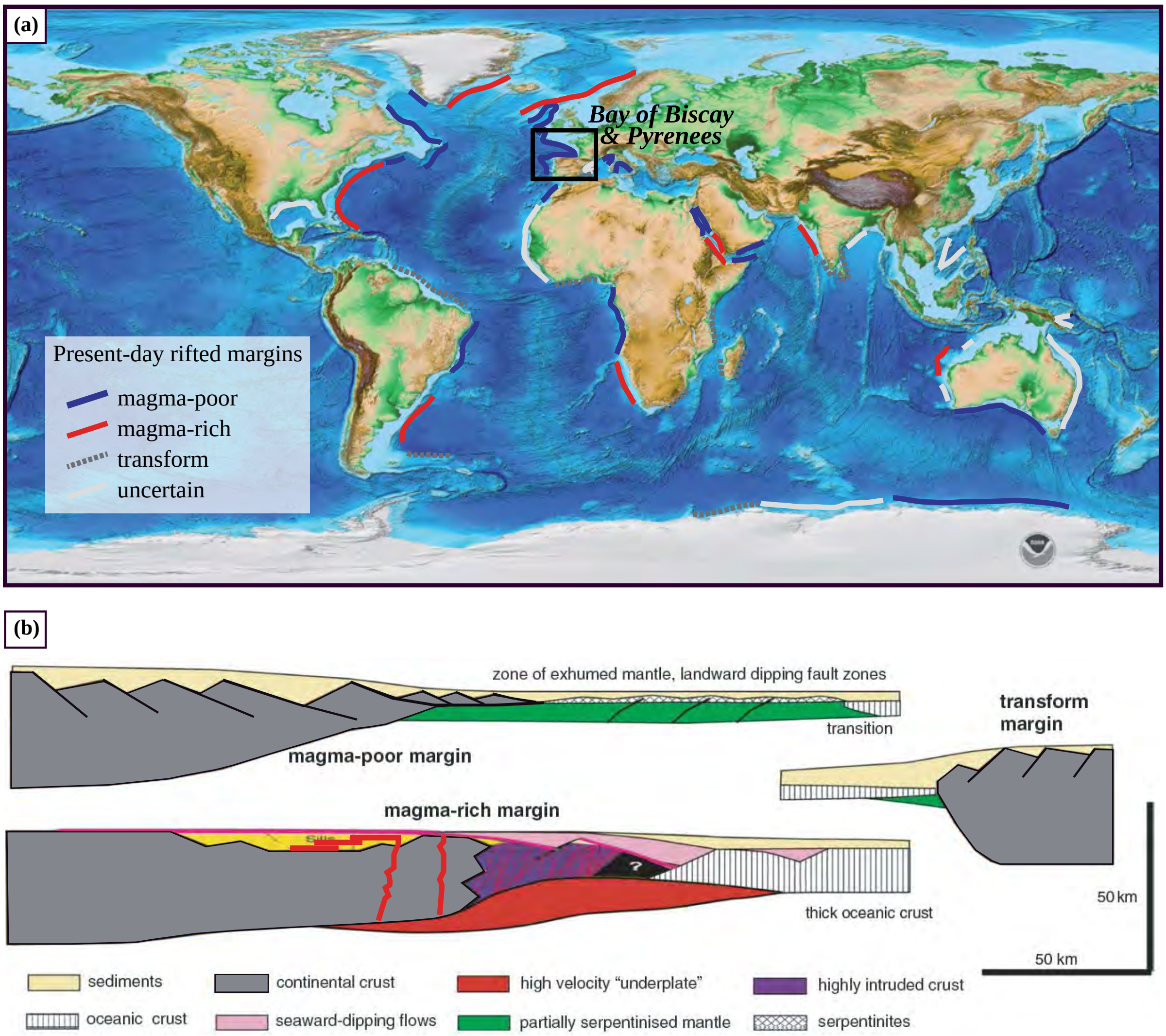
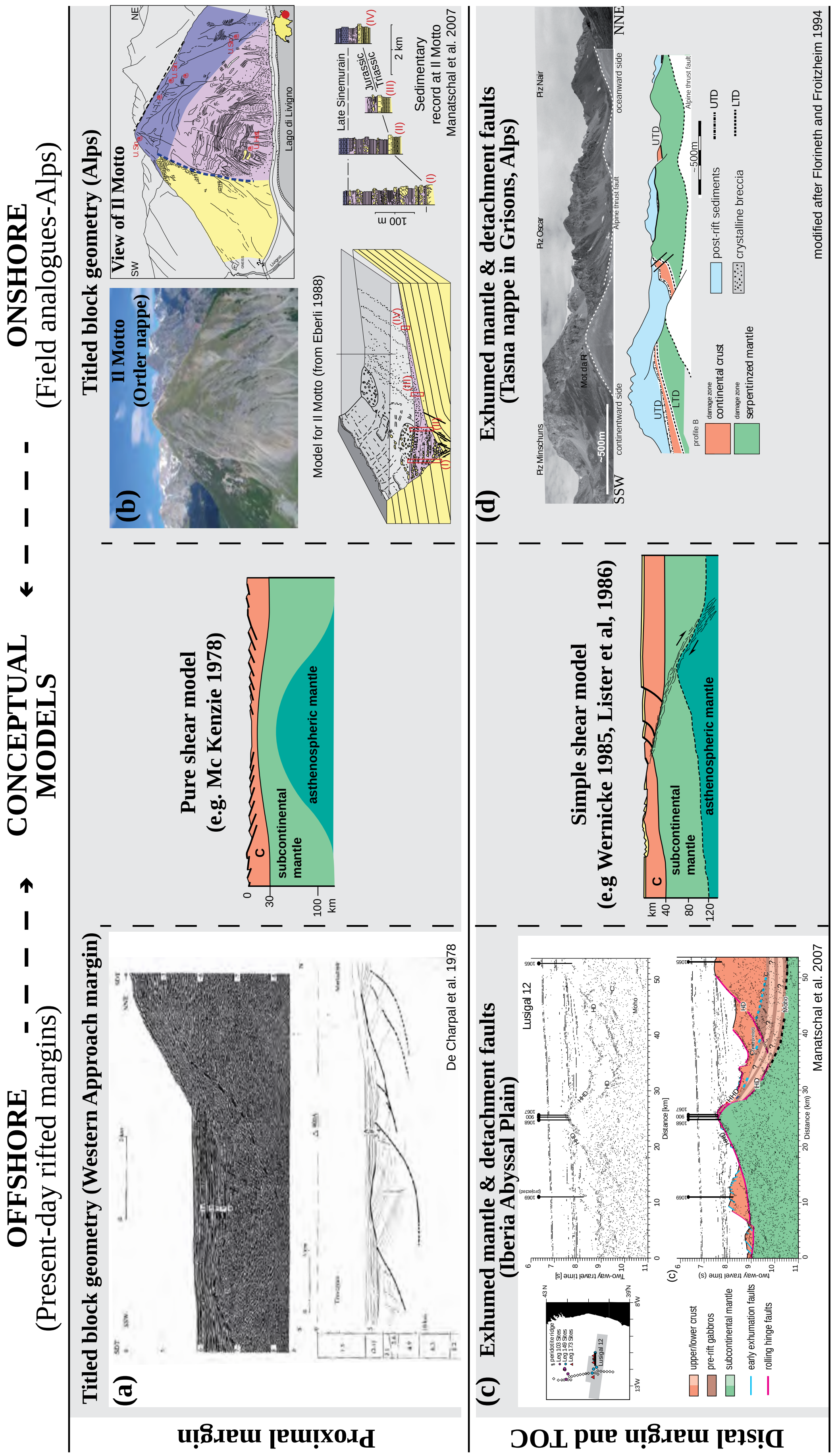


***Fig. 1**: (a) Carte bathymétrique illustrant la répartition des différents types de marges passives actuelles (image modifiée de NOAA, National Geophysical Data Center) et la localisation du chantier Golfe de Gascogne-Pyrénées. (b) Architecture simplifiée des marges passives peu magmatique, magmatique et transformante (modifié de Reston 2009 ; Geoffroy 2005 ; Sage et al. 2000 ; Reston & Manatschal 2011).*

1.1. Premiers modèles, apport des observations des rifts intracontinentaux

Les observations faites à terre dans les chaînes de montagnes ont mis en évidence très tôt l'existence de bassins sédimentaires précédant la formation des orogènes notamment dans le cas du domaine Alpin (e.g. Bertrand 1884 ; Argand 1916 ; Staub 1917 ; Bernoulli 1964 ; Trümpy 1984 ; Lemoine *et al.* 1986). Par suite, les travaux sur les mécanismes associés à la formation de ces bassins ont montré qu'ils étaient contrôlés par des failles normales à fort pendage (e.g. Bernoulli 1964 ; Trümpy 1984 ; fig. 2). Dès les années 1970, la découverte du premier bloc basculé imagé en sismique dans la marge des Entrées de la Manche (De Charpal *et al.* 1978 ; Montadert *et al.* 1979) a permis les premières analogies entre les structures observées à terre et les données acquises en mer, révélant des géométries de blocs basculés contrôlées par des failles normales à faible rejet. Ces observations ont permis l'élaboration d'un premier modèle conceptuel cohérent pour expliquer les mécanismes d'extension de la lithosphère et la formation des bassins sédimentaires associés (Mc Kenzie 1978 ; fig 2). Ce modèle implique un processus d'amincissement constant et uniforme en fonction de la profondeur (βCroûte=βManteau) en contexte de cisaillement pur. Ce modèle détermine une relative symétrie des marges conjuguées (fig. 2) et permet d'expliquer les architectures des bassins observés dans les marges proximales et dans de nombreuses marges passives peu magmatiques (fig. 1).

Une seconde génération de modèle est apparue dans les années 1980, développée à la suite des travaux sur les «metamorphic core complexes» dans la région du «Basin and Range» aux Etats-Unis (Wernicke 1981, 1985). Les processus d'extension sont contrôlés par un détachement à faible pendage et d'échelle lithosphérique caractérisé par un mécanisme de cisaillement simple (fig. 2). L'architecture des marges conjuguées est alors asymétrique.



OFFSHORE
(Present-day rifted margins)
CONCEPTUAL MODELS
ONSHORE
(Field analogues-Alps)
Proximal margin
Distal margin and TOC
(a) Titled block geometry (Western Approach margin)
De Charpal et al. 1978
Pure shear model
(e.g. Mc Kenzie 1978)
subcontinental mantle
asthenospheric mantle
C
(b) Titled block geometry (Alps)
Il Motto
(Ortler nappe)
View of Il Motto
SW
NE
Lago di Livigno
Model for Il Motto (from Eberli 1988)
Late Sinemurian
Jurassic
Triassic
100 m
2 km
Sedimentary record at Il Motto
Manatschal et al. 2007
(c) Exhumed mantle & detachment faults
(Iberia Abyssal Plain)
Lusigal 12
Two-way travel time [s]
Distance [km]
upper/lower crust
pre-rift gabbros
subcontinental mantle
early exhumation faults
rolling hinge faults
Manatschal et al. 2007
Simple shear model
(e.g Wernicke 1985, Lister et al, 1986)
subcontinental mantle
asthenospheric mantle
C
(d) Exhumed mantle & detachment faults
(Tasna nappe in Grisons, Alps)
Piz Minschuns
Mot da l'Ö
Piz Oscar
Piz Nair
SSW
NNE
continentward side
oceanward side
Alpine thrust fault
UTD
LTD
~500m
continental crust
serpentinized mantle
damage zone
post-rift sediments
crystalline breccia
modified after Florineth and Froitzheim 1994

1.2. Les Transitions Océan Continent (TOC) : apport des reliques fossiles alpines

La découverte de manteau sous-continental au large du Banc de Galice dans la marge Ibérique (Boillot *et al.* 1987) a conforté la nécessité de proposer de nouveaux concepts pour expliquer l'extension lithosphérique. Les travaux menés à terre dans les reliques des marges de la Téthys Alpine ont mis en évidence des structures similaires (fig. 2, Froitzheim & Eberli 1990, Manatschal & Nievergelt 1997). L'analogie entre les observations faites à terre et en mer a permis de caractériser l'architecture des Transitions Océan Continent (TOC) et révélé la présence des failles de détachement au toit du manteau sous-continental exhumé et de la croûte continentale dans les parties distales de la marge. Le modèle du détachement d'échelle lithosphérique (Wernicke 1981, 1985 ; Lister *et al.* 1986 ; Lister & Davis 1989) est alors appliqué pour expliquer l'exhumation du manteau dans les TOC de la marge Ibérique (e.g. Manatschal *et al.* 2001) et les reliques fossiles préservées dans les Alpes (Lemoine *et al.* 1987). Cependant, contrairement aux déformations mylonitiques associées à la mise en place des «metamorphic core complexes», les failles de détachement qui caractérisent l'architecture distale de la marge et les TOC sont actives dans le domaine cassant, impliquant des températures inférieures à 300°C (Marge Ibérique : Whitmarsh *et al.* 1996a, 1996b ; Alpes : Manatschal 2004).

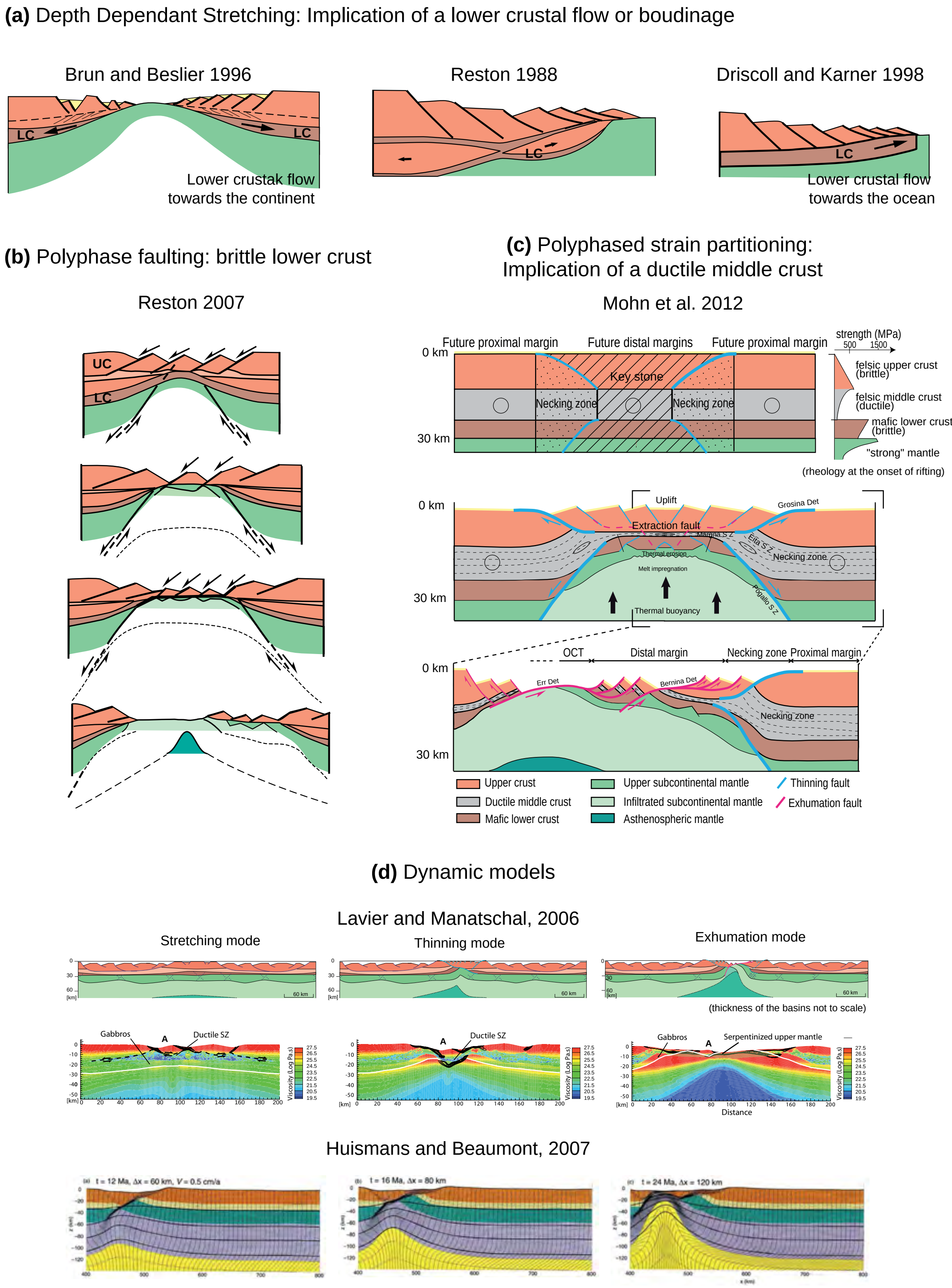
De nombreuses études se sont alors attachées à la caractérisation des TOC et des domaines distaux des marges, et plus précisément sur les processus associés à l'exhumation du manteau (Whitmarsh *et al.* 2001), sur le rôle de la serpentinisation (Pérez-Gussinyé &Reston 2001 ; Reston & Pérez-Gussinyé 2007), sur le magmatisme (Minshull *et al.* 2001 ; Münterer & Manatschal 1996) et aussi sur leurs propriétés géophysiques (Dean *et al.* 2001 ; Whitmarsh *et al.* 2001 ; Minshull 2009). Ces nombreux travaux mettent en évidence que la transition entre croûte continentale et croûte océanique est en fait progressive et non pas abrupte comme suggéré par le modèle proposé par Mc Kenzie (1978).

***Fig. 2**: (a) Premier bloc basculé imagé par sismique réflexion dans la marge des Entrées de la Manche (De Charpal et al. 1978 ; Montadert et al. 1979). (b) Exemple de bloc basculé préservé dans les Alpes: Il Motto et modèle d'évolution des faciès (Eberli 1988 ; Manatschal et al. 2007). Ces géométries de blocs basculés ont inspiré le modèle de Mc Kenzie (1978) partie centrale de la figure. (c) Carte de la marge passive Ibérique et localisation de la coupe sismique Lusigal 12. Coupe sismique Lusigal 12 et interprétation géologique associée de la transition océan continent mettant en avant le jeu de faille de détachement pour l'exhumation de manteau (d'après Manatschal et al. 2007). (d) Panorama de la nappe de Tasna préservant l'architecture et les structures de la transition océan continent dans les Alpes Suisses. Cet affleurement préserve le passage d'une croûte amincie jusqu'au manteau exhumé (figure d'après Florineth & Froitzheim 1994). Ces observations sont intégrées dans les modèles de cisaillement simple proposés par Wernicke (1985) et Lister et al. (1986).*

1.3. Rôle de l'amincissement crustal et modèles polyphasés

Les travaux menés dans les TOC et les parties distales des marges montrent que dans ces domaines, la croûte est déjà amincie à l'extrême et avoisine les 10 km d'épaisseur (Manatschal *et al.* 2001 ; Pérez-Gussinyé &Reston 2001). L'amincissement crustal majeur qui marque le passage entre une croûte de 30 km à 10 km entre les domaines proximaux et distaux est observé le long des domaines de necking (domaines d'étranglement) qui sont bien imagés en sismique réfraction (marge Ibérique, Péron-Pinvidic & Manatschal 2009 ; Norvégienne, Osmundsen & Ebbing 2008 ; Sud Atlantique, Contrucci *et al.* 2004 ; Armoricaine, Thinon *et al.* 2003). Dès lors, la compréhension des processus d'amincissement devient un enjeu majeur. De récentes études mettent en évidence l'importance de l'évolution polyphasée des processus d'amincissement et leur évolution en fonction de la profondeur (Depth-Dependent Streching, DDS, e.g. Kusznir & Karner 2007). Plusieurs modèles sont proposés combinant des processus de cisaillement pur et simple et des mécanismes d'amincissement en fonction de la profondeur (DDS, fig. 3). Certains de ces modèles impliquent un fluage ou un boudinage de la croûte inférieure (Brun & Beslier 1995 ; Driscoll & Karner 1998 ; Reston 1988). D'autres suggèrent une activation polyphasée des failles normales (Reston 2007 ; Ranero & Pérez-Gussinyé 2010). Cette évolution polyphasée des marges passives peu magmatiques a été conceptualisée (Lister *et al.* 1986, 1991 ; Whitemarsh *et al.* 2001 ; Reston 2007) avant d'être modélisée numériquement (fig. 3 ; e.g. Nagel & Buck 2004 ; Lavier & Manatschal 2006 ; Huismans & Beaumont 2007). L'identification de reliques fossiles de ces domaines de necking a été plus tardive (Mohn *et al.* 2012). La combinaison de l'architecture crustale observée par la géophysique marine et les observations de terrain ont permis de conceptualiser l'évolution rhéologique et structurale des domaines de necking (Mohn *et al.* 2012 ; fig. 3). L'amincissement crustal de ces domaines s'effectue par le jeu de failles de détachement conjuguées et concaves qui s'enracinent dans la croûte moyenne ductile. Ces structures délimitent un bloc de croûte caractérisé par la disparition des niveaux ductiles (Key stone block, Mohn *et al.* 2012 ; H-block, Lavier & Manaschal 2006; fig. 3).

***Fig. 3**: Résumé des principaux modèles et concepts proposés pour expliquer l'amincissement crustal extrême observé dans les marges passives (a) Modèles impliquant un fluage ou un boudinage de la croûte inférieure, l'amincissement est fonction de la profondeur, Reston (1988) ; Brun & Beslier (1995) ; Driscoll & Karner (1998) (b) Système de failles polyphasées : la croûte inférieure est rigide, Reston (2007), (c) Modèle d'évolution rhéologique et structural illustrant l'importance de la croûte moyenne (d) Modèles numériques (de haut en bas : modèle d'évolution conceptuelle et numérique proposé par Lavier & Manatschal 2006 et modèle numérique proposé par Huismans & Beaumont 2007).*


(a) Depth Dependant Stretching: Implication of a lower crustal flow or boudinage
Brun and Beslier 1996
LC
LC
Lower crustak flow
towards the continent
Reston 1988
LC
Driscoll and Karner 1998
LC
Lower crustal flow
towards the ocean
(b) Polyphase faulting: brittle lower crust
Reston 2007
UC
LC
(c) Polyphased strain partitioning:
Implication of a ductile middle crust
Mohn et al. 2012
Future proximal margin
Future distal margins
Future proximal margin
0 km
30 km
Key stone
Necking zone
Necking zone
strength (MPa)
500
1500
felsic upper crust
(brittle)
felsic middle crust
(ductile)
mafic lower crust
(brittle)
"strong" mantle
(rheology at the onset of rifting)
Uplift
Grosina Det
Extraction fault
Necking zone
Thermal buoyancy
OCT
Distal margin
Necking zone
Proximal margin
Err Det
Bernina Det
Upper crust
Ductile middle crust
Mafic lower crust
Upper subcontinental mantle
Infiltrated subcontinental mantle
Asthenospheric mantle
Thinning fault
Exhumation fault
(d) Dynamic models
Lavier and Manatschal, 2006
Stretching mode
Thinning mode
Exhumation mode
(thickness of the basins not to scale)
Gabbros
Ductile SZ
Serpentinized upper mantle
Distance
Huismans and Beaumont, 2007

1.4. Architecture des marges passives peu magmatiques : terminologie

L'architecture 2D des marges passives peu magmatiques est relativement bien contrainte par les observations issues de la géophysique marine (Reston 2009 et références citées) tandis que les analogues fossiles apportent des contraintes quant aux structures qui permettent d'accommoder l'amincissement lithosphérique (Müntener *et al.* 2010). Ainsi les travaux menés en mer et à terre mettent en évidence l'évolution polyphasée des processus de rifting (Whitemarsh *et al.* 2001 ; Reston 2007 ; Péron-Pinvidic & Manatschal 2009) ainsi que l'amincissement extrême de la croûte observé dans les marges passives peu magmatiques. L'architecture des marges dites «hyper-amincies» est caractérisée par différents domaines structuraux liés à l'évolution polyphasée du rift. La terminologie développée et utilisée dans ce mémoire de thèse est présentée sur la figure 4.

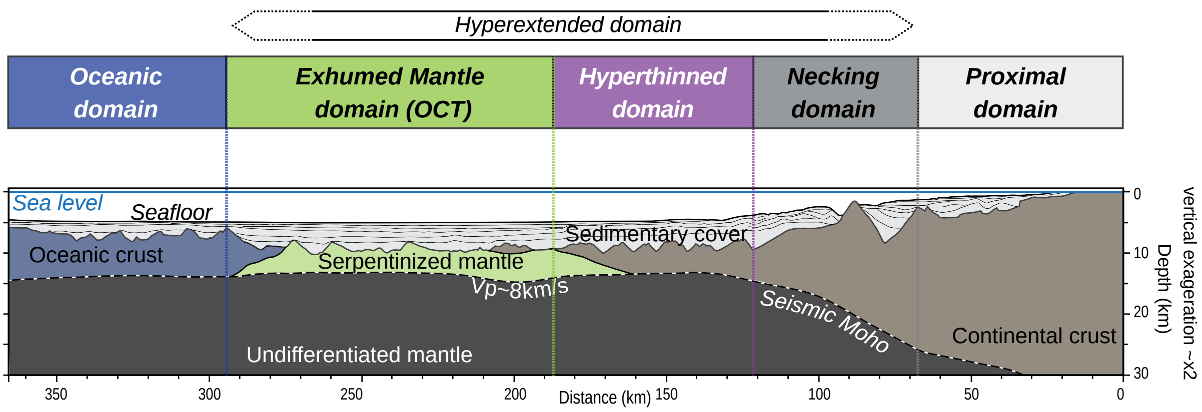


Fig. 4: *Architecture simplifiée d'une marge passive «hyper-amincie» : exemple de la marge Ouest Ibérique (basé sur la line de sismique réfraction IAM 5, Afilhado et al. 2008) et terminologie employée dans le manuscrit.*

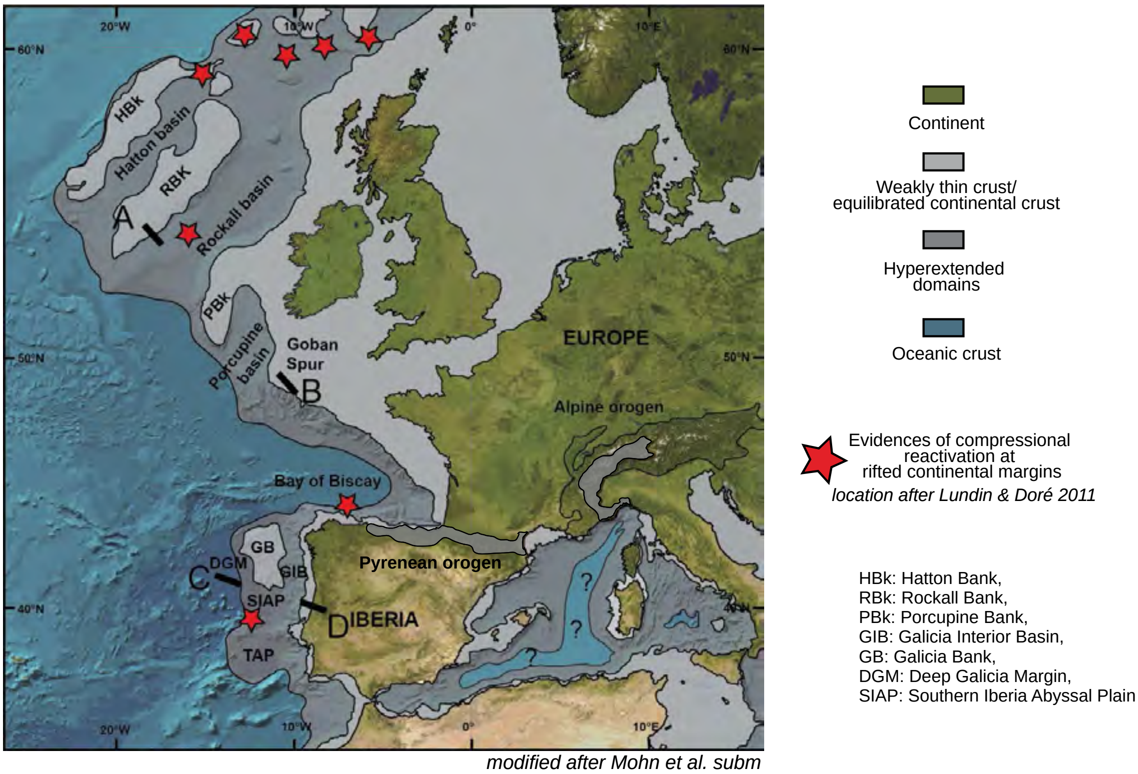
2. Questions et objectifs scientifiques de ce travail

Ainsi, comme cela a été exposé ci-dessus, la compréhension des processus associés à l'amincissement crustal et lithosphérique nécessite un aller-retour constant entre les observations de géologie de terrain et les données de géophysique marine. Les acquisitions de données de sismique réflexion et réfraction dans les marges passives actuelles permettent d'imager l'architecture crustale et stratigraphique. En revanche, les reliques fossiles préservées à terre sont déformées et intégrées dans les chaînes de montagnes (fig. 5a). Les analogues fossiles fournissent néanmoins un accès aux structures de rift, à la nature du socle ainsi qu'à l'enregistrement sédimentaire qui caractérise les différents domaines de la marge. Dans les exemples de marges actuelles, ces observations ne sont accessibles que localement par de rares forages dans les domaines les plus distaux des marges. La complémentarité entre les observations faites à terre et en mer est indéniable, cependant, la majorité des travaux est généralement focalisée sur l'un ou autre type de données. En effet, les différents types d'informations sont souvent difficiles à relier entre elles de par les différentes résolutions et échelles d'observations qui proviennent de régions caractérisées par des évolutions géodynamiques différentes. Ainsi il est nécessaire de développer une approche à l'interface entre les observations faites à terre et en mer.

D'autre part, les connaissances des processus de rift ont fortement évolué depuis l'introduction de la tectonique des plaques et les premiers modèles conceptuels. Malgré tout, l'évolution de ces processus dans le temps et dans l'espace ainsi que l'architecture sédimentaire associée restent mal contraintes même si certains modèles numériques et conceptuels ont déjà été proposés (fig. 5b). De plus, l'utilisation d'un exemple géologique concret est nécessaire afin de caractériser l'évolution 3D de l'architecture mais également le rôle de l'héritage crustal et lithosphérique pré-rift pour les variations d'architectures observées le long des marges (e.g. marge Nord Gascogne, Thinon 1999 ; marge de Vøring et Lofoten : Tsikalas *et al.* 2008 ; marge Ibérique : Sutra & Manaschal 2012).

Finalement, puisque les domaines de rift sont observés en mer mais également intégrés dans les chaînes de montagnes, la question du rôle de l'héritage extensif dans la formation des orogènes peut également être posée. De plus, de récents travaux ont mis en évidence un lien étroit entre les processus d'hyperextension et notamment le rôle de la serpentinisation du manteau exhumé pour la localisation des processus de réactivation (Lundin & Doré 2011 ; fig 5a).

(a) Distribution of hyperextended domains and onshore analogues in the Southern North Atlantic



(b) 3D evolution of continental break-up: conceptual and numerical model

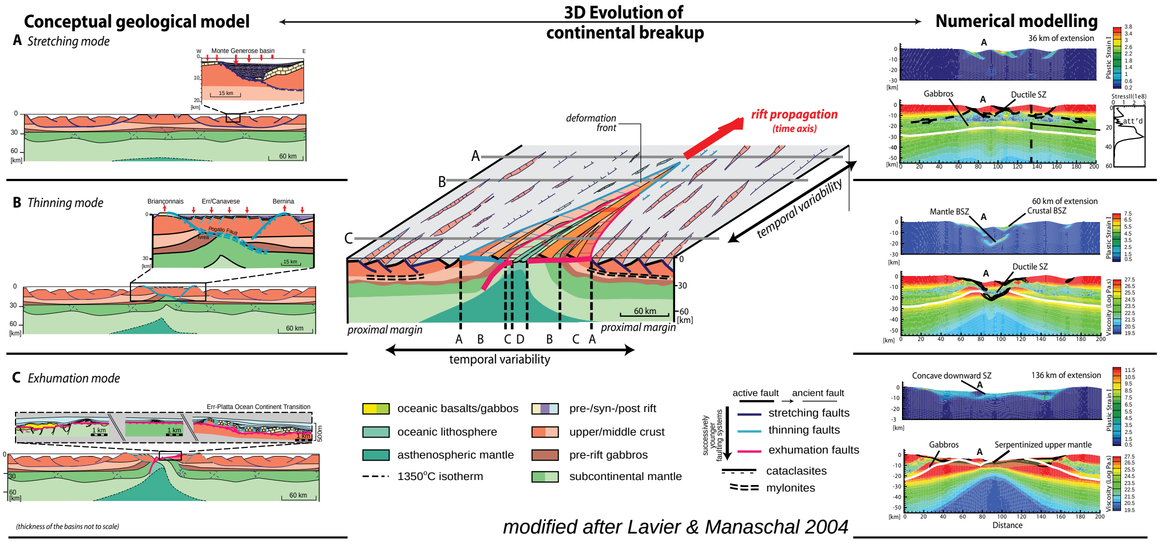


Fig. 5: (a) Distribution des domaines hyper-amincis dans le Sud de l'Atlantique Nord (d'après Mohn et al. subm) et évidences de réactivation compressive associée à ces domaines (d'après Lundin et Doré 2011). (b) Modèle conceptuel et numérique de l'évolution spatiale et temporelle du processus de rupture continentale, d'après Lavier & Manaschal 2006.

L'objectif de cette thèse est d'apporter des éléments de réponse à ces thématiques majeures pour la compréhension et la caractérisation des marges passives actuelles. Ainsi, trois questions essentielles représentent le fil conducteur de mon travail de thèse et sont rattachées à la structure du mémoire.

A. Comment peut-on caractériser et identifier des domaines de rift en mer et dans leurs analogues fossiles à terre?

B. Quelle est l'évolution tectonique, spatiale et temporelle des systèmes de rift soumis à un amincissement extrême dans le contexte des marges passives peu magmatique?

C. Quel est le rôle de l'hyper-extension lors des processus de réactivation et de formation des orogènes?

3. Le Golfe de Gascogne et les Pyrénées, un laboratoire naturel

Le choix du chantier représente un point capital afin d'aborder les thématiques précédemment exposées. Ainsi, dans ce travail, je me focaliserai sur l'étude du système Golfe de Gascogne–Pyrénées à la frontière entre la France et l'Espagne (fig. 6). En effet, ce système est l'un des rares exemples où l'on peut avoir accès à des exemples de marges passives actuelles dans le Golfe de Gascogne et à leurs analogues fossiles préservés à terre dans les Pyrénées (e.g. bassin d'Arzacq–Mauléon : Jammes *et al.* 2009 ; Lagabrielle *et al.* 2010).

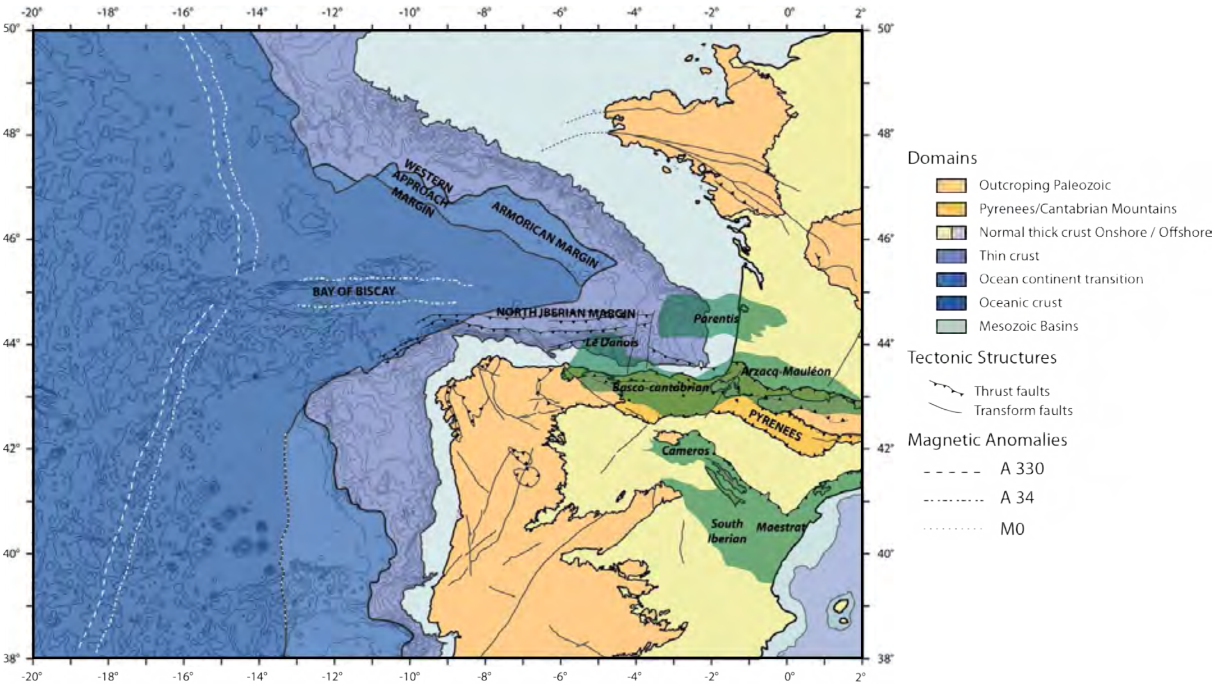


***Fig. 6**: Carte du Golfe de Gascogne et des Pyrénées : structures tectoniques majeures et différents domaines du Golfe de Gascogne (zonation établie d'après Thinon 1999 et Jammes et al. 2009)*

3.1. Contexte régional : cadre structural

Le Golfe de Gascogne correspond à un bassin océanique en forme de V ouvert à l'Ouest vers l'Océan Atlantique et se terminant à l'Est dans le bassin de Parentis. Il est bordé au Nord par la marge Nord Gascogne et au Sud par la marge Nord Ibérie ou Cantabre (fig. 6). A l'extrémité Est du Golfe de Gascogne, plusieurs bassins de rift ont été identifiés. Les données géophysiques et géologiques de ces bassins indiquent qu'ils ont subi un amincissement extrême de la croûte. Parmi eux, il est possible de citer les bassins de Parentis (Pinet et al. 1987; Bois & Gariel 1994; Tomassino & Marillier 1997; Jammes *et al.* 2010a, 2010c), d'Arzacq–Mauléon (Daignières *et al.* 1994; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010), d'Aulus (Lagabrielle & Bodinier 2008; Lagabrielle *et al.* 2010; Clerc *et al.* 2012), ainsi que le bassin Basque–Cantabre (Pedreira *et al.* 2003, 2007 ; Roca *et al.* 2011).

3.1.1. Le Golfe de Gascogne

Le Golfe de Gascogne est bordé au Nord par la marge Nord Gascogne considérée par de nombreux auteurs comme typique des marges passives peu magmatiques (Montadert *et al.* 1974, 1979 ; de Charpal *et al.* 1978 ; Barbier 1986 ; Le Pichon & Barbier 1987). L'architecture traditionnellement proposée est caractérisée par une série de blocs basculés, délimités par des failles listriques dont la taille diminue progressivement vers l'océan (Duvergé & Poutchkovsky 1983 ; Barbier *et al.* 1986 ; Le Pichon & Barbier 1987). En détail, la structuration de la marge Nord Gascogne est plus complexe et présente des variations latérales d'architecture. Trois segments de marges sont ainsi distingués et caractérisés par des orientations différentes : la marge de Goban, la marge des Entrées de la Manche et la marge Armoricaine qui se termine dans le bassin de Parentis (Thinon 1999 et références citées ; Thinon *et al*. 2003).

Contrairement à la marge Nord Gascogne, l'architecture du rift n'est pas préservée dans la marge conjuguée Nord Ibérie. Elle est classiquement interprétée comme une marge active fossile caractérisée par une subduction du domaine océanique (Boillot 1984 ; Alvarez-Marron *et al.* 1997). De récentes études de sismique réfraction remettent en cause l'hypothèse d'une importante subduction océanique (Ruiz 2007) et suggèrent que le prisme d'accrétion observé au pied de la marge résulte de la réactivation partielle à totale de l'ancienne marge passive (Gallastegui *et al.* 2002; Fernández-Viejo & Gallastegui 2005; Alonso *et al.* 2007; Roca *et al.* 2011; Fernández-Viejo *et al.* 2012) et de son intégration au système orogénique Pyrénéen (Fernández-Viejo *et al.* 1998 ; Roca *et al.* 2011).

Le domaine central du Golfe de Gascogne correspond à un domaine océanique qui a été caractérisé très tôt par sismique réfraction (Bacon *et al.* 1969) et également défini par la présence d'anomalies magnétiques orientées E–W à ESE–WNW (Matthews & Williams 1968). La limite entre les croûtes continentale et océanique a longtemps été considérée comme une limite structurale simple (e.g. Avedik & Howard 1979 pour le Golfe de Gascogne). Cependant, de récents travaux menés sur la marge Armoricaine et la marge Nord Ibérie suggèrent que cette transition pourrait être en partie caractérisée par un domaine de manteau exhumé (Thinon *et al.* 2003 ; Roca *et al.* 2011; Fernández-Viejo *et al.* 2012).

3.2.1. Les Pyrénées et le bassin Basque–Cantabre

Le système orogénique Pyrénéen au sens large s'étend sur près de 1000 km de long depuis la Provence à l'Est jusqu'aux Asturies en Espagne. L'architecture des Pyrénées est caractérisée par une structure en éventail fortement asymétrique et essentiellement déversée vers le Sud (Casteras 1933 ; Mattauer 1968). Les Pyrénées sont classiquement divisées en 5 domaines structuraux majeurs (fig 7 ; Mattauer 1968 ; Choukroune & Séguret 1973 ; Mattauer & Henry 1974). Au Nord et au Sud, la chaîne Pyrénéenne est bordée par deux bassins d'avant–pays : respectivement le bassin Aquitain et celui d'Ebro.

La zone Nord Pyrénéenne chevauche le bassin Aquitain au Nord, le long du chevauchement Nord Pyrénéen, et est limitée au Sud par la Faille Nord Pyrénéenne. Cette structure est classiquement interprétée comme une discontinuité majeure entre les plaques Européenne et Ibérique (Mattauer 1968; Choukroune & Mattauer 1978; Choukroune & ECORS Team 1989). Les sédiments mésozoïques de ce domaine sont caractérisés par un métamorphisme de haute température et de basse pression qui devient de plus en plus important à la bordure Sud du domaine. Les massifs paléozoïques Nord Pyrénéens (e.g. Agly, Saint-Barthélemy, Arize, Trois Seigneurs…) caractérisent l'Est de la chaîne et présentent souvent un contact tectonique avec les sédiments encaissants.

La zone Axiale au Sud de la faille Nord Pyrénéenne est formée par des terrains Paléozoïques hérités de l'orogène varisque. Bien souvent, le socle paléozoïque préserve sa couverture permienne à Trias inférieur. Ce domaine est bien défini dans les Pyrénées Orientales et Centrales mais n'affleure quasiment plus à l'Ouest de la chaîne.

La zone Sud Pyrénéenne, limitée au Nord par la faille Sud Pyrénéenne, est caractérisée par des sédiments mésozoïques à cénozoïques, transportés vers le Sud sur le bassin d'Ebro.

Cette structuration ne s'applique pas aussi bien à l'ensemble des Pyrénées. En effet, la continuité vers l'Ouest de certaines structures, et notamment la faille Nord Pyrénéenne, est fortement discutée.

Le bassin Basque–Cantabre forme la terminaison Ouest du système orogénique Pyrénéen, au sens large, mais est caractérisé par une stratigraphie et une structuration différente (fig 7). Il est limité à l'Est par la faille de Pamplona orientée NE–SW, en partie héritée de l'histoire post-varisque (Capote 1983 ; Rat 1988 ; Martinez-Torrez 1989). Cette structure représente une discontinuité crustale majeure (Jammes *et al.* 2010c) et contrôle à la fois la sédimentation mésozoïque (Larrasoaña *et al.* 2003) et la déformation Pyrénéenne.

Les structures compressives sont orientées E–W immédiatement au contact de la faille de Pamplona (e.g. faille de Leiza) et NW–SE à l'Ouest de la faille d'Hendaye, parallèle à la structure de Pamplona. Le bassin est caractérisé par des structures anticlinales et synclinales à vergence NE qui sont, du SW au NE : l'anticlinal de Bilbao, le synclinal de Viscaya (Biscay), l'anticlinal Norvizcaíno (Nord Biscay) et le synclinal de Zumaya.

La faille de Leiza est souvent considérée comme la continuation de la faille Nord Pyrénéenne dans le bassin Basque–Cantabre (Boillot *et al.* 1973; Choukroune 1976; Rat 1988; Combes *et al.* 1998; Mathey *et al.* 1999). Elle délimite la nappe des marbres (Lamare 1936; Mendia & Gil-Ibarguchi 1991) qui est caractérisée par des affleurements de manteau et de sédiments mésozoïques métamorphisés, similaires à ceux observés dans la zone Nord Pyrénéenne.

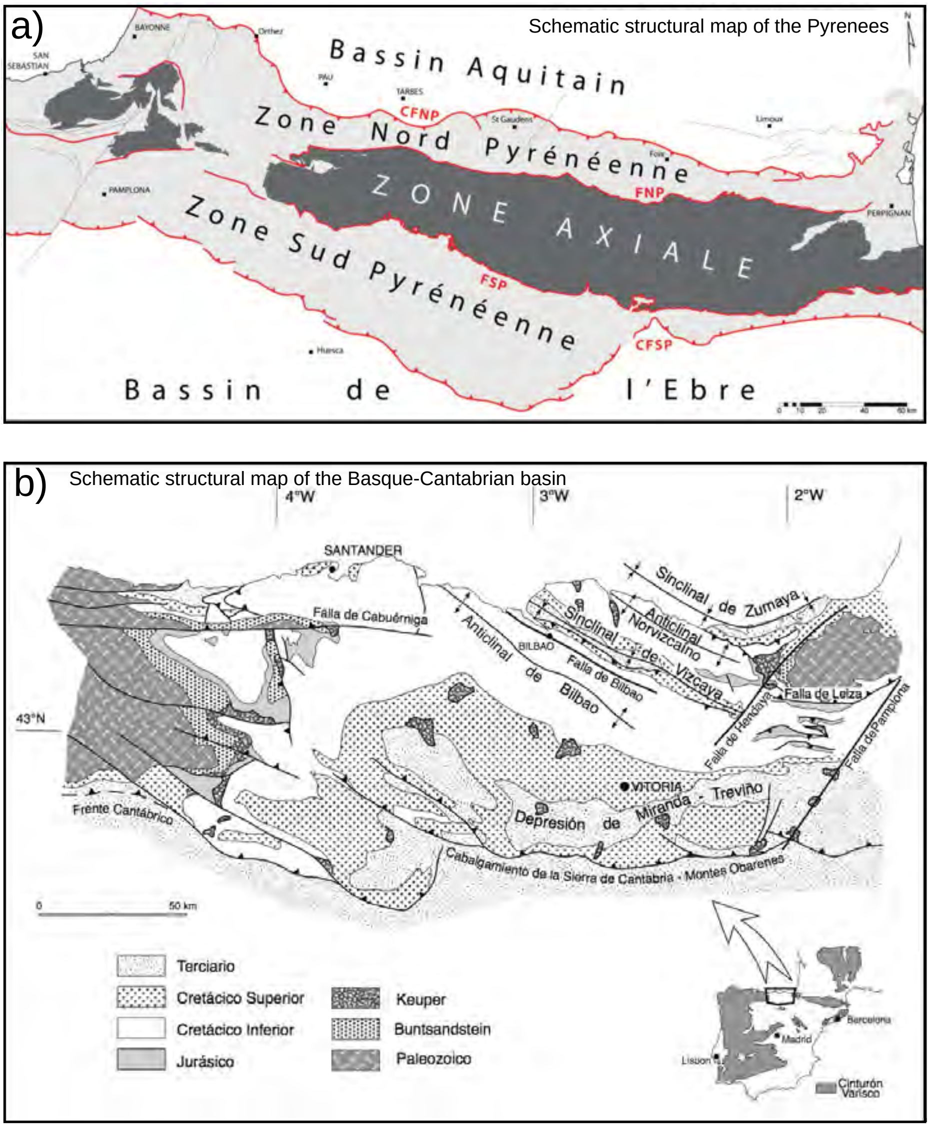


***Fig. 7**: (a) Schéma structural de la chaîne pyrénéenne présentant cinq unités structurales distinctes ainsi que les failles qui les délimitent. CFNP : Chevauchement Frontal Nord Pyrénéen ; FNP : Faille Nord Pyrénéenne ; FSP : Faille Sud Pyrénéenne ; CFSP : Chevauchement Frontal Sud Pyrénéen (d'après Clerc 2011). (b) Schéma structural du bassin Basque-Cantabre montrant les structures principales (d'après Pedreira 2004).*

3.2. Evolutions paléogéographique, tectonique et sédimentaire

La structuration actuelle du Golfe de Gascogne et des Pyrénées résulte de la succession de plusieurs cycles tectoniques extensifs et compressifs initiés dès le début du Paléozoïque. Les épisodes tectoniques majeurs sont synthétisés ci-après.

3.2.1. Héritage pré-rift

3.2.1.1. Orogénèse hercynienne (évolution paléogéographique, structuration)

La formation de l'orogène varisque au cours du Carbonifère résulte de la collision des plaques Laurussia et Gondwana, suite à la fermeture progressive de domaines océaniques (océans Rhéic et Galice–Massif Central) séparés par des micro-continents comme Armorica (fig 8 ; Matte 1991, 2001). Selon Matte (1991), le domaine Pyrénéen appartenait aux domaines externes de la chaîne varisque (fig 9). Au Carbonifère, suite à la compression liée à la formation de la chaine varisque, des sédiments de faciès Culm (grès, silts et pélites turbiditiques) sont déposés dans des bassins profonds. Cette phase compressive est associée à la mise en place de nombreux plutons granitiques et d'un métamorphisme de contact avec les terrains environnants.

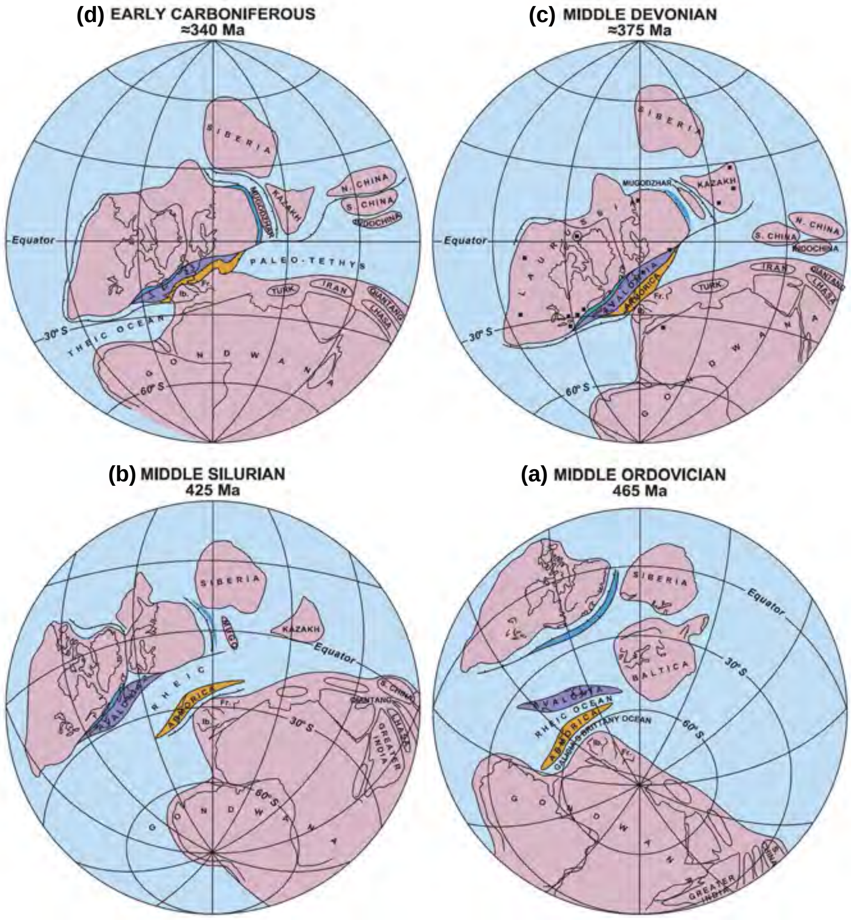


***Fig. 8**: Reconstruction de l'évolution Paléozoique de l'Ordovicien moyen (465 Ma) au Carbonifère inférieur (340 Ma). Orange : microplaque continentale d'Armorica ; Violet : microplaque continentale Avalonia ; Bleu arc insulaire (d'après Matte 2001).*

Fig. 9*: Schéma structural de la chaîne varisque d'Europe (Matte 1986). 1 : principaux chevauchements ; 2 : nappes cristallines internes et sutures ophiolitiques ; 3 : domaines à schistosité de flux ou foliation ; 4 : zones externes ; 5 : bassins dévono-carbonifères ; 6 : socle prévarisque peu ou pas déformé ; 7 : sens de transport des nappes et déversement des grands plis couchés ; 8 : grands décrochements ductiles.*

3.2.1.2. Extension tardi-hercynienne

L'évolution tardi-varisque est caractérisée par la mise en place de failles de décrochement ou de transfert à partir de la fin du Carbonifère et jusqu'au début du Permien (e.g. failles de Toulouse, des Cévennes, de Ventaniella ou encore la faille Nord Pyrénéenne). Deux scénarios sont principalement proposés pour caractériser cette évolution tardi-varisque : un épisode de fracturation antérieur au Trias et indépendant de la tectonique varisque (Arthaud & Matte 1975, 1997 ; fig 10) ou une phase extensive à transtensive liée à l'effondrement gravitaire de la chaîne varisque (Burg *et al.* 1994a, 1994b ; fig 10). Ces deux interprétations restent débattues mais l'évolution tardi-varisque semble avoir structuré durablement la lithosphère.

Au cours du Permien certaines de ces structures sont réactivées en failles normales contrôlant la mise en place de bassins intracontinentaux (fig 10b). Cet épisode est également associé à la mise en place d'un épisode magmatique calco-alcalin.

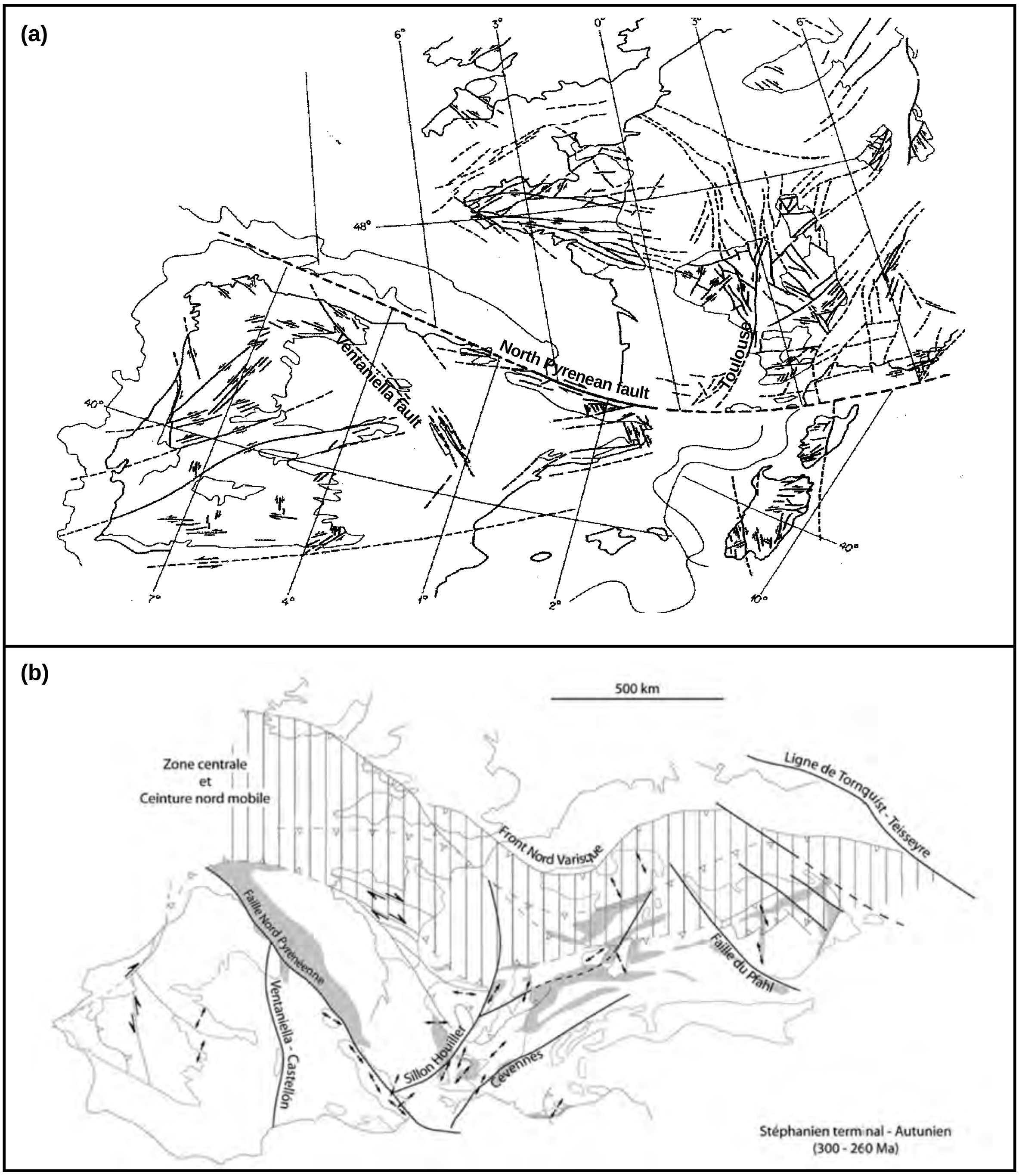


***Fig. 10**: (a) Carte des principaux décrochements tardi-varisques reconnus ou supposés d'après Arthaud & Matte 1975. (b) Directions d'extensions du Carbonifère supérieur au Permien inférieur dans la chaîne varisque d'Europe occidentale (d'après Burg et al. 1994b). Les principaux bassins sédimentaires mis en place au cours du Stéphanien supérieur au Permien inférieur sont grisés.*

3.2.1.3. Rifting triassique

Un épisode de rift succède à la phase tardi-varisque et correspond à la mise en place de bassins intra-continentaux délimités par des failles normales orientées NE–SW au cours du Trias (figs 11 & 12). Les anciennes structures paléozoïques peuvent être en partie réactivées (Curnelle *et al.* 1982). D'épaisses séquences de sédiments silico-clastiques, de carbonates et d'évaporites, caractéristiques du Trias, sont ainsi déposées (Faciès Germanique : Curnelle 1983; Fréchengues 1993). Dans les domaines les plus profonds des bassins, les formations sédimentaires du Keuper sont associées à un magmatisme tholéitique (e.g. Montadert & Winnock 1971; Winnock 1971; Rossi *et al.* 2003). De la fin du Trias au Jurassique, une transgression marine généralisée est associée à la formation d'une plateforme carbonatée dont l'étendue exacte est mal contrainte (fig 12).

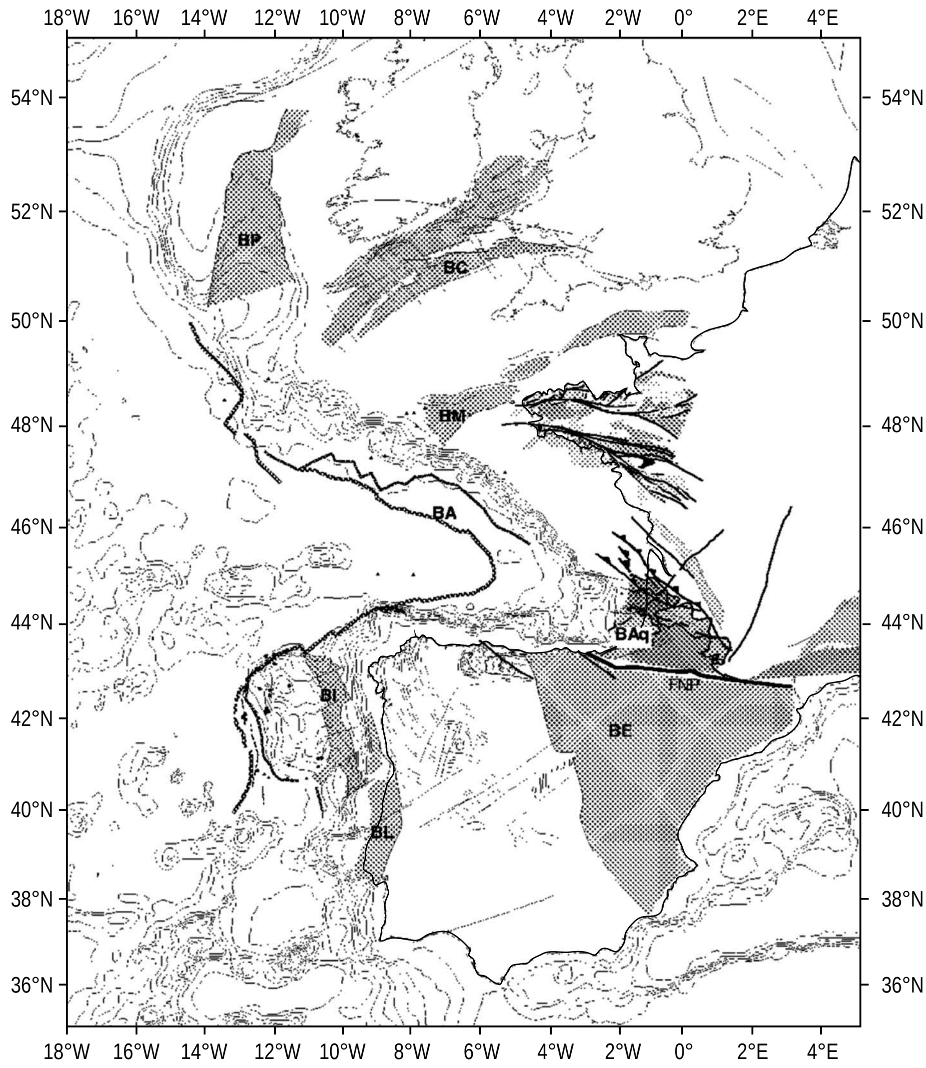


***Fig. 11**: Carte des bassins Triasiques des plaques Ibérie et Europe. BA : bassin armoricain, BM : bassin de la Manche ; BC : bassin Celtique ; BE : bassin d'Ebro ; BL : bassin de Lusitanie ; BI : bassin intérieur de Galice ; Baq : bassin Aquitain (d'après Thinon 1999).*

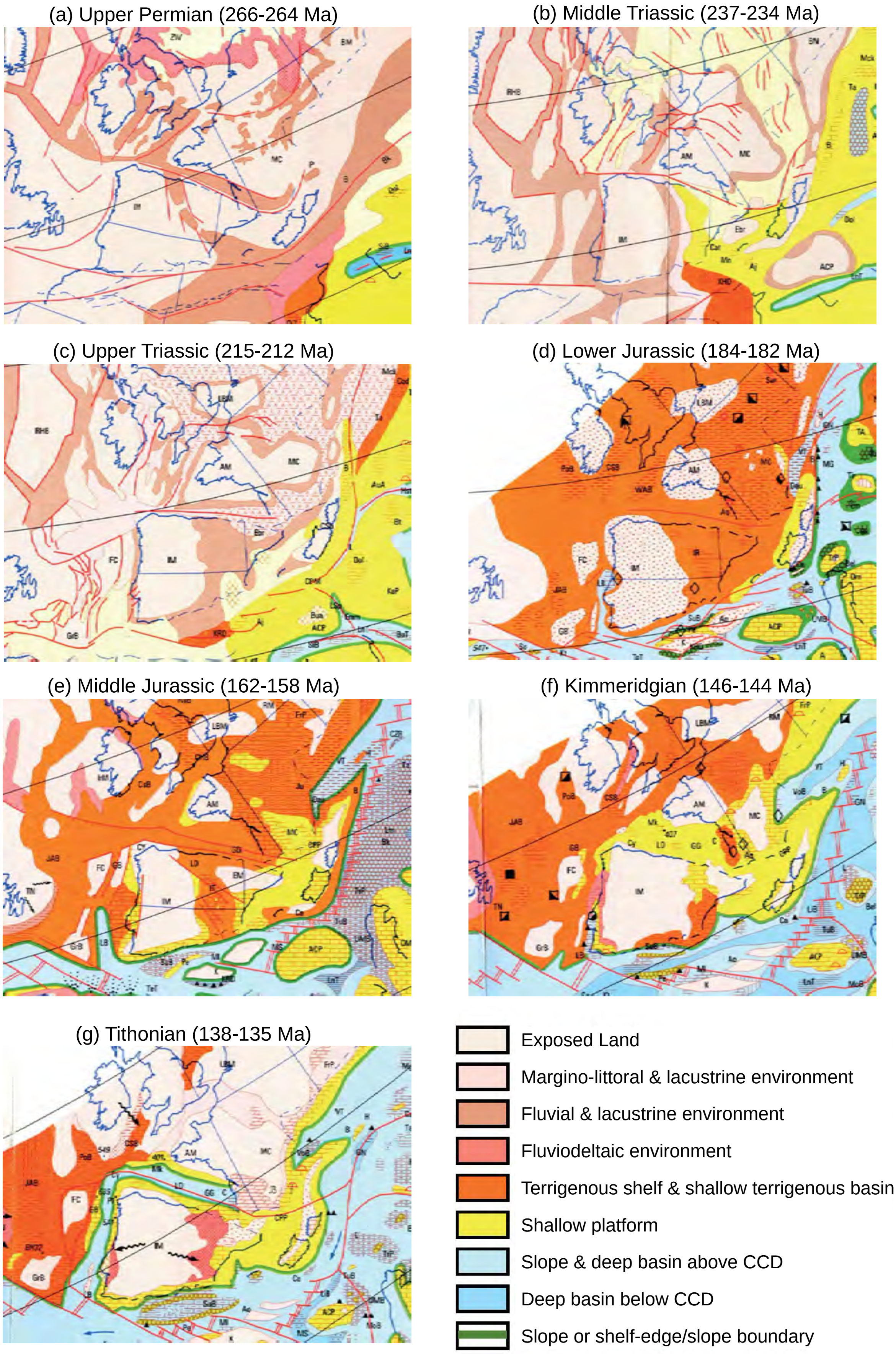


***Fig. 12**: Evolution paléogéographique proposée par Dercourt et al. (2000) pour la période du Permien au Jurassique*

3.2.2. Rifting et amincissement extrême de la lithosphère (Jurassique sup-Crétacé)

3.2.2.1. Contexte cinématique

L'évolution cinématique associée à la formation du Golfe de Gascogne, au cours du Jurassique terminal jusqu'au Crétacé inférieur, est le sujet de nombreuses controverses. La plupart des débats associés au mouvement de l'Ibérie concernent la quantité de déplacement, la cinématique des évènements ainsi que l'interprétation des pôles de rotation. Des contextes tectoniques complétement différents sont ainsi proposés.

Sibuet et al. (2004) suggèrent que l'ouverture du Golfe de Gascogne se fait en deux temps. Entre le Jurassique Supérieur et l'Aptien (M25 : 156.5 Ma à M0 : 118 Ma), une large extension N–S est accommodée dans le domaine Pyrénéen aboutissant à la formation d'un océan que les auteurs appellent la néo-Téthys. Une seconde phase d'extension dans le Golfe de Gascogne, entre l'Aptien Supérieur et le Santonien Supérieur (M0 à A33o : 85 Ma), est accommodée par une subduction entrainant la formation de bassins d'arrière-arc dans le domaine Pyrénéen (fig. 13). D'autres hypothèses proposent que le déplacement de l'Ibérie soit associé à des mouvements décrochants ou transtensifs accommodés le long de la faille Nord Pyrénéenne ou par la formation de bassins en pull-apart (e.g. Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Choukroune & Mattauer 1978 ; Olivet 1996 ; fig 14).

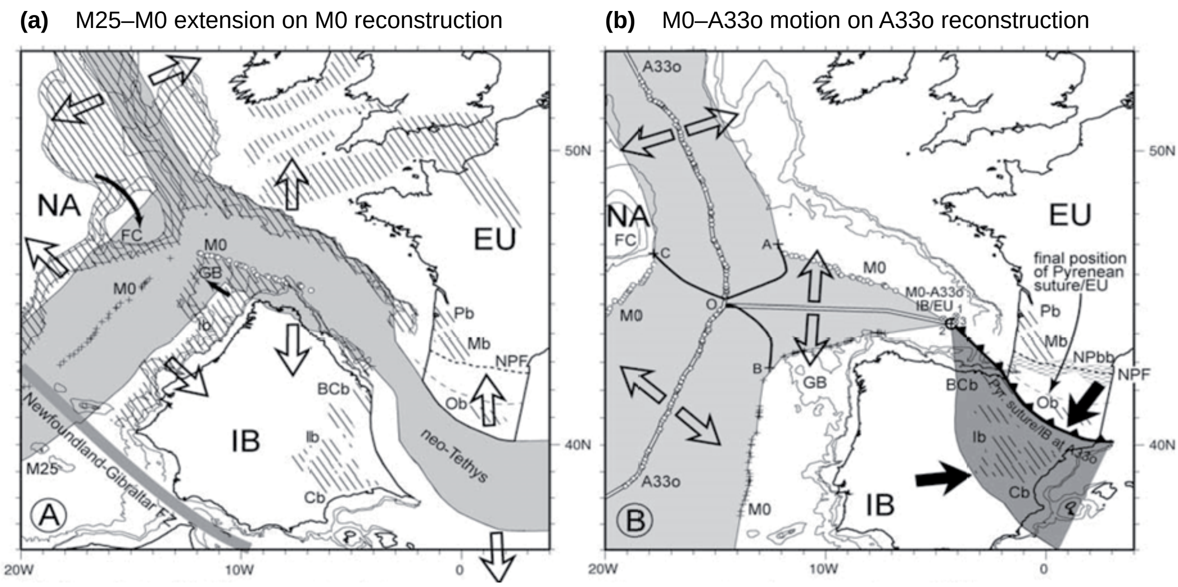
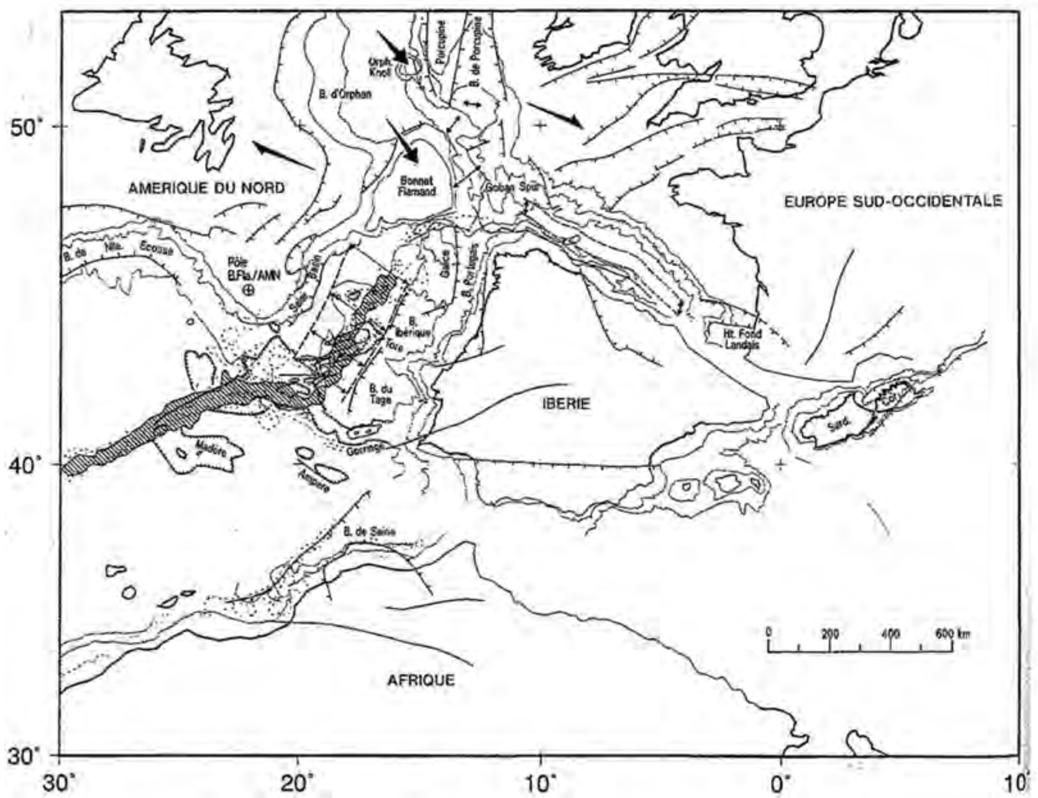


Fig.13: Modèles d'évolution cinématique proposé par Sibuet et al. 2004. L'Europe est supposée fixe. (a) Restauration des continents à l'anomalie M0. (b) Restauration à l'anomalie A33o.

(a) Reconstruction à l'anomalie J (Barremien 130–125 Ma)



(b) Reconstruction à l'Albien moyen (107 Ma)

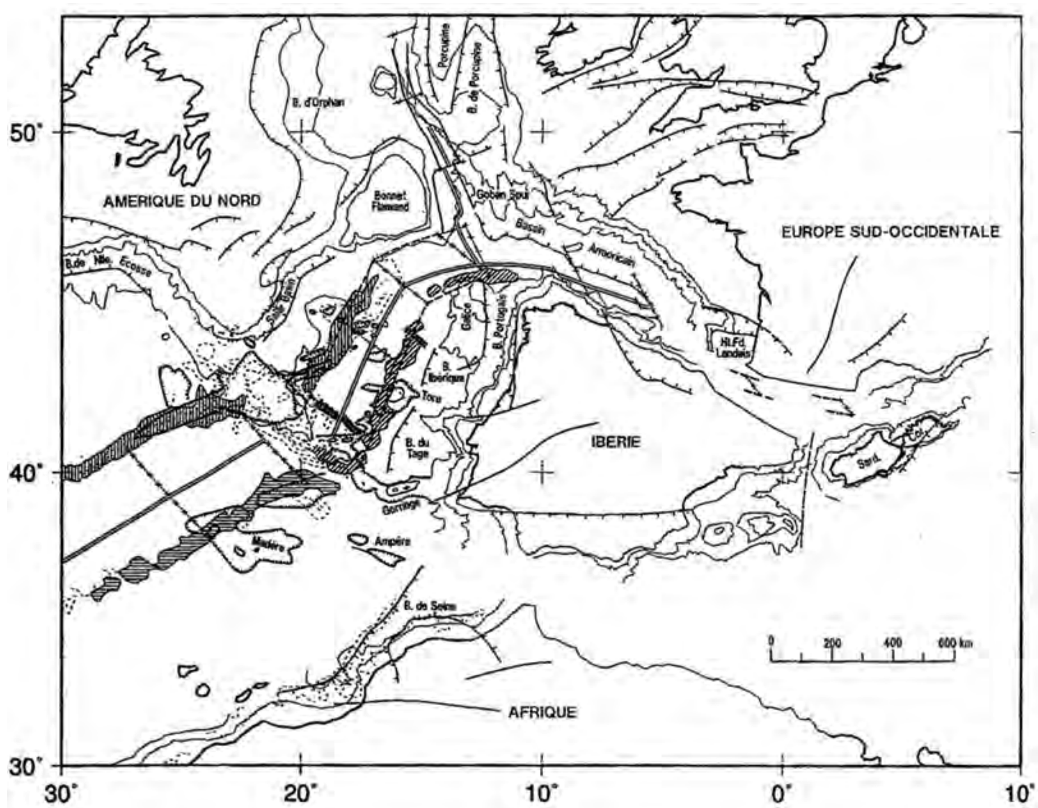
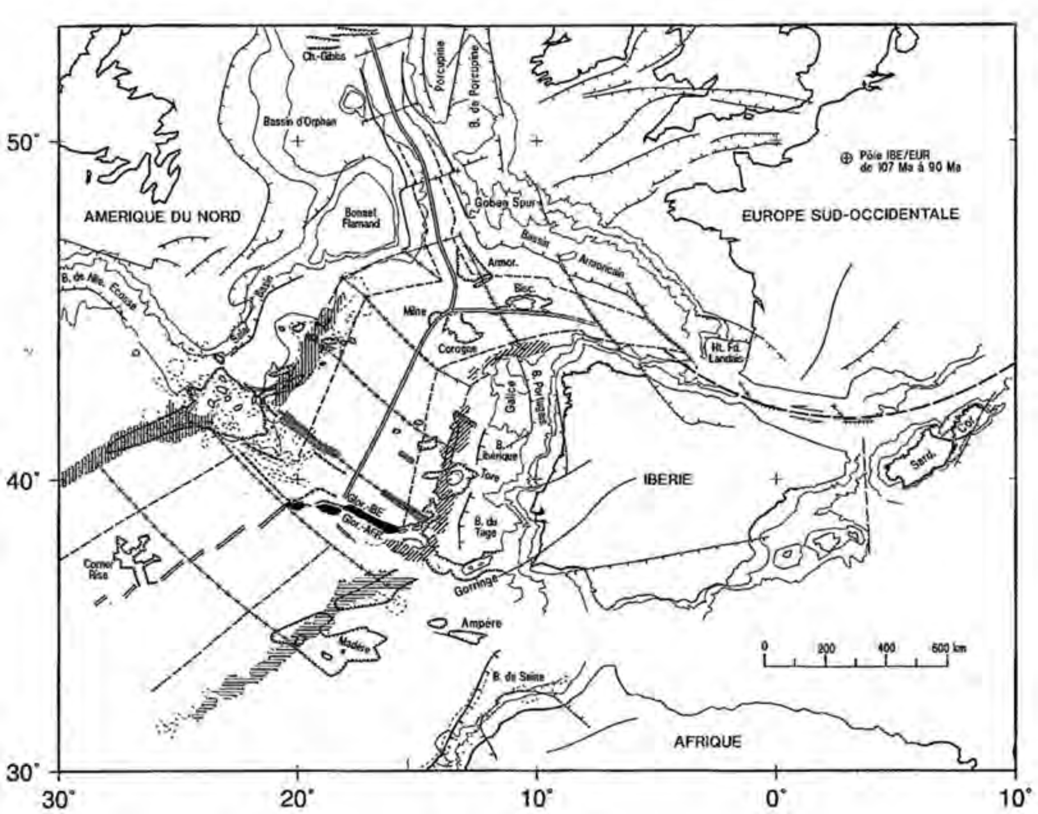
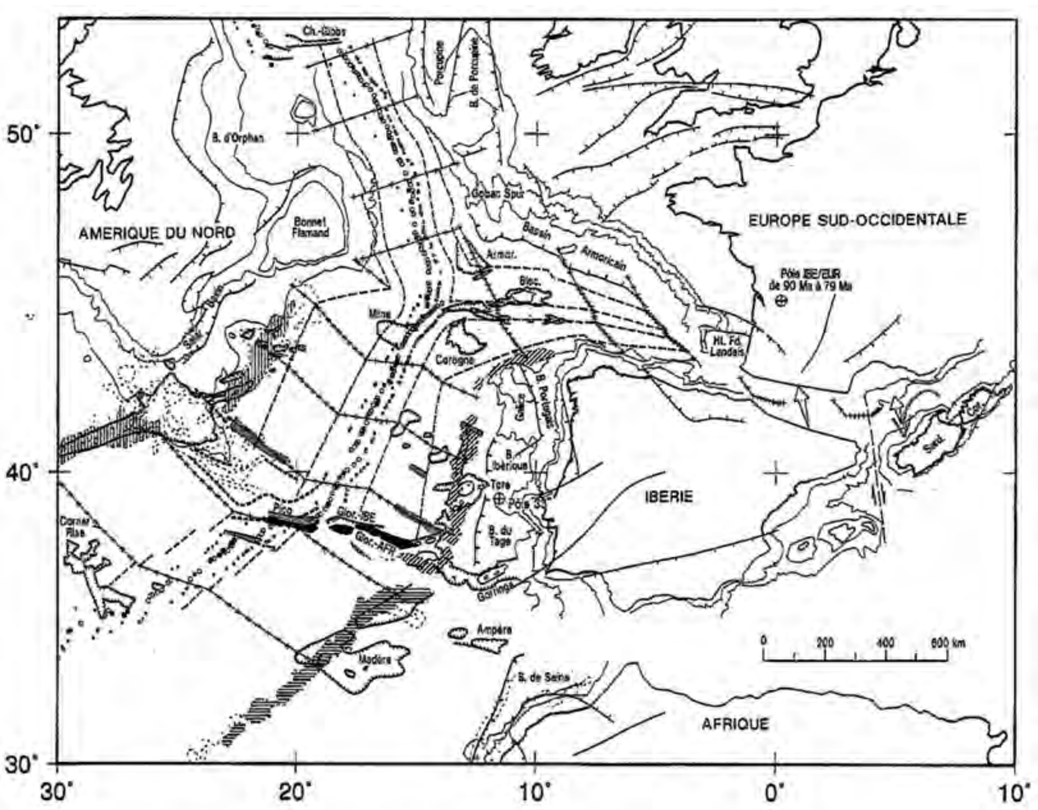


***Fig. 14**: Reconstruction cinématique de l'Ibérie proposé par Olivet 1996. L'Europe est supposée fixe. (a) Reconstruction à l'anomalie J. (b) Reconstruction à l'Albien moyen. (c) Reconstruction au Sénonien inférieur. (d) Reconstruction au Campanien moyen.*

(c) Reconstruction au Sénonien inférieur (90 Ma)



(d) Reconstruction au Campanien moyen (79 Ma)



Les nombreuses incertitudes sont en partie associées à l'identification, à l'interprétation et à la restauration controversée des anomalies magnétiques de la série M, dans le Golfe de Gascogne et dans l'Atlantique Nord en général (M3–M0, 126 to 118.5 Ma). Ainsi, la plupart des restaurations basées sur les anomalies magnétiques considèrent que les déplacements pré-rupture continentale sont mineurs. Cette hypothèse est remise en question par l'existence de domaines hyperamincis et de manteau exhumé qui sont cartographiés sur plusieurs centaines de kilomètres entre le continent et la première anomalie magnétique attribuée à la rupture lithosphérique (Tucholke *et al.* 2007; Péron-Pinvidic & Manatschal 2009). En prenant en compte les mouvements précédents la rupture lithosphérique, le mouvement de l'Ibérie pourrait avoir été initié dès la fin du Jurassique et accommodé dans un couloir de déformation transtensive entre les plaques Européenne et Ibérique (Wortmann *et al.* 2001; Schettino & Scotese 2002; Canérot 2008; Jammes *et al.* 2009, 2010a).

De plus, de récentes études proposent des origines différentes pour expliquer la formation des anomalies M3 à M0, identifiées entre les marges Ibérie et Terre-Neuve. Selon Sibuet *et al.* (2007), elles pourraient être liées à l'exhumation du manteau. Bronner *et al.* (2011) proposent qu'elles soient la conséquence d'un évènement magmatique majeur qui précéderait la rupture lithosphérique et partiellement surimposée à un ancien domaine de manteau exhumé. Les nombreuses incohérences observées, entre les observations géologiques et les restaurations basées sur l'anomalie M0, suggèrent que ces anomalies ne représentent pas des isochrones et ne devraient donc pas être utilisées pour les restaurations cinématiques, comme discuté par Norton *et al.* (2007) et Bronner *et al.* (2012).

3.2.2.2. Enregistrement des processus d'amincissement

La migration du rift Atlantique vers le Nord est associée à des changements majeurs qui sont observés au cours du Jurassique Supérieur jusqu'au Crétacé Supérieur. Des dépôts marins profonds sont déposés dans le Golfe de Gascogne tandis qu'une réorganisation E–W est observée dans les bassins d'Arzacq, Tarbes et Parentis (BRGM 1974; Biteau *et al.* 2006). Les processus d'amincissement extrême dans le Golfe de Gascogne aboutissent à la rupture continentale et à l'initiation de l'accrétion océanique entre la fin de l'Aptien et le début de l'Albien (e.g. Montadert *et al.* 1979; Boillot 1984). Au même moment, les basins, à présent préservés à terre, enregistrent une accélération de leur subsidence (Debroas 1987 ; 1990) qui est attribuée aux processus d'hyper-extension et d'exhumation de manteau (e.g. basins Basque–Cantabrian et d'Arzacq–Mauléon: Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Roca *et al.* 2011). Cette progressive création d'espace d'accommodation est enregistrée par les dépôts marins profonds (Ducasse & Velasque 1988) qui peuvent être localement intrudés par des roches alcalines entre la fin de l'Aptien et le début du Santonien (~113 to 85 Ma; Lamolda *et al.* 1983; Montigny *et al.* 1986).

3.2.3. Convergence Crétacé supérieur au miocène

L'évolution paléogéographique des marges du Golfe de Gascogne et d'Ibérie-Terre Neuve est relativement bien contrainte après la période magnétique calme du Crétacé et notamment grâce à l'identification de l'anomalie magnétique 34 (83 Ma, Santonian) (e.g. Roest & Srivastava 1991; Rosenbaum *et al.* 2002). La mise en mouvement de la plaque Africaine, dès la fin du Crétacé, initie la déformation compressive dans le Golfe de Gascogne et les Pyrénées (fig 15). L'initiation de la déformation compressive est enregistrée dans les sédiments d'âge Santonien (Garrido-Megías & Ríos 1972; McClay *et al.* 2004) dans les Pyrénées et est associée à une réactivation hétérogène de la marge Nord Gascogne (Thinon *et al.* 2001). La phase majeure de la compression débute au cours de l'Eocène et se poursuit jusqu'à la fin de l'Oligocène (Muñoz 2002; Vergés *et al.* 2002). Elle est associée au soulèvement généralisé de la chaîne Pyrénéenne et à la formation des bassins d'avant-pays.

Même si les directions de la compression sont cohérentes entres les différents modèles cinématiques proposés (N–S pour Roest & Srivastava 1991; et NE–SW pour Rosenbaum *et al.* 2002 ; fig 15), la part entre le raccourcissement accommodé à la fin crétacé et celui accommodé au cours du Cénozoïque reste mal défini. Ainsi, à l'échelle du Golfe de Gascogne et de l'orogène Pyrénéenne au sens large, la transition entre la subduction embryonnaire de la marge Nord Ibérie et la collision continentale observée dans les Pyrénées est interprétée comme la conséquence d'une augmentation de la convergence d'Ouest en Est. La restauration des anomalies magnétiques proposée par Rosenbaum *et al.* (2002) suggère un raccourcissement de 144 km dans les Pyrénées de l'Ouest comparé aux 206 km de raccourcissement estimé pour les Pyrénées de l'Est (fig 15).

3.3. Questions liées au chantier

Ainsi le choix du Golfe de Gascogne et des Pyrénées et sa complexité géologique soulève de nouvelles questions et thématiques plus locales qui seront également abordées dans le manuscrit.

D. Quelle est l'origine de l'évolution de l'architecture à travers le système ?

E. Quelle est la nature de la limite de plaque entre Ibérie et Europe ?

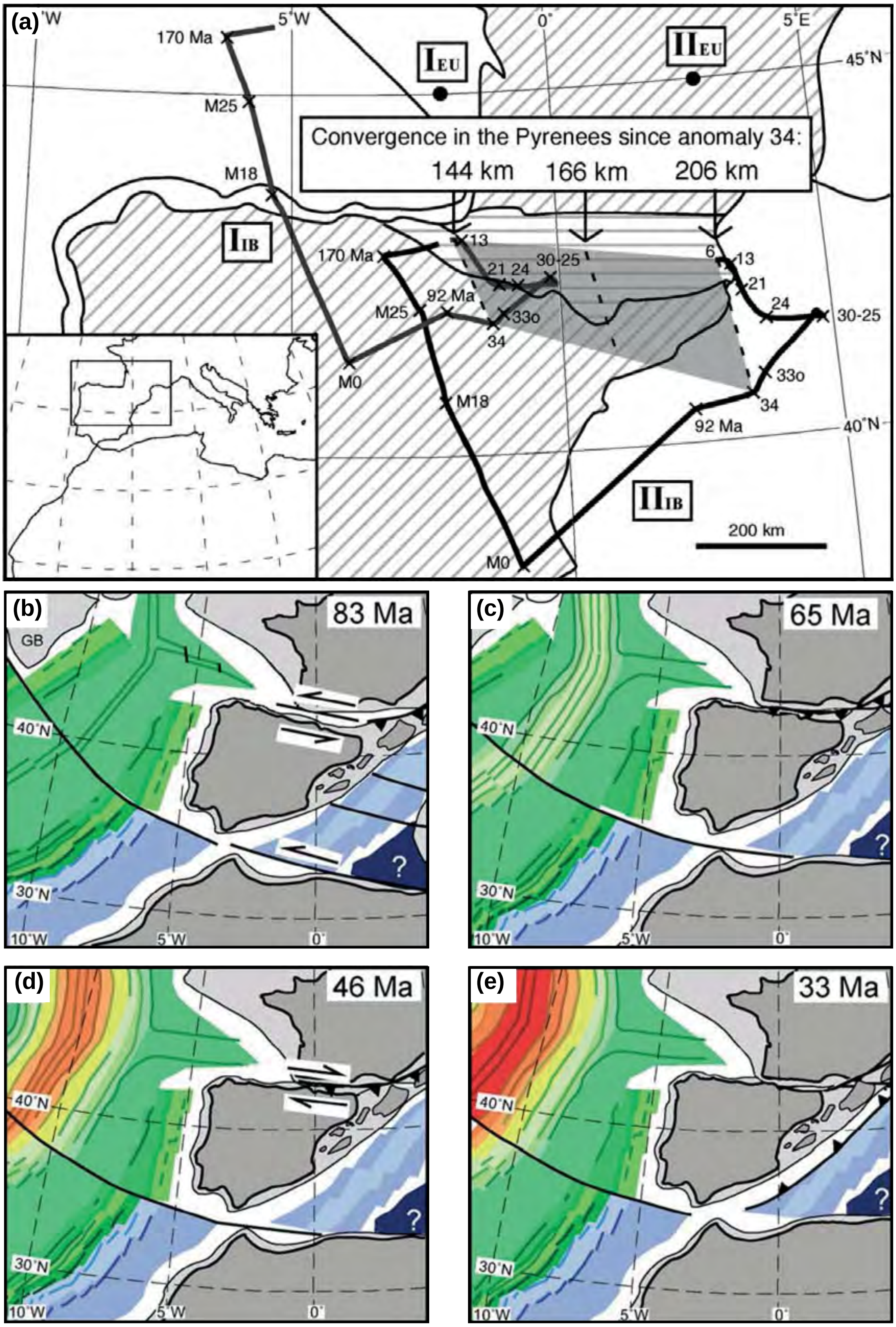


***Fig. 15**: (a) Trajectoire de deux points de l'Ibérie par rapport à deux points fixe d'Europe. La zone grisée indique la quantité de raccourcissement accumulée dans les Pyrénées depuis l'anomalie 34 (d'après Rosenbaum et al. 2002). (b)-(e) Restaurations des mouvements cinématiques de l'Ibérie proposées par Rosenbaum et al. 2002. (b) Au Santonien. (c) A la fin du Maastrichien. (d) A l'Eocène moyen. (e) A la fin de l'Eocène. (D'après Rosenbaum et al. 2002)*

4. Démarche du travail et plan de la thèse

4.1. L'approche Terre-Mer

Le choix du chantier Golfe de Gascogne et Pyrénées impose une approche pluridisciplinaire et pluri-échelle à l'interface entre la géologie de terrain, les données de forage et les méthodes issues de la géophysique marine. Ainsi, la complémentarité des observations terre-mer, au cœur de ce mémoire de thèse, a nécessité dans un premier temps de développer une approche qui me permette d'intégrer et de confronter des observations issues de différentes résolutions et d'échelle d'observation. Ma thèse a aussi pour objectif de faire le lien entre différentes échelles d'observations : du travail de terrain (dans les chaînons béarnais et le bassin de Mauléon) à l'échelle de la limite de plaque entre Ibérie et Europe.

4.1.1. Le travail de terrain (Annexe 1)

Le travail de terrain effectué dans le cadre de la thèse est essentiellement focalisé dans la région des chaînons béarnais avec l'objectif de caractériser les structures associées à la déformation compressive et de comprendre le rôle de l'héritage extensif. Ce travail de terrain a ensuite été replacé et intégré à l'échelle plus globale pour comprendre d'une part l'évolution du bassin d'Arzacq-Mauléon et d'autre part les processus de mise en place des domaines de rift dans l'orogène Pyrénéenne *(Annexe 1 : cartes géologiques)*.

4.1.2. Interprétation sismique et données de forage (Annexe 2 et 3)

Les interprétations sismiques m'ont permis en premier lieu de caractériser l'architecture spécifique des différents segments de marge en 2D. D'autre part, l'association de l'interprétation de ces lignes à des données de forages (*annexe 2 : les données de forage*) aux approches géophysiques présentées ci-après m'a permis de déterminer l'évolution des domaines de rift en mer dans le Golfe de Gascogne et le bassin de Parentis. La plupart des interprétations utilisées et faites dans le cadre de cette thèse sont présentées en annexe, seules certaines de ces lignes sont présentées dans la partie des résultats.

4.1.3. Approches quantitatives (Annexe 4)

Des techniques d'analyses quantitatives ont été utilisées pour caractériser les domaines de rift en 2D et en 3D. Ces méthodes ont été mises en œuvre en collaboration avec Nick Kusznir (Université de Liverpool) et incluent des techniques d'inversion gravimétrique et des analyses des anomalies résiduelles de subsidence. Le principe de ces méthodes est présenté dans l'annexe 4. L'association de ces différentes techniques m'a permis de caractériser quantitativement les domaines de rift et notamment l'étendue du domaine de manteau exhumé. Les résultats obtenus ne sont exposés qu'en partie dans les chapitres 1 et 2 et leur synthèse ne reflète donc que partiellement le travail effectué en amont, et en particulier en ce qui concerne le choix des paramètres et son influence sur les différents modèles testés. De plus une compilation des résultats de ces méthodes est proposée pour les différentes interprétations sismiques présentées (*annexe 4 : analyses quantitatives*).

4.2. Organisation de la thèse

Cette introduction, est essentiellement consacrée à la présentation des thématiques abordées et questions majeures associées. Ainsi les principaux résultats de travaux antérieurs menés dans le chantier considéré ont été synthétisés.

La seconde partie de la thèse présente les principaux résultats issus du travail de thèse, exposés sous forme d'articles en préparation ou soumis.

Le chapitre 1 correspond à un article soumis dans un volume spécial de la Société Géologique de Londres intitulé : «Sedimentary Basins and Crustal Processes at Continental Margins: From Modern Hyper-extended Margins to Deformed Ancient Analogues». Cet article intitulé: "*Characterizing and identifying structural domains at rifted continental margins: application to the Bay of Biscay margins and its western Pyrenean fossil remnants*" développe une nouvelle approche terre-mer à partir d'exemple provenant du Golfe de Gascogne et des Pyrénées.

Le chapitre 2 est un article en préparation pour soumission au journal *Tectonics* et intitulé : "*Formation and deformation of hyperextended rift systems: insights from the mapping of the Bay of Biscay–Pyrenean rift system*". Deux aspects principaux y sont développés. Dans un premier temps l'approche terre-mer précédemment développée est appliquée à l'ensemble du système Golfe de Gascogne–Pyrénées afin de caractériser d'évolution spatiale et temporelle des différents systèmes de rift soumis aux processus d'amincissement extrême. Ensuite, la carte illustrant la répartition des domaines de rift sur l'ensemble du système est utilisée pour discuter le rôle de l'architecture extensive sur la formation et l'architecture finale des orogènes.

Le chapitre 3 est un article en préparation pour *Geology*, intitulé : "*Formation and deformation of the Bay of Biscay–Pyrenean domain: implication for nature, kinematics and timing of the Iberian–European plate boundary*". L'évolution de la déformation extensive et compressive est discutée à l'échelle de la limite entre les plaques Ibérie et Europe.

Une troisième partie, sous forme de discussion synthétise les principaux résultats et apporte des éléments de réponse aux thématiques et questions posées dans l'introduction.

-Comment peut-on caractériser et identifier des domaines de rift en mer et dans leurs analogues fossiles à terre?

-Quelle est l'évolution tectonique, spatiale et temporelle des systèmes de rift soumis à un amincissement extrême dans le contexte des marges passives peu magmatique?

- Quel est le rôle de l'hyper-extension lors des processus de réactivation et de formation des orogènes?

- Quelle est l'origine de l'évolution de l'architecture à travers le système ?

- Quelle est la nature de la limite de plaque entre Ibérie et Europe ?

Une conclusion générale vient ensuite clore ce travail et présente de nouvelles perspectives de recherches.

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CHAPITRE 1

Le premier chapitre présente une nouvelle approche terre-mer couplant des données de géophysique marine issues de marges actuelles et de géologie de terrain provenant de reliques fossiles préservées dans les chaînes de montagne. Cette approche a pu être développée grâce à la richesse des données du Golfe de Gascogne (sismiques et forages) et à la présence d'analogues fossiles du même système dans les Pyrénées. Ce chapitre a pour objectif de définir des domaines de rifts comparables dans les marges passives actuelles et leurs analogues fossiles.

Pour les exemples en mer (marges des Entrées de la Manche et bassin de Parentis), les interprétations sismiques proposées sont converties en profondeur et accompagnées d'analyses quantitatives (*inversion gravimétrique et backstripping*). Plusieurs éléments diagnostics sont ainsi identifiés pour caractériser les exemples en mer : *l'espace d'accommodation, l'amincissement crustal et lithosphérique* ainsi que deux types de *domaines extensifs* (distingués par l'architecture des bassins associés).

L'étude des analogues fossiles (bassin de Mauléon) repose sur l'analyse structurale du bassin qui permet de délimiter des unités tectoniques cohérentes et de déterminer les différentes étapes de déformation. L'identification d'affleurements clefs dans ces différentes unités permet de décrire la *nature des sédiments* et du *socle* ainsi que les *structures* qui forment les restes fossiles des domaines de marge.

Cette analyse qualitative et quantitative d'observations à terre et en mer fournit des éléments de diagnostic essentiels pour identifier 5 domaines distincts dans les marges passives actuelles et fossiles : les domaines proximaux, de necking, hyper-amincis, de manteau exhumé et océaniques. Cette nouvelle approche peut ainsi être utilisée pour proposer des analogies entre les observations faites à terre et en mer. A terre, les reliques fossiles peuvent être replacées dans un contexte de marge passive actuelle et les analyses quantitatives permettent de restaurer l'architecture grande échelle. Les observations de terrain permettent de suggérer la nature du socle, des sédiments et des structures de rift pour les interprétations sismique en mer.

Les analyses quantitatives (*conversion profondeur, inversion gravimétrique et backstripping*) proposées dans cette partie du travail ont été effectuées au cours de plusieurs séjours à l'Université de Liverpool (collaboration avec Nick Kusznir). La plupart des affleurements présentés dans ce chapitre sont basées sur des travaux antérieurs, l'analyse structurale du bassin repose sur un travail de synthèse et de cartographie mené au cours de cette thèse.

Ce chapitre a été soumis dans un volume spécial de la Société Géologique de Londres (*The Geological Society Special Publication*), intitulé « *Sedimentary Basins and Crustal Processes at Continental Margins: From Modern Hyper-extended Margins to Deformed Ancient Analogues* ».

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Characterizing and identifying structural domains at rifted continental margins: application to the Bay of Biscay margins and its western Pyrenean fossil remnants

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Abstract

We use the Bay of Biscay and Western Pyrenees as a natural laboratory to develop and apply an innovative approach to characterize and identify distinctive rifted margin domains in offshore and onshore settings. The Bay of Biscay and Western Pyrenees offer the unique possibility to have access to seismically imaged, drilled and exposed parts of one and the same hyperextended rift margin system.

Offshore, we use a gravity inversion technique and flexural backstripping combined with seismic interpretation to provide quantitative estimates of accommodation space, crustal thickness and lithosphere thinning. Onshore, we focus on key outcrops preserving remnants of the former rift domain to describe the nature of sediment and basement rocks and of their interface. This qualitative and quantitative characterisation provides the essential diagnostic elements for the identification of five distinct domains at magma-poor rifted margins and their fossil analogues. We name these 5 domains proximal, necking, hyperthinned, exhumed mantle and oceanic.

This new approach can be used to reconcile offshore and onshore observations and aid interpretation especially when only local observations are available. Onshore remnants can be placed in an offshore rifted margin context, enabling the prediction of first order crustal architecture. For the interpretation of offshore seismic reflection sections, geological insights on rift structures and basement nature can be suggested based on onshore analogies.

1. Introduction

Knowledge on magma-poor rifted margin structure has significantly evolved over the past decades. Present-day models of continental rifted margin include the existence of hyperextended rifted domains made of extremely thinned continental crust and/or exhumed mantle, as described at present-day rifted margins (e.g. Reston & Manatschal 2011; Lundin & Doré 2011 and reference therein) and in onshore fossil analogues preserved in collisional orogens (e.g. Alps: Manatschal 2004; Mohn *et al.* 2010; 2012; Masini *et al.* 2011; 2012; Pyrenees: Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Caledonides: Anderson *et al.* 2012). Both marine and onshore observations offer complementary datasets, however, the different scale and resolution of observations restrain straightforward correlations and most studies are either focused on offshore or onshore examples.

As only few drill-holes enable the direct exploration of distal parts of offshore continental rifted margins, their investigation mainly relies on indirect approaches such as the interpretation of reflection and refraction seismic sections (e.g. Osmundsen & Ebbings 2008; Péron-Pinvidic *et al.* 2013; Sutra *et al.* 2013; Franke 2013), potential field methods or geophysical quantitative techniques (e.g. Pérez-Gussinyé *et al.* 2003; Reston 2009; Roberts *et al.* 2013). In contrast, fossil analogues of rifted margins preserved in collisional orogens offer a direct access to observations on the nature of sediments, basement and of their interface. Remnants of ancient rifted margins preserving primary relationships provide valuable insights on the interaction between tectonic and sedimentary processes but also on the subsidence history during and after rifting (e.g. Mohn *et al.* 2010; Masini *et al.* 2013).

In this study, we aim to characterize distinctive structural domains at continental rifted margins in order to propose qualitative and quantitative criteria to identify and interpret hyperextended domains onshore and offshore. We use the Bay of Biscay and Western Pyrenees to develop and apply this geological/geophysical approach. This natural laboratory preserves different evolutionary stages of a hyperextended rifted margin system that are seismically imaged and drilled, as well as exposed parts of the former rift system, due to the subsequent inversion and integration in the Pyrenees.

This combined approach can be used to reconcile offshore and onshore observations on continental rifted margins. First order relationships can be established between accommodation space creation, lithosphere thinning, extensional modes and field observations on the nature of sedimentary and basement rocks for the different domains. As a consequence, this approach can be used to aid interpretations of local observations both onshore and offshore. Onshore fossil remnants can be placed back in the context of the whole rifted margin predicting the first order crustal architecture. Onshore analogies can be used to suggest geological insights on rift structure and basement nature for offshore seismic interpretations.

Finally, we believe that this approach may also be used, if limitations are respected, in a more general way to characterize, identify and interpret hyperextended rift systems, either buried under thick sedimentary sequences at present-day margins or disrupted in collisional orogens.

2. Rifted margins: terminologies, approaches and observations

In spite of the wide spectrum of rift architectures observed at continental rifted margins, first-order observations can be used to describe, characterize and sub-divide them into different domains (Reston 2009, Péron-Pinvidic *et al*. 2013). The identification of different domains is either based on seismic studies (e.g. Dean *et al.* 2000; Minshull 2009, Osmundsen & Ebbings 2008; Péron-Pinvidic *et al*. 2013, Sutra *et al.* 2013), quantitative approaches (e. g. Reston, 2009) or rheological considerations on how the crust and lithosphere deform (Pérez-Gussinyé *et al.* 2003; Sutra & Manatschal 2012). As a consequence, the terminology and definition of domains depends on the approach, as well as on the resolution and scale of observations (fig1). In fact, most studies of present-day rifted margins rely on geophysical approaches (e.g. seismic reflection or refraction and potential field methods) that give large-scale images of the crustal structure but no direct insights on the nature of the rocks and structures that form the margins. In contrast, the investigation of remnants of fossils margins exposed in collisional orogens offers direct access to outcrops corresponding to "high resolution" pixels of the former margin section. However, since these outcropping remnants represent disconnected samples from the former rifted margin, the relation between them is often complex, difficult to determine and they can only be interpreted if they can be restored back to their initial position (e.g. Jammes *et al.* 2009; Mohn *et al.* 2010, fig 1). Drill-hole observations are the only dataset that offers a local in-situ geo-referenced 1D access to detailed geological information within a present-day rift section, enabling to bridge direct outcrop observations with indirect geophysical observations. The Bay of Biscay and West Pyrenean rift system offers the unique possibility to have access to seismically imaged, drilled and exposed parts of one and the same hyper-extended rift system. Using this example we aim to demonstrate how onshore (high resolution but discontinuous) and offshore (low resolution but continuous) observations can be combined and used to produce the first-order definitions of genetic rift domains proposed in this work (proximal, necking, hyperthinned, exhumed mantle and oceanic domains, fig 1).

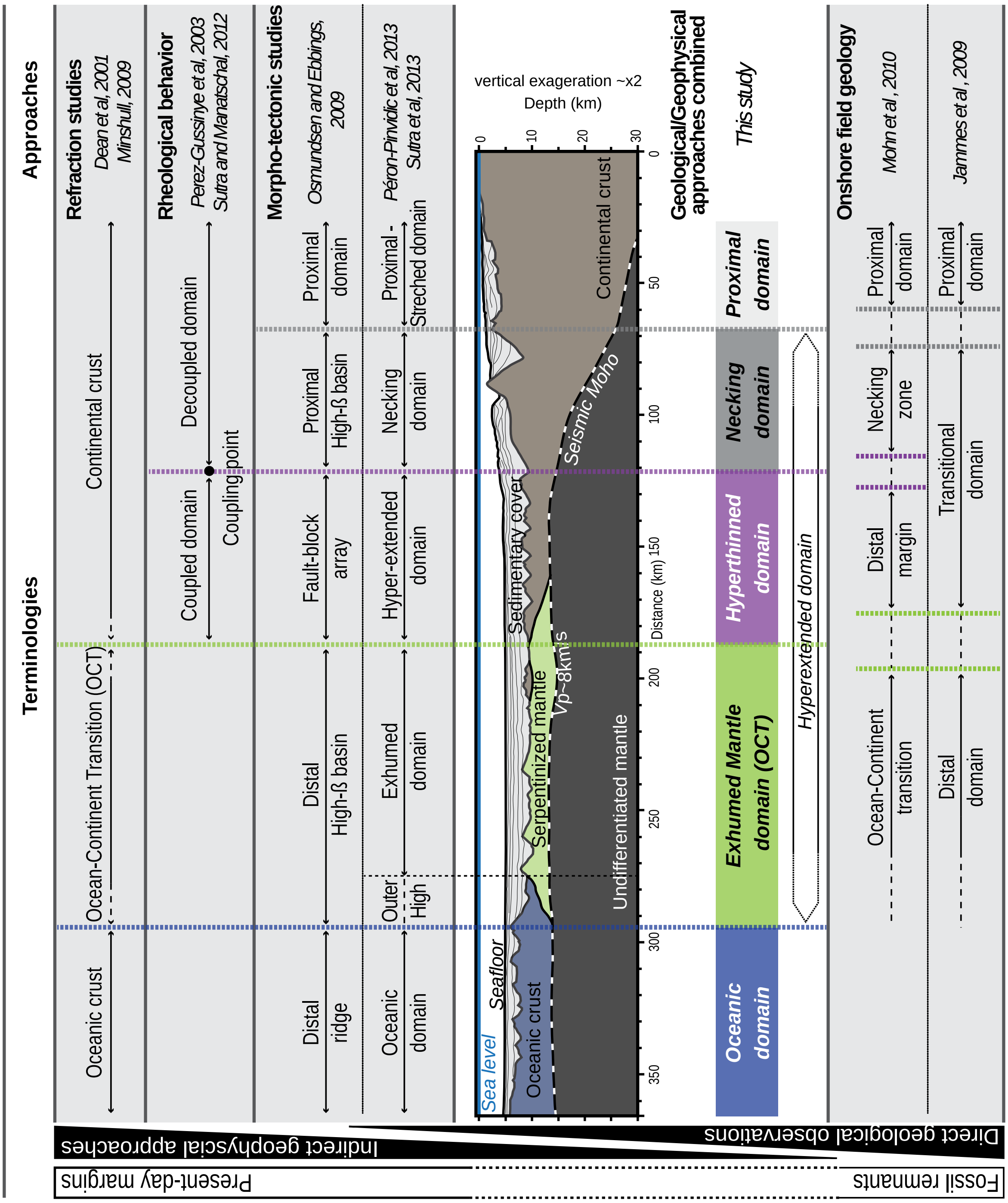


Fig.1: *Simplified architecture of a hyperextended magma-poor rifted margin (based on the IAM 5 refraction line from the West Iberia margin, Afilhado et al. 2008) and terminology adopted in this study (central part). A comparison with terminologies from other recent studies relying either on geophysical approaches (upper part of the figure) or geological field observations (lower part of the figure) is shown.*

2.1. Indirect geophysical approach

Geophysical studies such as seismic reflection and refraction surveys or potential field data provide first order continuous images of rift systems that are based on the mapping of physical properties of rocks forming continental rifted margins. Since different rocks can have similar densities and/or velocities, the interpretation of geophysical data is non-unique. However, first order interfaces, juxtaposing materials with different densities and/or velocities, such as sea-level, sea-bed/top of sedimentary infill, base of sediments/acoustic basement and seismic basement/mantle rocks velocities (seismic Moho) can be well imaged in seismic sections (fig 2a). These interfaces can also be used to define the total accommodation space (surface between sea-level and top basement, as defined in fig 2a) and crustal thickness (surface between top basement and Moho). Furthermore, since both are related to extension and subsidence, they also provide input parameters that can be used for analytical methods to link crustal thinning and subsidence history (creation of accommodation space) (e.g. Kusznir & Karner 2007; Roberts *et al.* 2013). Although there are limitations to these calculations, due to the restricted knowledge of some critical lithospheric processes such as dynamic topography or the detailed thermal and density structure of the lithosphere, these methods can quantify the evolution of accommodation space and crustal thinning during rifting and can therefore describe the evolution of rifting in time and space.

2.1.1. Accommodation space

Considering a section through a present-day thermally equilibrated rifted margin, the most straightforward observation is to consider the evolution of potential accommodation space (between sea level and seafloor) and total accommodation space (between sea level and top basement), as shown in fig 2. At a thermally equilibrated magma-poor rifted margin accommodation space commonly increases oceanwards as continental crust thins and is controlled, to first order, by crustal thickness, post-breakup thermal subsidence and the thickness of the sedimentary cover (contributing to compaction and flexural loading). Second order controls such as eustatic sea-level variations and mantle processes (dynamic topography) become important when attempting to restore the palaeo-bathymetry at breakup time.

2.1.2. Crustal thinning

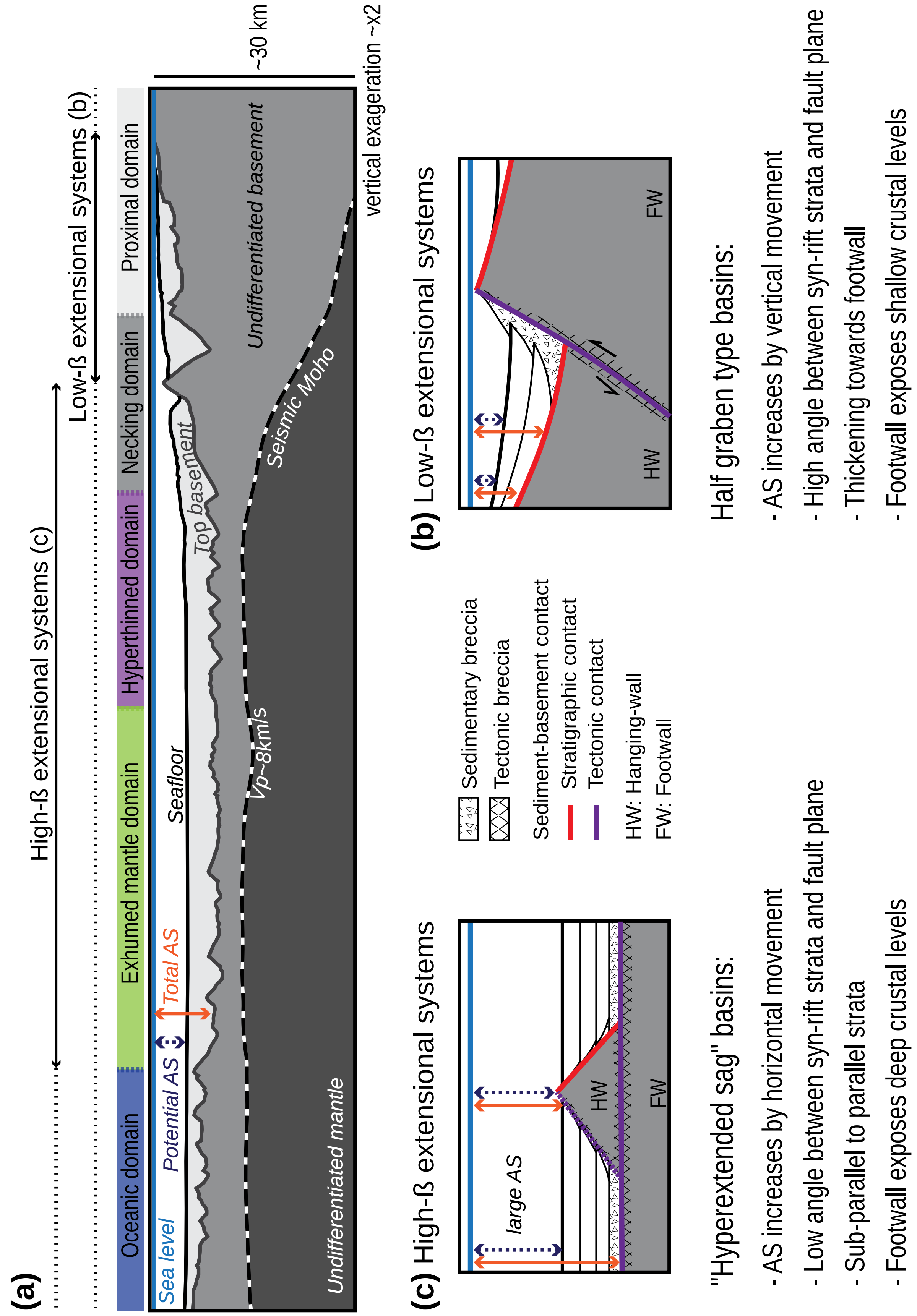
Crustal thickness variations along rifted continental margins can be quantified using the lithosphere stretching factor β (McKenzie 1978), defined for the continental lithosphere as the ratio of the initial continental crust (or lithosphere) thickness to the present-day thickness of the continental crust. As a consequence, for oceanic lithosphere (i.e. in oceanic domains), β-factor values tend to infinity. However, the occurrence of magmatic addition during the rift complicates the quantification and interpretation of the stretching factor since crustal thickness variations are not only function of thinning processes, but also depend on the magmatic additions. In this paper, we explore a magma-poor system, where magmatic additions during rifting are of subordinate importance (e.g. Montadert *et al.* 1979b; de Charpal *et al.* 1978; Avedik *et al.* 1982; Le Pichon & Barbier 1987; Thinon *et al.* 2003). Hereafter, crustal thinning is represented by the thinning factor (γ= 1-1/β), which is more convenient to represent graphically, since it only varies between zero (no crustal thinning) and one (complete thinning of the continental crust).

2.1.3. Low- vs. high-ß extensional systems (Fig. 2)

Two different extensional systems, referred to as low and high-ß systems, can be distinguished (fig 2), based on basin architecture and associated extensional structures (Wilson *et al.* 2001). Low-β extensional systems correspond to classical rift structures. They are characterized by high-angle normal faults and tilted blocks delimitating graben and half-graben type basins filled by wedge shaped sedimentary sequences. Low-ß systems can only accommodate small amounts of extension and therefore are frequently found over weakly thinned crust (i.e. in proximal margins, fig 2).

In contrast, high-β extensional systems are associated with long-offset extensional faults; also referred to as top basement detachment faults (Hölker *et al.* 2003) or large-offset normal faults (e.g. Wernicke 1985; Lister *et al.* 1986; Brady *et al.* 2000). In these systems, faults and their underlying footwall can be exhumed to the seafloor, leading to the creation of new surfaces tectonically exhumed to the seafloor (sometimes referred to as “new real estate”). This

***Fig. 2**: (a) Architecture of a hyperextended magma-poor rifted margin (based on the IAM 5 refraction line from the West Iberia margin, Afilhado et al. 2008) showing first order interfaces (seafloor, top basement and Moho), and the potential and total accommodation space (AS). (b-c) Architecture and characteristics of half graben and hyperextended sag basins also referred to as low and high-ß extensional systems. Note the difference in potential and total accommodation space between the two types of basins and the nature of sediment-basement contact.*


(a)
High-ß extensional systems (c)
Low-ß extensional systems (b)
Oceanic domain
Exhumed mantle domain
Hyperthinned domain
Necking domain
Proximal domain
~30 km
Sea level
Potential AS
Total AS
Seafloor
Top basement
Vp~8km/s
Seismic Moho
Undifferentiated basement
Undifferentiated mantle
vertical exageration ~x2
(c) High-ß extensional systems
large AS
HW
FW
(b) Low-ß extensional systems
HW
FW
Sedimentary breccia
Tectonic breccia
Sediment-basement contact
Stratigraphic contact
Tectonic contact
HW: Hanging-wall
FW: Footwall
"Hyperextended sag" basins:
- AS increases by horizontal movement
- Low angle between syn-rift strata and fault plane
- Sub-parallel to parallel strata
- Footwall exposes deep crustal levels
Half graben type basins:
- AS increases by vertical movement
- High angle between syn-rift strata and fault plane
- Thickening towards footwall
- Footwall exposes shallow crustal levels

type of faults is difficult to observe on seismic sections, because in contrast to high-angle faults, they form relatively smooth fault related topography. Since these faults can accommodate large displacements and thin the crust, they are interpreted to floor large sedimentary basins within the hyperextended distal parts of magma-poor rifted margin (fig 2). These "hyperextended sag" basins are characterized by overlying sub-parallel sedimentary sequences and a low angle on-lap of the strata onto basement (fig 2). This type of basins is described from several deep-water rifted margins e.g the Mid Norwegian margin (High-β basins, Osmundsen & Ebbing 2008), the Angola and Iberia margin (sag basins, Unternehr *et al.* 2010; Péron-Pinvidic *et al*. 2007; 2013).

2.2. Direct geological approach

The discovery of dislocated fragments of ancient rifted margins in collisional orogens gives access to detailed geological observations and unlimited sampling that enable to characterize the nature and age of sediments, type of the basement rocks as well as the nature of the contact between basement rocks and sediments. The combination of sedimentological, petrological and structural observations provides information on depositional environment, kinematics and timing of faulting, thermal and isostatic history of a rift system. However, well-exposed and preserved outcrops of former rift systems are only locally preserved due to reactivation of the former rift structures during their emplacement in the orogenic system. Therefore, the restoration of these "remnants" of former rift systems back into their original position within the margin is often difficult and related to large uncertainties. Studies of remnants derived from the former Jurassic Tethys margins, today exposed in the Alps, provided detailed descriptions of rift structures and their link to sediments and underlying basement rocks. Because the alpine fossil remnants show many similarities with structures that were drilled and seismically imaged along the Western Iberia margin, they are considered as "analogues" of low- and high-ß extensional rift systems (fig 2, e.g. Wilson *et al.* 2001; Manatschal 2004). While low-ß systems can be well defined from seismic data, the high-ß systems and in particular the exhumation faults related to these systems are only observable at an outcrop scale. Therefore, being able to resolve the nature of top basement (fault or stratigraphic contract) and the type of hanging-wall and footwall rocks is a key in interpreting seismic data. In offshore systems, this information can only be established by drill-hole data.

2.2.1. Nature of sediments and creation of accommodation space

In order to quantify the total accommodation space created during extension, the palaeo-bathymetry and the decompacted stratigraphic thickness need to be determined. In particular for deep-water environments, the estimation of the palaeo-bathymetry is associated to large uncertainties. Therefore, field data can only qualitatively determine the formation of accommodation space, either by the observation of aggradation and formation of thick sedimentary sequences or by the deepening of the depositional environment.

In proximal margins, characterized by low-ß extensional systems, the crust shows little thinning and hence accommodation space and/or palaeo-water depth is small or zero, in the latter case resulting in sub-aerial exposure and basement erosion. As a consequence, the occurrence of shallow marine post-rift sediments onlapping directly onto basement is characteristic for this domain. Locally, syn-rift sediment thickness can be more important but restricted to half graben type basins (fig 2b).

High-ß extensional systems are associated with distal margins (Wilson *et al.* 2001) or highly extended rift settings (Friedmann & Burbank 1995). Well-preserved remnants of high-ß extensional systems are described from the Alps and Pyrenees (i.e.: supra-detachment basins in Masini *et al.* 2011). The complex architecture of these systems, including top basement detachment faults, break-away blocks and extensional allochthons (see Fig. 3 in Masini *et al.* 2013) are reflected in the sedimentary infill history that strongly depends on the sedimentary supply (Masini 2011). In sedimentary starved systems, these basins are filled by deep-water turbidites interleaved with pelagic or hemi-pelagic material, whereas in sediment-rich systems, these basins show thick aggradational sequences. In cases where the base level of these basins is not anymore controlled by the global sea-level, these basins can be air filled basins, often associated with the formation of evaporates.

2.2.2. Nature of top basement

A key observation to distinguish between low and high-β extensional settings is the nature of basement rocks and of its contact with the overlying sedimentary deposits. This contact can either be a stratigraphic contact, a high-angle fault or a tectonically exhumed basement, in which case the top of the basement is either an erosion surface or an exhumed fault surface (Masini *et al.* 2011; fig 2b/c). The sharp topography observed along low-ß extensional systems is due to the small offset of the top basement along high-angle normal faults. Since the faults can only accommodate little extension they cannot exhume deep pre-rift lower crustal levels or mantle rocks to the surface. Hence, in proximal domains, low-ß extensional systems dominantly preserve pre-rift upper crust in rotated fault blocks (fig 2).

Top basement detachment faults and associated high-ß extensional systems can be best recognized in outcrops. These faults can exhume deep crustal levels and mantle rocks while thinning the crust. The presence of exhumed sub-continental mantle at the seafloor suggests an entire removal of the continental crust corresponding to a complete crustal thinning.

2.2.3. Geometry and kinematics of extensional systems

While on seismic sections the recognition of the large-scale architecture of extensional systems mainly relies on the study of the infill stratigraphic geometry, in orogens, the overall architecture of rift basins is usually not preserved. However, insights may come from the study of the relation between fault geometry and the related syn-tectonic sediments. Low-β settings are classically recognized by wedge shaped syn-tectonic sediments (Wilson *et al*. 2001) that can be interbedded with footwall-derived breccia along fault scarps. The syn-tectonic sediments are typically deposited on the tilted hanging-wall and the contact with the normal fault is at a high angle (fig 2b). In high-ß settings, the relationships between the syn-tectonic sediments and the fault surface are very different (fig 2c, Wilson *et al*. 2001). Top basement detachment faults are associated with a characteristic sequence of rocks that include cataclastic basement and gouges, overlain by tectono-sedimentary breccias; grading upwards into sedimentary breccias and turbidites reworking hanging-wall and footwall-derived material (Masini *et al.* 2012). The occurrence of clasts derived from pre-rift lower crust and mantle in syn-rift breccias provide an indirect hint for the existence of exhumation faults in hyperextended domains. Another characteristic of top-basement detachment faults is the occurrence of hanging-wall derived extensional allochthons overlying the fault surface. These blocks can range in size from tens of kilometres to tens of meters (for examples see Péron-Pinvidic & Manatschal 2010).

2.3. Drill-hole data: the link between direct and indirect approaches

Drill-hole data enable the linkage of rock-derived data (physical properties, ages, and kinematics) with seismic data. In contrast to outcrops that sample remnants of former margins that are disrupted and often out of context, drill-hole data are localized and can sample well-defined parts of the margins. At present, most observations based on drill-hole data come from the proximal margins, while data coming from the distal margins are limited to the conjugate Iberian and Newfoundland margins, which at present is the only margin system where drill holes penetrated the hyper-extended basement. The ODP drill hole data (Legs 103, 149, 173 and 210) from the Iberian/Newfoundland margins enabled to validate the link between the deep structures seismically imaged along these margins with outcrops exposed in the Alps (e.g. Wilson *et al.* 2001; Manatschal 2004). The recognition of analogies between these two margins enabled to link the outcropping fossil remnants with seismic interpretations. However, since these margins formed independently from each other and their post-rift evolution is very different, the integration of the two datasets, beyond using them as analogues, is not possible.

3. The Bay of Biscay and Western Pyrenees natural laboratory

In this paper we use the Bay of Biscay/Western-Pyrenees as a natural laboratory to show how indirect geophysical and direct field observations can be integrated into a coherent and consistent approach to characterize and interpret hyperextended rifts systems. The Bay of Biscay/Western-Pyrenees preserves, along one and the same system, differenat evolutionary stages of a rift and compressional system (fig 3). Thanks to decades of exploration, the access to critical drill-hole observations is possible making this a unique natural laboratory to link on- and offshore observations.

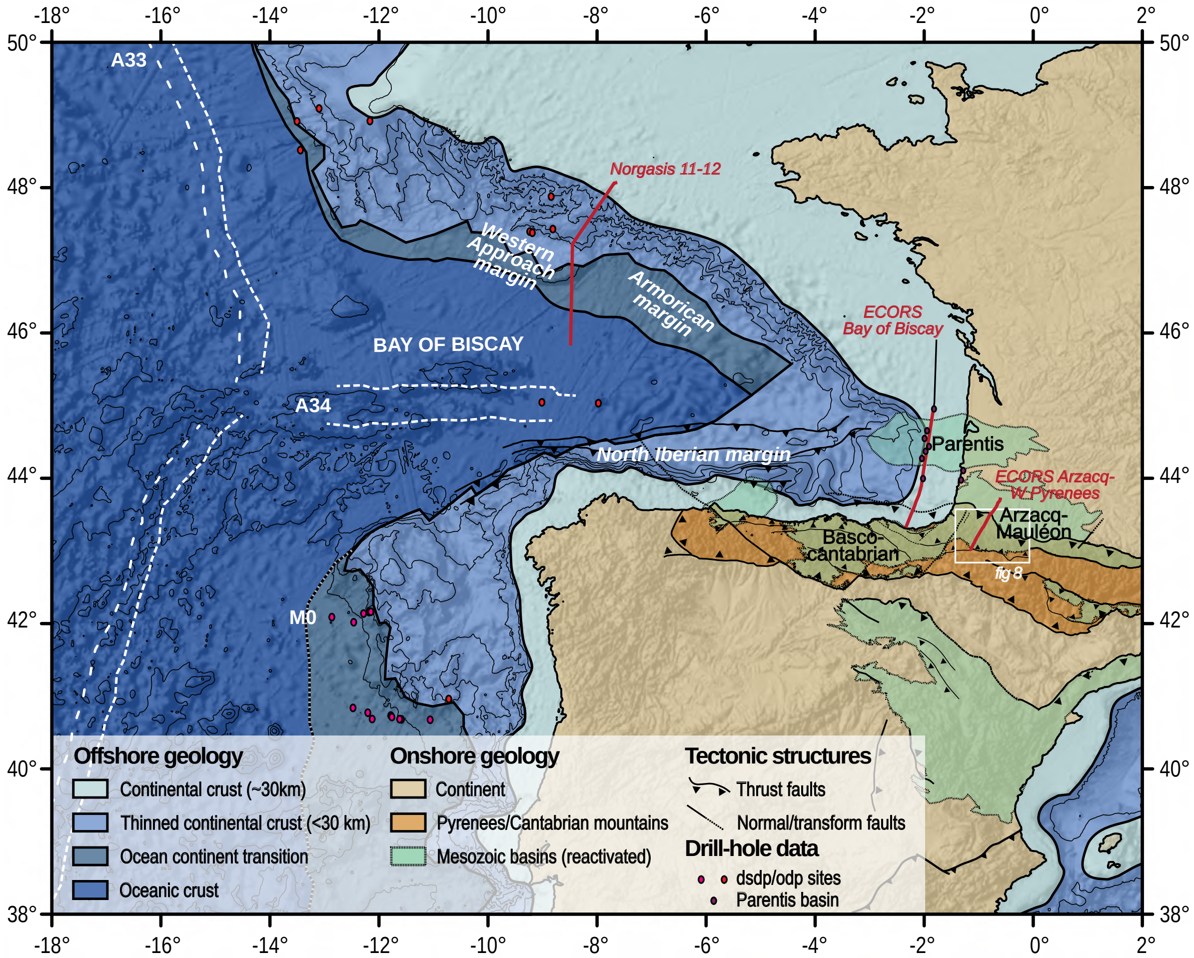


***Fig. 3**: Map of the Bay of Biscay and Pyrenees showing the major tectonic structures and extensional domains (zonation established after Thinon 1999 and Jammes et al. 2009) and location of the sections presented in this study. Magnetic anomalies are based on Sibuet et al. 2004*

3.1. General Setting

The Bay of Biscay, located between France and Spain, is characterized offshore by an embryonic oceanic basin, bounded to the North by the Western Approach and Armorican margin and to the South by the North Iberian/Cantabrian margin. At the eastern termination, several Mesozoic rift basins preserving geophysical and geological evidence of extreme crustal thinning are identified both offshore (e.g. the Parentis basin: Pinet *et al.* 1987; Bois & Garriel 1994; Tomassino & Marillier 1997; Jammes *et al.* 2010 a,b,c) and onshore (e.g. the Arzacq Mauléon basin: Daignières *et al.* 1994; Grandjean 1992; Jammes *et al* 2009; Lagabrielle *et al.* 2010; and the Basque-Cantabrian basin: Pedreira *et al.* 2007).

The present-day structuration of the region results from the succession of several extensive and compressive tectonic events initiating with the Variscan orogen and early Permian orogenic collapse (Burg *et al.* 1994a, 1994b), followed by several rift phases, Triassic to Lower Cretaceous in age. The Late Cretaceous northward movement of the African plate resulted in the partial closure of the Bay of Biscay - Pyrenean domain.

As this work will be mainly focused on the Early Cretaceous hyperextensional event leading to oceanic crust formation at Aptian-Albian time in the Bay of Biscay (Montadert *et al.*. 1979b; Boillot 1984), we will subdivide the sedimentary record in the study area into pre-, syn-, and post-hyperextension sedimentary sequences.

3.2. Plate kinematic context, Cretaceous hyperextension and Pyrenean compression

The paleogeographic and plate kinematic evolution related to the opening of the Bay of Biscay remains at present strongly debated. These large uncertainties can be related to the lack of reliable magnetic anomalies (including the M-series and the following magnetic quiet zone) indicative of break up time in the Iberia-Newfoundland margins (Bronner *et al.* 2011) and the Bay of Biscay (Norton *et al.* 2007). Furthermore, the discovery of hyper-extended crust and exhumed mantle at the Iberia-Newfoundland margins over hundreds of kilometres continentwards of the first unambiguous magnetic anomaly, related to breakup (for a review see Tucholke *et al.* 2007 or Péron-Pinvidic & Manatschal 2009) is at present not taken into account by most restorations (e.g. Olivet 1996). As a result, several geodynamic models are suggested for the Late Jurassic to Late Cretaceous evolution of the Bay of Biscay including a back arc setting (e.g. Sibuet *et al.* 2004; Vissers & Meijer 2012) or a transtensional setting forming pull-apart basins within the Pyrenean domain (Choukroune & Mattauer 1978). Based on the mapping of the Iberia-Newfoundland margins (e.g. Péron-Pinvidic & Manatschal 2009; Sutra *et al.* 2013) and combining it with field observations from the Mauléon and Parentis basins, Jammes *et al.* (2009) and (2010a) proposed an alternative plate kinematic scenario. Their model, similar to the one already proposed by Williams (1975), suggests that the main rifting episode is linked with a counter clock wise rotation of Iberia during Aptian time (Gong *et al.* 2008), overprinting a Late Jurassic to Early Aptian sinistral strike-slip or transtensional event.

The kinematics of the area is reasonably well constrained after the magnetic quiet period of the Cretaceous with the identification of the magnetic anomaly 34 (about 84 Ma, Campanian-Santonian) in the Bay of Biscay (e. g. Rosenbaum *et al.* 2002). First evidences of compressional deformation are recorded in Santonian to Campanian time in the Pyrenees (Garrido-Megías & Ríos 1972; McClay *et al.* 2004) and are associated with heterogeneous and weak reactivations in the Northern and Eastern part of the Bay of Biscay (Thinon *et al.* 2001; Pinet *et al.* 1987). After Eocene time, the study area almost acquired its present-day structure. The Southern Cantabrian margin and the Basque-Cantabrian and Mauléon basins are reactivated and integrated into the Pyrenean orogen.

3.3. A natural laboratory to observe, characterize and interpret hyperextensional systems

The Bay of Biscay and Western Pyrenees do not only preserve different evolutionary stages of a rift system, including hyperthinned crust, exhumed mantle and oceanic domains (terminology defined in fig 1), but also show different steps of reactivation along one and the same system. In the following, we will consider three sections preserving different genetic and reactivation steps. A “margin” stage is investigated using the Norgasis 11-12 seismic section across the Western Approach margin (Thinon 1999, Avedik *et al.* 1996, location fig 3). The “hyperextended” stage is imaged on the ECORS Bay of Biscay profile across the Parentis basin (Pinet *et al.* 1987, location fig 3). Finally, the integration of the Mauléon rift basin into the Western Pyrenees enables to explore the reactivated stage of a hyperextended rift system (Jammes *et al*. 2009, location fig 3). The three sections do not only represent different stages in a Wilson cycle, but also result from the interpretation of different datasets with different resolutions and observational scales.

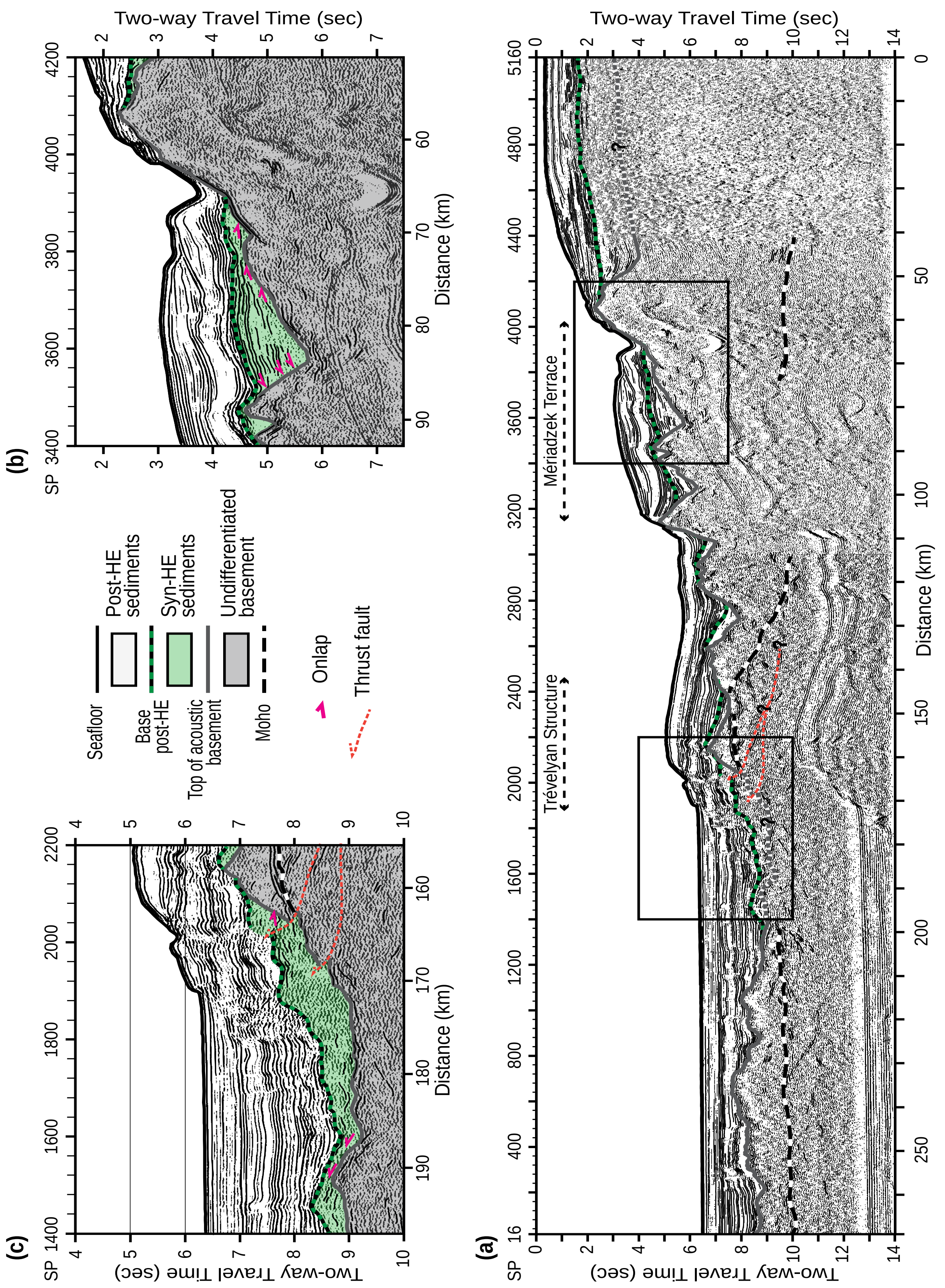
4. The margin stage: the Western Approach margin

The Western Approach margin corresponds to the north-western segment of the Bay of Biscay (fig 3). It shows the typical "building blocks" of a hyperextended rifted margin, including indirect hints supporting the existence of exhumed mantle domains (Thinon 1999; Thinon *et al*. 2003; Bullock 2004; Bullock & Minshull 2005). Only few drill-holes penetrated the sedimentary cover (leg 48, Montadert & Roberts 1979a), enabling some seismic stratigraphic correlations across the margin to be proposed and to extend them towards the adjacent Armorican margin (Thinon *et al*. 2002). Hereafter, we will focus on the Norgasis 11-12 sections across the Western Approach margin, corresponding to consecutive seismic sections acquired in 1994 by Ifremer during the Norgasis campaign (Avedik *et al.* 1996; Thinon 1999). In the following, we will mainly rely on the mapping of first order interfaces (seafloor, top basement and Moho). The mapping of these interfaces allows us to define low-ß and high-ß extensional system (Wilson *et al*. 2001), but also to quantify crustal thickness and accommodation space and to describe and characterize genetic domains across the margin.

4.1. Seismic interpretation: first order interfaces

First order interfaces in seismic sections correspond to interfaces between materials with different physical properties, which are: 1) the water/sediment interface (seafloor topography), 2) the interface between sediments and basement corresponding to seismic basement and 3) the crust/mantle boundary (Moho). The first limit can be well observed in the time migrated Norgasis 11-12 sections (fig 4). Top basement is also rather well imaged along most of the section and corresponds to the interface between the base of syn-hyperextensional sediments (Berriasien to Aptian, Thinon *et al.* 2002) and the acoustic basement, characterized by strong low frequency reflectors either corresponding to older well cemented sediments, meta-sediments or crystalline basement. However, in some places, top basement is not well defined and its interpretation is related to larger uncertainties either because of multiples (to the North, where the section cuts through the Western Approach basin, SP 5160-4400), or because of the lack of distinctive basement reflections (Southward of Trevelyan, SP 2000-1400). On the Norgasis 11-12 sections, deep low frequency sub-parallel reflections are observed in the seismic basement

Fig. 4*: (a) Interpretation of the seafloor, base post-hyperextension (HE), top basement and moho reflections on the Norgasis 11-12 seismic section (seismic sections from Thinon 1999). Zoom over hyperextended sag type basins characterizing high-β extensional settings on (b) the Meriadsek terrace and (c) South of the Trevelyan structure.*


(a)
(b)
(c)
SP
Two-way Travel Time (sec)
Distance (km)
Mériadzek Terrace
Trévelyan Structure
Seafloor
Post-HE sediments
Base post-HE
Syn-HE sediments
Top of acoustic basement
Undifferentiated basement
Moho
Onlap
Thrust fault

between 9 and 10 seconds (twtt). These reflections are highly discontinuous north and south of the Meriadsek terrace (SP 4400-3800; 3000-2600) but more continuous oceanwards, between SP 16-1400 at the southernmost termination of the section. In the Trevelyan area (SP ~2400-2000), where evidence of compressional deformation is observed, similar strong reflectors can be described but at shallower levels ~7.5 sec (twtt), due to the tectonic inversion (Thinon *et al*. 2001). A possible interpretation of these deep reflections is to relate them to the velocity contrasts between continental or oceanic crust and the underlying mantle rocks, corresponding to seismic Moho.

4.2. Stratigraphic and basin architecture

A detailed description of the morphology and stratigraphic architecture of the Western Approach margin was provided by Thinon (1999). In this study we focus on observations enabling the distinction between high- or low-β extensional systems (fig 2). Hence, only the base of post-hyperextensional sediments is discussed (Late Aptian, Thinon *et al*. 2002). The Western Approach margin is classically only interpreted as a succession of titled blocks (e.g. Avedik *et al.* 1979; Montadert *et al.* 1979b) characteristic of low-β extensional settings. However, based on the descriptions of Wilson *et al.* (2001), two high-β extensional domains can be observed on the Norgasis sections. Over the Meriadzek terrace almost sub-parallel syn-hyperextensional sedimentary sequences are either onlapping at a low angle the underlying basement (SP ~3800) or terminate towards a basement high (SP ~3600) delimiting a large sedimentary basin (fig 4b). The second one, more distal is located southward of Trevelyan. In this domain the seismic unit 3B, Early Aptian in age (Thinon *et al.* 2002) directly overlies the almost flat top-basement and shows a stratigraphic architecture similar to that of classical post-rift sediments. However, it is important to note that these sediments are older than the first sediments that overlie oceanic crust (Late Aptian to Albian in age, Thinon *et al.* 2002) (fig 4c). This “hyperextended sag” type architecture is indicative of high-β extensional settings that may be floored by top basement detachment faults (Wilson *et al.* 2001, Masini *et al.* 2012; Péron-Pinvidic *et al.* 2013), related to thinning and exhumation processes.

4.3. Estimations of crustal thickness and lithosphere thinning

In our study, we depth converted the first order interfaces defined on the Norgasis 11-12 section, which are: 1) seafloor, 2) base post-hyperextension and 3) top seismic basement. The depth conversion uses a constant velocity (1500 m/s) for water and a linear velocity model with depth for sediment thickness (Slotnick 1936; Japsen 1993). Moho depth, crustal thickness and continental lithosphere thinning factor (fig 5) have been determined using a gravity inversion technique incorporating a thermal gravity anomaly correction and a parameterization of decompression melting to predict volcanic addition (detailed scheme described by Greenhalgh & Kusznir 2007; Chappell & Kusznir, 2008; Alvey *et al.* 2008, parameters for this study in table 1). The data used in the gravity inversion are public domain free-air gravity (Sandwell & Smith 2009) and bathymetry (Smith & Sandwell 1997), the depth converted sediment thickness from the seismic profiles and oceanic isochrones (Muller *et al.* 1997). A compaction controlled density-depth relationship is assumed for sediment.

Two end-member results are presented in figure 5 a-b. A first one corresponds to a magma-starved model and a second one to a model with "normal" volcanic additions. The magma-starved solution (fig 5a) implies that no melt is generated during rifting and break-up; hence the crustal basement thickness derived from gravity inversion only represents continental crust. The solution computed with a "normal" volcanic addition (i.e. maximum oceanic crustal thickness of 7 km) (fig 5b) assumes that decompression melting occurs during rifting, producing magmatic additions to the extended continental crust once the lithosphere is thinned below a critical value (fig 3 in Chappell & Kusznir 2008). Therefore, the crustal basement thickness derived from gravity inversion represents both the residual continental crust and volcanic additions. We are aware that none of these two models completely represent the Norgasis 11-12 section, but they provide end-members. Indeed, considering gravity inversion results for the Norgasis sections (Thinon 1999), the Moho inflection at the termination of the residual continental crust (fig 5b) is compatible with a basement composed of serpentinized mantle.

Comparing the two thinning factor profiles (fig 5d) and thickness variations of the residual continental crust (fig 5c) of both models, three main trends can be observed delimiting different domains (fig 5). In the most proximal domain, the thinning factor is relatively low and constant (γ~0.2) in both cases, coincident with a limited thinning of the continental crust. Contrastingly, the initiation of the extreme thinning of the continental crust coincides with the high-β domain identified in fig 4 in the Meriadsek terrace. In this domain a strong increase in thinning values is observed that extends oceanwards until complete thinning of the crust is achieved (resulting in $\gamma \approx 1$ for the "normal" volcanic addition model). This domain corresponds to the beginning of the ocean continent transition, where mantle exhumation is proposed on the Western Approach margin (Thinon 1999; Thinon *et al.* 2003) and where another more distal high-β setting is identified (fig 4).

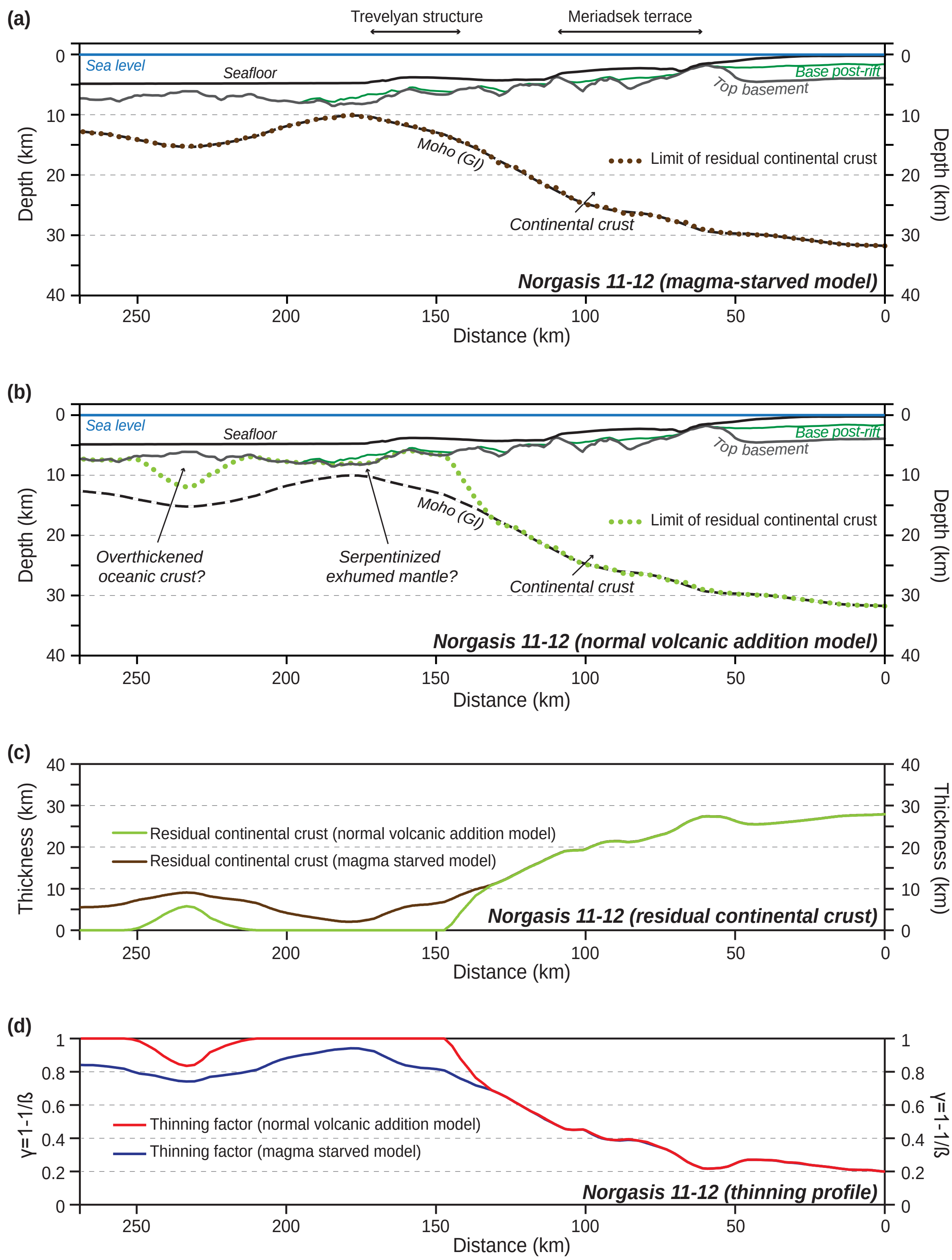


***Fig. 5**: (a) & (b) Crustal cross section along the Norgasis 11-12 profiles showing depth converted seafloor, base post-hyperextension and top basement. The Moho depth and residual continental crust limit are derived from gravity inversion results (GI) for (a) a "normal" volcanic addition model and for (b) a magma-starved model. (c) Residual continental crust and (d) thinning factor variations along the Norgasis 11-12 profiles are determined from gravity inversion for both "normal" volcanic addition and magma-starved models*

TABLE 1: Parameters used for gravity inversion

Parameters	Value and reference dataset
Reference crustal thickness	40 km
Break up age	110 Ma
Volcanic additions	
Standart thickness	7 km
Magma-starved	0 km

4.4. Quantification of accommodation space evolution

In thermally equilibrated magma-poor rifted margins, total and potential accommodation space are to first-order controlled by crustal thickness and sedimentary supply. The estimation of the potential and total accommodation space for the Norgasis 11-12 section relies only on the identification of seafloor and top basement from seismic reflection data (fig 6 a/b). In order to determine the sediment corrected post-rift bathymetry at a mature (thermally equilibrated) margin, 2D flexural isostatic backstripping and decompaction has been performed on the Norgasis 11-12 section. This post-rift sediment corrected bathymetry enables the evolution of accommodation space without the loading and compaction effects of post-rift sediments to be determined (fig 6c).

Flexural backstripping and reverse post-rift thermal subsidence modelling has also been applied to the section in order to predict bathymetry at the start of post-breakup subsidence. The distribution of reverse thermal subsidence is determined using the model of McKenzie (1978) and the thinning factor profile resulting from gravity inversion (using normal volcanic addition model, fig 5c); The methodology for flexural backstripping, decompaction and reverse thermal subsidence modelling is described in Kusznir *et al.* (1995) and Roberts *et al.* (1998). A breakup age of 110 Ma for the Bay of Biscay, corresponding to Late Aptian (Montadert *et al.* 1979b; Boillot 1984), is used to define the reverse thermal subsidence time. This age is assumed to represent the transition from tectonic dominated subsidence to thermal subsidence (parameters used in our models are presented in table 2). The modelled bathymetry at break-up time prior to thermal equilibration (fig 6c) enables the evaluation of the available syn- to post-rift accommodation space.

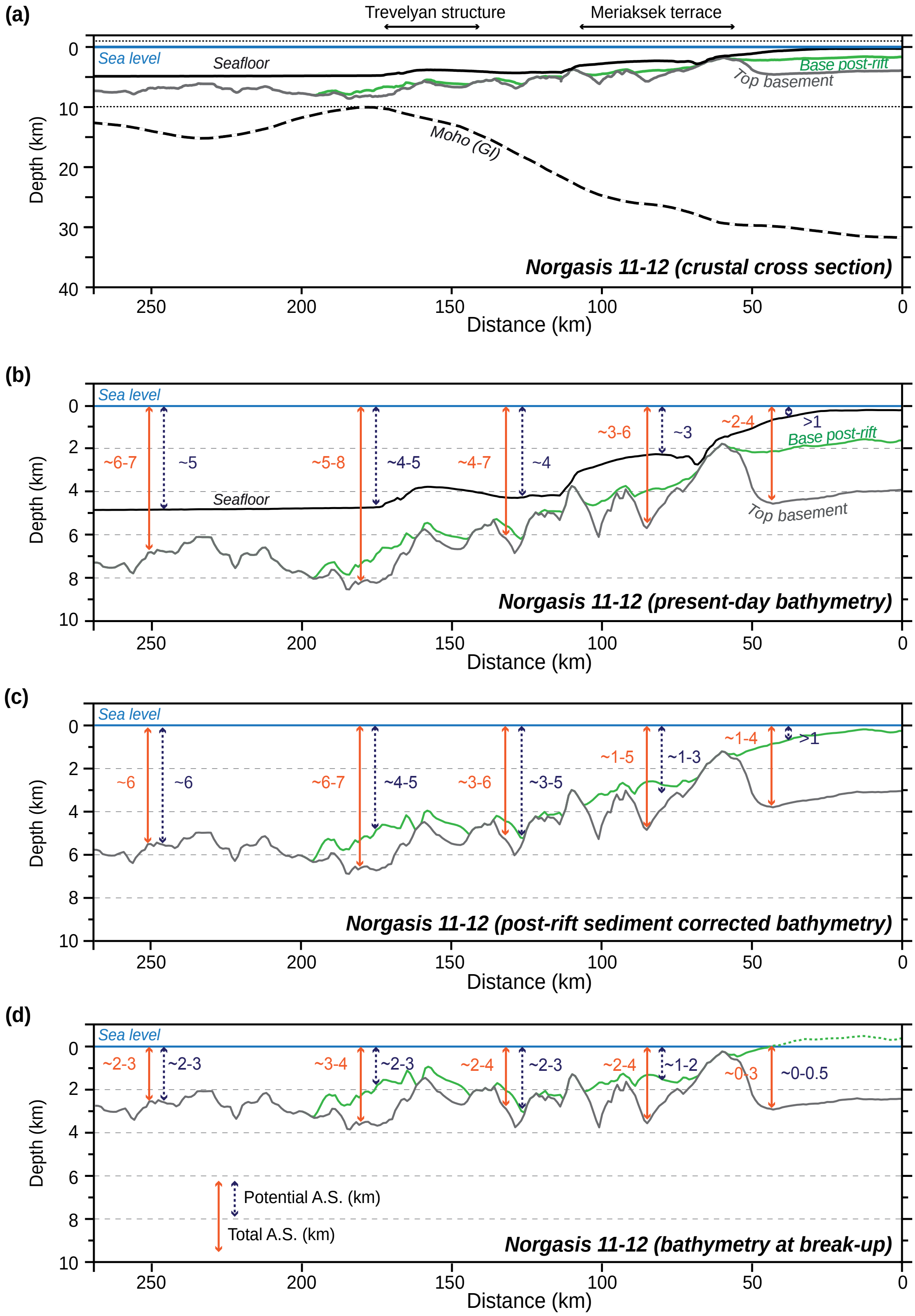


***Fig. 6**: (a) Crustal cross section from gravity inversion (GI) along the Norgasis 11-12 profiles. The first 10 km are framed and the evolution is quantified below (b), (c) and (d): Potential and total accommodation space evolution for (b) the present-day bathymetry (c) backstripped considering flexural unloading and decompaction only (post-rift sediment corrected bathymetry) and (d) backstripped applying reverse thermal modelling (bathymetry at break up).*

The bathymetry values resulting from flexural backstripping (fig 6 c/d) calculated for our case study provide first order isostatically balanced estimates of how much accommodation space can be created during continental margin rifting and breakup. The general trend of bathymetries predicted can then be related to the thinning domains previously described (fig 5c and fig 6). Hence, in the most proximal domain, where the crust is only weakly thinned, erosion is predicted at the beginning of post-rift (fig 6d) and the potential accommodation space created during rifting is minor in this part of the sections (fig 6). Contrastingly, the total accommodation space can record greater pre- to syn rift sedimentary aggradation. An increase of the potential accommodation space is related to the extreme thinning of the continental crust; the total accommodation slightly varies, depending on basement topography. The transition from continental to oceanic domains (interpreted as exhumed mantle, Thinon *et al.* 2003) preserves locally the largest total accommodation space. Oceanward, the total accommodation space slightly diminishes, recording the isostatic response due to the increasing thickness of magmatic additions in the oceanic domain (fig 6).

TABLE 2: Parameters used for flexural backstripping

Description of parameters	Value and units
Flexural properties	
Effective elastic Thickness	3 km
Brittle layer Thickness	15 km
Tectonic Model	
Eustatic sea level	Global eustasy
Main rift age (Ma)	110 Ma
Age of previous rift (Ma)	200
Previous Uniform Beta	1,2
Age of layer	
Layer 1 (Base Post rift)	108 Ma
Layer 2 (Top Basement	135 Ma
Dynamic subsidence Correction	400 m

5. The hyperextended stage: the Parentis basin

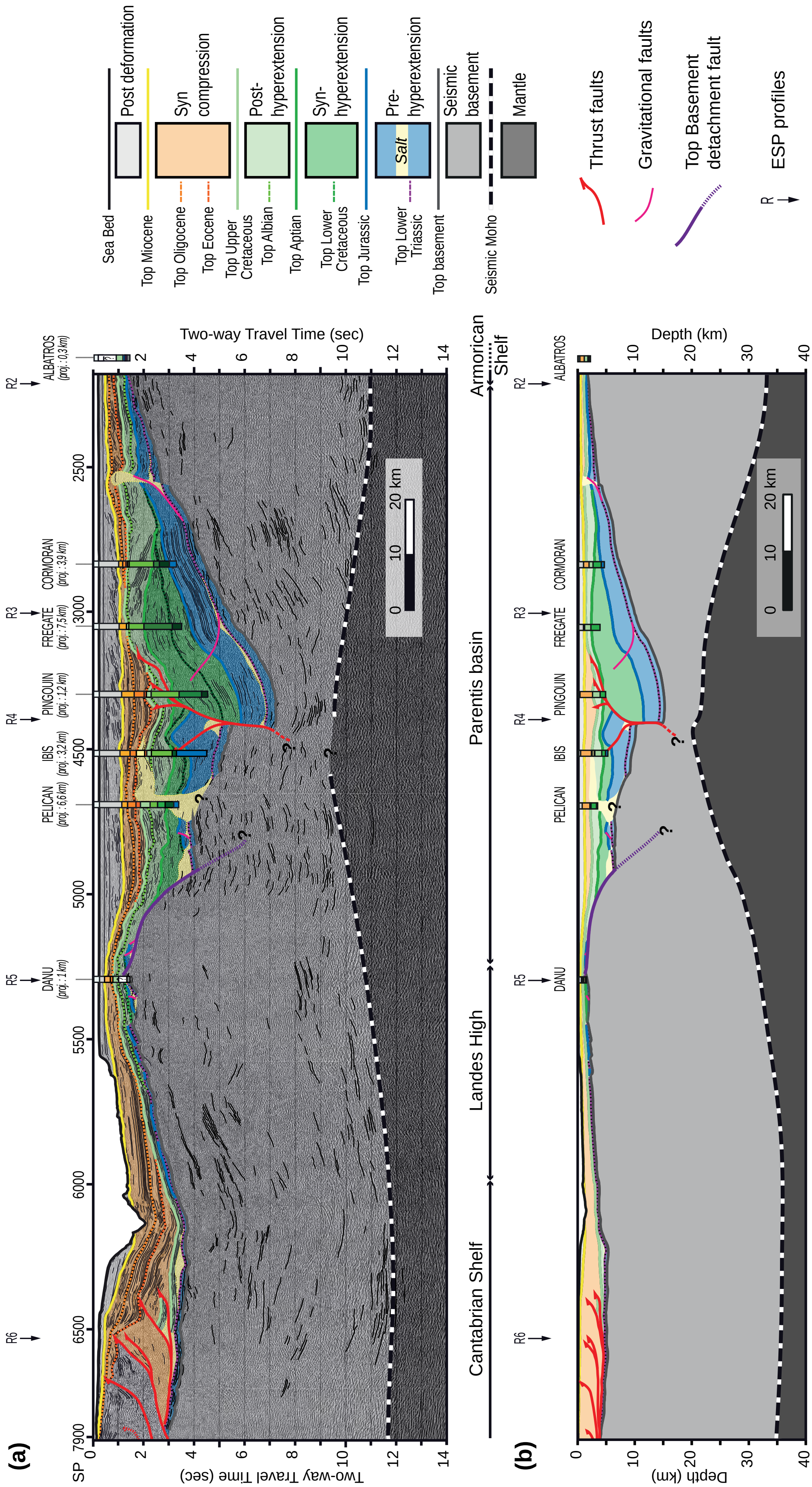
Located at the eastern termination of the Bay of Biscay, the offshore Parentis basin is a roughly 100 km wide rift basin bounded by the Landes High to the South and the Armorican shelf to the North (fig 3). The extreme crustal thinning of the basin to less than 10 km is relatively well captured by numerous geophysical surveys, including the ECORS Bay of Biscay profile (Pinet *et al*. 1987; Bois & Gariel 1994; Tomasino & Marillier 1997) and more recently by the MARCONI survey (Gallart *et al*. 2004; Ruiz 2007).

In the following, we will use the ECORS Bay of Biscay seismic profile (Pinet *et al.* 1987) to map first-order interfaces such as seafloor, top basement and Moho (relying on seismic reflection and refraction, Tomassino & Marillier 1997) similarly as for the previous example. Additional constraints on the stratigraphic architecture and sedimentary evolution are provided by numerous industry drill holes. These data enable us to define several key stratigraphic horizons that can be mapped across the basin (fig 7).

5.1. Seismic interpretation of top basement and Moho

Even though no drill hole is located directly on the ECORS Bay of Biscay profile, some drilling results can be projected (Pinet *et al.* 1987; Bois *et al.* 1997; Jammes *et al*. 2010a; 2010b; 2010c) to facilitate stratigraphic correlations (fig 7a). For the purpose of this study, we defined stratigraphic units that are pre-hyperextension (Triassic to Late Jurassic), syn-hyperextension (~Valanginian-Barremian to Aptian), post-hyperextension (Albian to Upper Cretaceous) and syn-compressional (Tertiary) sedimentary sequences. However, except from the Danu well over the Landes high, no drill-hole reached the Palaeozoic basement and only three of them (the Cormoran, Ibis and Pelican wells) penetrated Upper Jurassic sediments. Furthermore, the identification of the top basement is complicated by the presence of Triassic salt, associated with diapiric and gravitational structures (Biteau & Canérot 2007), strongly affecting the imaging of the southern border of the basin.

Fig. 7*: (a) ECORS Parentis seismic section and projected boreholes (Pinet et al. 1987; Bois et al. 1996; Jammes et al. 2010a; 2010b; 2010c, location of the profile fig 3). The main stratigraphic horizons and the interpreted top basement and moho are indicated. (b) Depth section of the ECORS Bay of Biscay profile. The interpreted stratigraphic horizons and the top basement have been depth converted and are reported. Moho depth is based on Expending Spread profiles (ESP) results (Marillier et al. 1988; Tomassino & Marillier 1997). For both sections, the position of the top basement detachment fault penetrated by the Saint Girons en Marensin and Contis boreholes is reported (Jammes et al. 2010b)*


(a)
(b)
Two-way Travel Time (sec)
Depth (km)
SP
ALBATROS
CORMORAN
FREGATE
PINGOUIN
IBIS
PELICAN
DANU
R2
R3
R4
R5
R6
20 km
Armorican Shelf
Parentis basin
Landes High
Cantabrian Shelf
Post deformation
Syn compression
Post-hyperextension
Syn-hyperextension
Pre-hyperextension
Salt
Seismic basement
Mantle
Sea Bed
Top Miocene
Top Oligocene
Top Eocene
Top Upper Cretaceous
Top Albian
Top Aptian
Top Lower Cretaceous
Top Jurassic
Top Lower Triassic
Top basement
Seismic Moho
Thrust faults
Gravitational faults
Top Basement detachment fault
ESP profiles

Insights on the seismic Moho depth are provided by the study of the expanding spread profiles (ESP), acquired at the same time as the ECORS Bay of Biscay section (location on fig 7, Marillier *et al.* 1988; Tomassino & Marillier 1997). The interface between the crustal basement and the seismic Moho is marked by a major increase in velocities (from ~6km/s in the crust to ~8km/s in the mantle).

5.2. Stratigraphic architecture and insights on the sedimentary infill history

Drill-hole observations constitute a critical dataset to investigate the infill history of the basin (Bois *et al.* 1997) but also to model the related subsidence evolution of the Parentis basin (as proposed by Brunet 1984; 1997). In this work, the sedimentary sequences are simplified with reference to the hyperextensional rift event preceding the opening of the Bay of Biscay.

The pre-hyperextensional sediments (from Triassic to Upper Jurassic) record a diffuse and widely distributed Triassic to Liassic rifting episode evidenced by the locale deposition of more than 2 km thick siliciclastic and evaporitic sequences, followed by the accumulation of thick Jurassic shallow marine limestones, dolomite and shales (Brunet 1984; Désegaulx & Brunet 1990). In the northern part of the basin this pre-hyperextensional sequence is continuous as shown in seismic sections (fig 7, SP~2500-4500); whereas over the Landes High and towards the Aquitaine shelf, this sequence is either absent or very thin (e.g. the Danu borehole, Bois *et al.* 1997).

The syn-hyperextensional succession record an Early Cretaceous (~Valanginian-Barremian) to Late Aptian subsidence (Désegaulx & Brunet 1990), related to rifting in the Bay of Biscay. During this phase of hyperextension, deltaic sequences were deposited in the basin (Curnelle & Dubois 1986). Similar to the pre-hyperextensional successions, the syn-hyperextension sequence is absent or highly reduced over the Landes High (Danu borehole, Bois *et al.* 1997), contrasting with the thick (~2 km to 3 km) time equivalent succession deposited in the basin. Furthermore, the syn-hyperextension infill is asymmetric in the Parentis basin. In the northern part of the basin the sequence is thickening towards the centre of the basin (fig 7, SP ~3000) while in the southern part (fig 7, SP ~5000-4500), the sequence is thinner (Ibis borehole, Jammes *et al.* 2010b; 2010c).

The deposition of the post-hyperextension sequence (Albian to Upper Cretaceous) is related to the thermal subsidence consecutive to seafloor spreading in the Bay of Biscay and can be subdivided into a rapid Albian subsidence, and reduced rates in Early Cretaceous time (Curnelle 1983; Désegaulx & Brunet 1990). During this post-rift phase, the basin was differentiated into a platform domain (to the South, over the Landes High, fig 7, SP ~7900-5000) preserving shallow water carbonates. In contrast, the northern basin (fig 7, SP~5000-

2500) is characterized by the deposition of more than 2.5 km of tubidites and shales (Curnelle 1983) at deep water depth (~2 km, estimations of Curnelle 1983).

The syn-compressional sequences of the Parentis basin record the progressive formation of the Pyrenean orogen starting from Late Cretaceous and lasting until Oligocene to Miocene time. In the central part of the Parentis basin only a little deformation is observed, related to a complex network of thrusts, sealed by Eocene to Oligocene sequences. The southernmost sediments of the ECORS Bay of Biscay section are deformed by thrust sheets and, over the Cantabrian shelf (SP~6500-6000, fig 7), present the classical architecture of a foreland basin (Roca *et al.* 2011).

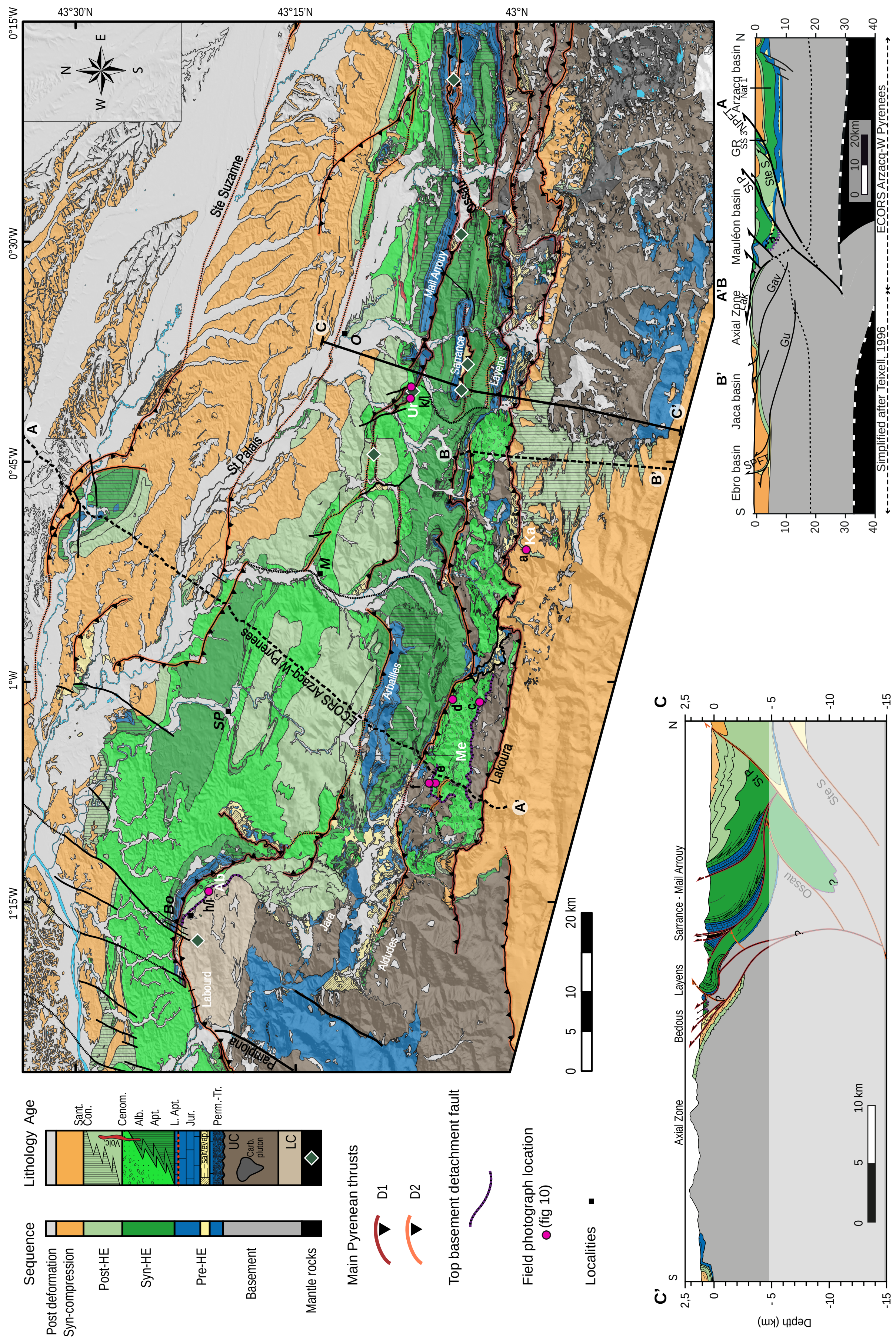
5.3. Crustal architecture and thinning structures: insights from depth sections and drill-hole data

Different interpretations have been proposed to explain the geometry of the Parentis basin, ranging from a synclinal basin (Bois & Gariel 1994; Bois *et al.* 1997) to a half graben shaped basin bounded to the South by one major crustal scale listric fault (interpretation of the MARCONI 3 profile, Ferrer *et al.* 2008). However, as already highlighted by Pinet *et al.* (1987), the amount of extension inferred from the few normal faults observed across the Parentis basin cannot explain the observed extreme crustal thinning. This crucial observation lead Jammes *et al.* (2010 a, b, c) to propose an alternative interpretation, suggesting that the architecture of the basin and its extreme crustal thinning are controlled by a top basement detachment fault, flooring the southern part of the basin. This interpretation is based on analogies with the Mauléon basin (Jammes *et al.* 2009) and supported by drill-hole data eastward of the ECORS Bay of Biscay profile (Saint Girons en Marensin and Contis boreholes, Jammes *et al.* 2010b). Hence, the authors suggested that the architecture of the Parentis basin is strongly asymmetric, related to a different evolution between an upper plate to the North and lower plate setting to the South. The upper and lower plate positions are defined with respect to a major, north dipping detachment system (Jammes *et al.* 2010a; 2010b; 2010c).

The interpreted depth section proposed in this work relies on published drill-hole data and includes some of the observations previously described by Jammes *et al.* (2010 a, b, c). Based on the identification of the main stratigraphic horizons and on the interpretation of the top basement (fig 7a), a depth section is proposed (fig 7b), using the same velocity structure as described above for the Norgasis 11-12 sections. Insights on the Moho depth are again provided by the expanding spread profiles (ESP), studied by Tomassino & Marillier, 1997. The evolution of the crustal architecture on the section is well recorded in the syn- to post-hyperextension sediments (a shallow carbonate platform to the South and a deep basinal domain to the North). As previously highlighted by Jammes *et al.* (2010a; 2010b; 2010c), the sedimentary sequences from the northern part of the Parentis basin show a relatively simple sag geometry and no major fault structures can be identified. Crustal thinning is progressive from ~30 km (on the Armorican shelf) to less than ~7km in the central part of the basin (towards the Ibis fault), coinciding with a thickening of the syn-hyperextension sediments and a deepening of syn- to post-hyperextension sequences (Curnelle 1983), indicating a progressive transition from a proximal to hyperthinned domain (fig 7). The architecture of the southern border of the basin is more complex. Over the Cantabrian shelf and Landes High, the post hyperextension shallow water carbonate platform (Curnelle 1983) overlies a ~30 km thick crust and the Moho is at 35 km depth, indicating only a weak crustal thinning during the rift event (i.e. proximal domain). Major crustal thinning only occurs northward of the Landes High. The Saint Girons en Marensin and Contis boreholes (Jammes *et al.* 2010b) that penetrated sequences of rocks characteristic of top basement detachment faults are located on a similar structural position, at the southern border of the Parentis basin. Hence the initiation of thinning, northward of the Landes High is coincident with the existence of a high-β extensional system. To the North, the extreme crustal thinning continues towards a major crustal discontinuity, the Ibis fault (Jammes *et al.* 2010a; 2010b; 2010c), delimiting a southward ~12 km thick crust from a ~5 km thick crust to the North.

6. The reactivated stage: the Arzacq-Mauléon basin

The Arzacq-Mauléon basin is located in the north-western part of the Pyrenean orogen. It is one of the best-documented examples of a reactivated hyperextended rift basin (fig3). Evidences of extreme crustal thinning and local mantle exhumation have been reported and discussed by several authors based on geophysical and geological approaches (e.g. Daignières *et al.* 1994; Grandjean 1992; Canérot 2008; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010; Jammes *et al*. 2009). Jammes *et al.* (2010a; 2010c) suggested that prior to its Pyrenean overprint the Arzacq-Mauléon basin showed a similar basin architecture to the present day Parentis basin, relying on comparable geological and geophysical (mainly gravity) observations in the two basins. The Mauléon basin was inverted as a large tectonic pop-up structure along north and south vergent thrusts (fig 8). In the North it was thrust over the Grand Rieu high and southern Arzacq basin along the north-vergent Saint Palais and Sainte Suzanne thrust systems (Daignères *et al.* 1994; Teixell 1996; 1998) and in the South over the Axial zone and southern foreland basin along the south-vergent Lakoura thrust system (fig 8, Teixell 1998; Muñoz 1992). The inversion and erosion of the Mauléon basin enable us to have locally access to the basal sedimentary sequences and the basement flooring the deepest parts of the basin (e.g. Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al*. 2010). In the following, we will focus on several outcrops preserving primary relationships between the sedimentary cover and the underlying basement. We will characterise the nature of sediments and basement rocks, define the nature of the top-basement and identify the position of these units within the present-day nappe stack (see map of the basin and a geological section across the basin fig 8 and 9).



Sequence
Lithology
Age
Post deformation
Syn-compression
Post-HE
Syn-HE
Pre-HE
Basement
Mantle rocks
Sant.
Con.
Cenom.
Alb.
Apt.
L. Apt.
Jur.
Perm.-Tr.
UC
LC
Main Pyrenean thrusts
D1
D2
Top basement detachment fault
Field photograph location (fig 10)
Localities
Ste Suzanne
St Palais
Mail Arrouy
Sarrance
Layens
Arbailles
Lakoura
Labourd
Pamplona
Aldudes
ECORS Arzacq-W Pyrenees
Ebro basin
Jaca basin
Axial Zone
Mauléon basin
Arzacq basin
Simplified after Teixell, 1996
Bedous
Sarrance - Mail Arrouy
Depth (km)

6.1. Sedimentary evolution

Although the Parentis and Arzacq-Mauléon basins show many analogies from a structural point of view (Jammes *et al.* 2010a; 2010c), the timing of extension and subsidence related to hyperextension is slightly younger in the Arzacq-Mauléon (late Aptian to Early Cenomanian) compared to the Parentis that is (Early Cretaceous to Aptian).

The pre-hyperextension sequences, Permian to lower Aptian in age, records several ill-defined Triassic to Early Liassic rift events (Brunet 1984) leading to the deposition of shales, sandstones, conglomerates, limestones and evaporates (e.g. the Germanic facies, Curnelle 1983; Fréchengues 1993). The debated Jurassic to Early Cretaceous evolution is related to the deposition of carbonates and marls, locally eroded and associated with the formation of bauxites (Biteau & Canérot 2007; Canérot 2008). The thickness variations observed within these pre-hyperextension sequences range from a few hundred meters to a few kilometres. Upper Triassic evaporates (Keuper formation) represent an important rheological horizon that acted as a major decoupling level during Cretaceous hyperextension (Canérot *et al.* 2001; Jammes *et al.* 2010b) and Pyrenean compression (Casteras *et al*. 1970a/c, Casteras *et al.* 1971; Le Pochat *et al.* 1978). Subsalt formations are coupled and deformed together with the upper crustal basement rocks and mainly outcrop south of the basin (e.g. south of the Labourd massif, in the Jara-Arbailles area and south of the Layens, fig 8). In contrast supra-salt formations are often decoupled from the underlying units and form almost E-W trending ridges (e.g. Sarrance, Mail Arrouy, fig 8). The Arbailles massif is an exception where the pre-compressional stratigraphic sequences and their relationship with the underlying basement are locally preserved.

Fig. 8*: Geological map of the Mauléon basin and the Southern part of the Arzacq basin. Lithologies and age of sequences are synthetized in a log. The location of the top basement detachment faults identified by Jammes et al. 2009 (in the Labourd massif) and Johnson & Hall 1989a/b (in the Mendibelza massif) are reported. The sections AA' and BB' corresponds to the interpreted section of the ECORS Arzacq-western Pyrenees seismic profile (crustal architecture modified from Teixell 1998 and Daignières et al. 1994) that is extended into the Axial Zone and southern the foreland basin as proposed by Teixell (1996). (Ga: Gavarnie thrust; Lak: Lakoura thrust GR: Grand Rieu high; Gu: Guarga thrust; SPFT: South Pyrenean Frontal Thrust; St P: Saint Palais thrust and Ste S: Sainte Suzanne thrust, forming the North Pyrenean Frontal Thrust system NPFT). The geological section CC' illustrates the pop-up inversion of the Mauléon basin and the interaction between the two phases of deformation. Outcrops and field photographs presented in fig 10: Ab: Abarratia quarry; Ka: Kalkuetta canyon; Me: Mendibelza massif; Ur: Urdach quarry. Localities: Bo: Bonloc; M: Mauléon; O: Oloron-Sainte-Marie; SP: Saint Palais. This map is simplified after the BRGM (1/50000) geological map of: Arthez de Béarn, Argelès-Gazost, Gavarnie, Hasparren, Iholdy, Laruns-Somport, Larrau, Lourdes, Oloron-Sainte-Marie, Orthez, Mauléon-Licharre, Morlaàs, Pau, Saint-Jean-Pied-de-Port, Tardets (see references in the bibliography, authorization n°: 2011-071)*

During the hyperextensional event, from late Aptian to Cenomanian, sedimentation in the Arzacq basin remained carbonate dominated as indicated by the deposition of more than 3 km of carbonates and marls (Brunet 1984). In the Mauléon basin, Upper Aptian to Lower Albian sequences record the progressive formation of half graben basins, with the development of carbonate platform over basement highs (Urgonian facies and Melobesia carbonates) and marls in the basins (e.g. Spicula marls). The progressive transition to a siliciclastic deltaic system is related to a rapid creation of accommodation space and strong subsidence during the onset of hyperextension. During this stage, the basin was filled in its southern part by a prograding deltaic system associated with conglomeratic facies (e.g. the Mendibelza conglomerates and lateral equivalent) while in the deeper parts of the basin further to the north turbidites are deposited (fig 8). These siliciclastic deposits referred to as “flysh noir deposits” (e.g Souquet *et al.* 1985) progressively filled the Mauléon basin until Cenomanian time. The main depocenter to the North of the basin (near the city of Mauléon) can reach a total compacted thickness of 4 km (Roux 1983; Fixari 1984; Souquet *et al.* 1985).

The transition from syn-hyperextensional to post-hyperextensional phase is marked by the occurrence of alkaline magmatism in the northern part of the Mauléon basin. The post-hyperextension sequence (Cenomanian to Coniacian) record a major break in the sedimentation of the Mauléon basin, with the transition from siliciclastic to carbonate dominated sedimentation. This source change is recorded by the development of a thick shallow water carbonate platform (~400 m, e.g. the Calcaire des Cañons platform) to the south of the basin directly overlying the basement (in the Axial zone). Concurrently calciturbiditic and hemipelagic systems filled the northern part of the basin.

The deposition of Santonian breccia and calcarenites unconformably over a karstic substratum records the onset of Pyrenean convergence. Upper Cretaceous to Tertiary flysch-type sedimentation results from the uplift and erosion of the axial domain. Syn-orogenic deposits are mainly preserved in the Jaca foreland basin to the South and to the North of the Saint Palais and Sainte Suzanne frontal thrust system forming the present-day Aquitaine basin. In the Mauléon basin, the last evidences for compressional deformation are Eocene to early Oligocene (Désegaulx *et al.* 1990; Teixell 1998) whereas it continued until late Oligocene to early Miocene in the Southern Pyrenees (Teixell 1996, 1998).

6.2. Nature of basement rocks

Different types of basement rocks are outcropping in the Mauléon basin. Palaeozoic meta-sediments locally intruded by carboniferous plutons (e.g. Eaux-Chaudes granite, Ternet *et al.* 2004) are representative of pre-hyperextension upper crustal rocks, outcropping in the Jara and Arbailles massifs and further to the south and southwest in the Axial Zone and the southern Basque massif (e.g. Aldudes massif, fig 8). Meta-sedimentary and metabasic granulites (Boissonas *et al.* 1974) described in the Labourd massif by Vielzeuf (1984) indicate conditions of equilibration prior to hyperextension corresponding to approximately 20 km depth (mineral assemblages of the basic granulites 775°C ±50° and 6 kbars ±0.5), hence representative of pre-hyperextension mid to lower crustal rocks. Finally, numerous outcrops of mantle derived serpentinized rocks (of variable size) can also be observed in the basin, mainly derived from lherzolite and websterite ultra-mafic protolites and showing a strong hydrothermal alteration with various degrees of serpentinisation (Fabriès *et al.* 1998; for distribution of mantle rocks, see fig 8). In the Mauléon basin, mantle outcrops occur in the northern part of the basin within thrust sheets associated to strongly altered crustal rocks (less than 100 m thick) at the base of the pre- and syn-hyperextensional sediments (e.g. Sarrance, Mail Arrouy).

6.3. Deformation history

Any attempt to reconstruct the pre-Pyrenean rift related history depends on the ability to unravel the deformation history and to distinguish the Pyrenean compressive structures from the rift related ones. In the following, we will mainly focus on a N-S geological section located in the eastern part of the Mauléon basin (fig 8 and 9), where the different compressional events are well-preserved, defining coherent compressive tectonic units within the large scale Mauléon pop-up structure (fig 9). Two opposite directions of thrusting can be observed. An older set of south-directed thrusts are overprinted by north directed ones (fig 8 and 9, geological section and interpretation of the ECORS Arzacq profile, fig 8). Hence, the initial thrust sheet emplacement related to the initial shortening of the N-S Pyrenean compression is delimited by south-directed thrusts that can be mapped throughout the basin (e.g. the Sarrance, Mail Arrouy or Lakoura thrust systems that are all Late Cretaceous, Teixell 1998, fig 8 and 9). The recognition of this initial phase of deformation is crucial as it can be used to define coherent compressive tectonic units that sample remnants of the former rift system (fig 9). With increasing shortening, a second phase of deformation led to the formation of mainly north-directed thrusts in the Mauléon basin that may either deform or overthrust former south directed thrusts (e. g. the Ossau thrust, Canérot *et al.* 2001 or north of the Labourd and Arbailles, or at a larger scale, the Saint Palais and Sainte Suzannne thrusts, fig 8 and 9) and is also related to the formation of south directed backthrusts in the Axial zone, responsible for the formation of the Jaca foreland basin (Teixell 1996).

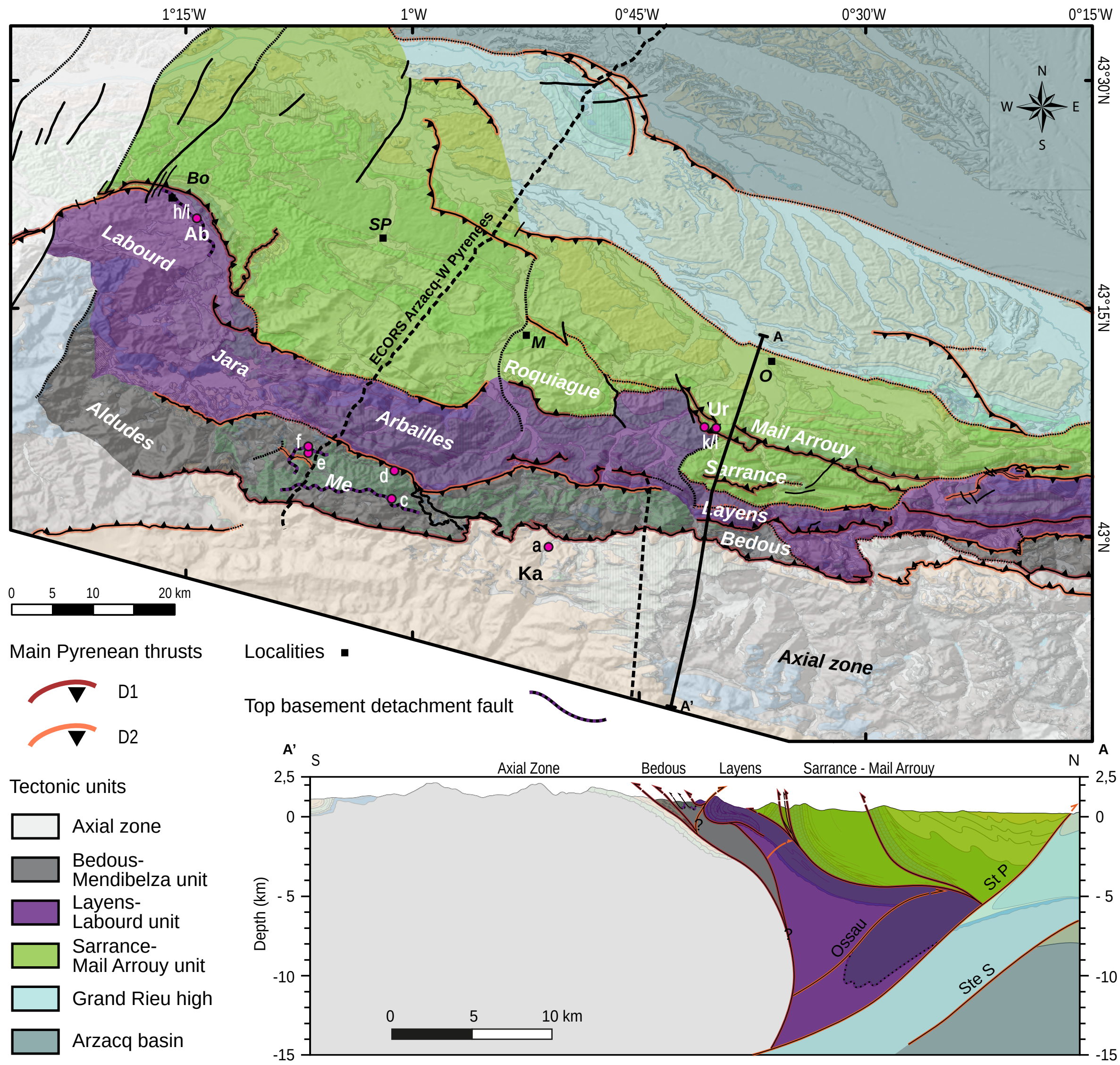


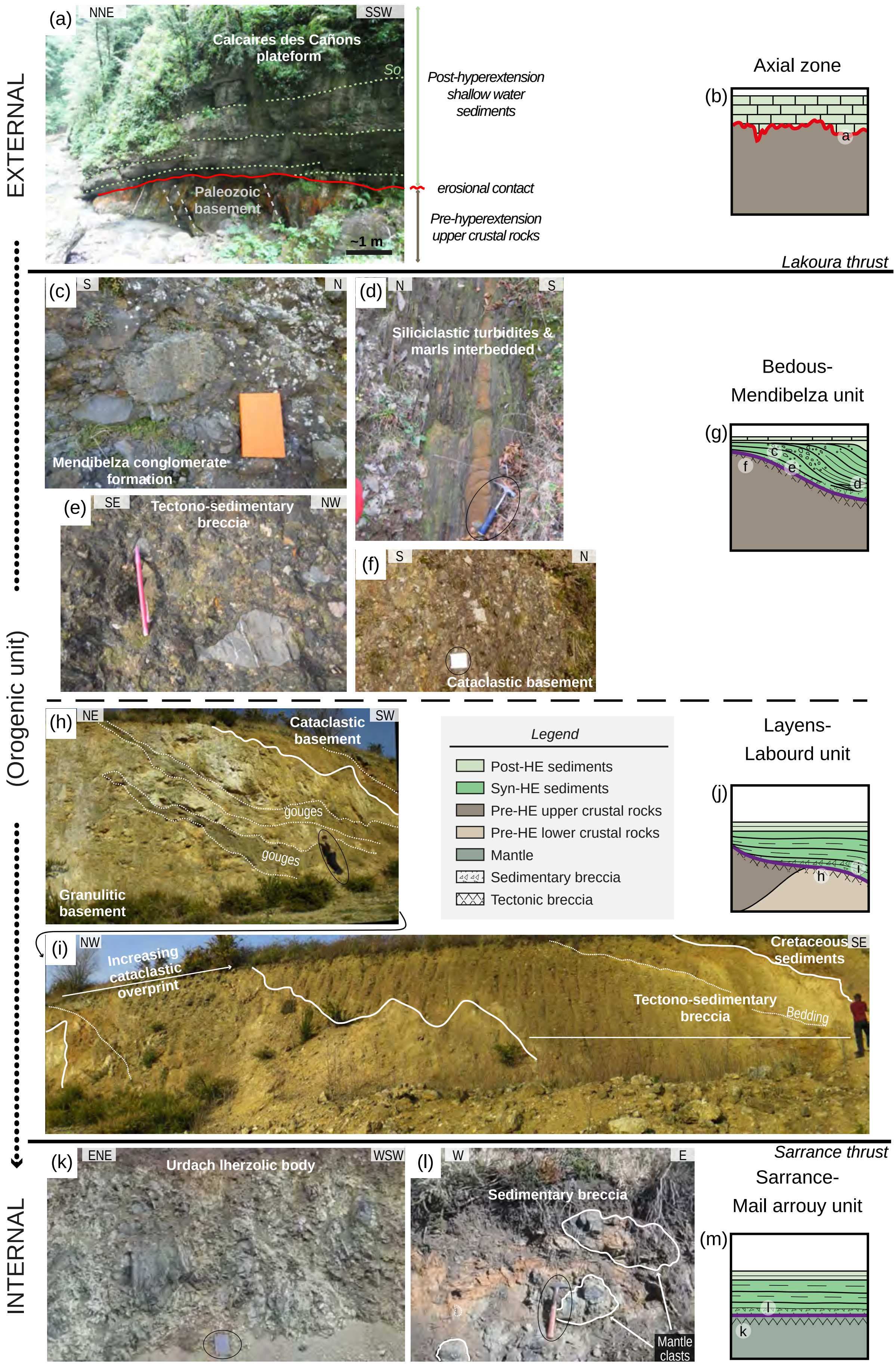
***Fig. 9**: Simplified structural map of the Mauléon basin and the Southern part of the Arzacq basin. The section AA' is the same as in fig 8 but only tectonic units are indicated. The location of the top basement detachment faults identified by Jammes et al. 2009 (in the Labourd massif) and Johnson & Hall 1989a/b (in the Mendibelza massif) are indicated. Outcrops and field photographs presented in fig 10: Ab: Abarratia quarry; Ka: Kalkuetta canyon; Me: Mendibelza massif; Ur: Urdach quarry. Localities: Bo: Bonloc; M: Mauléon; O: Oloron-Sainte-Marie; SP: Saint Palais.*

6.4. Remnants of the former rift system in the Pyrenean nappe stack

A structural analysis of the Mauléon basin enables the mapping of the compressive structures that define the present-day position of coherent tectonic units within the reactivated Mauléon basin (fig 9). The recognition and restoration of these units back into the initial position is essential as it defines the initial position of the units in the previous rift system. Additionally, the identification of pre-compressional relationships preserved in outcrops as well as observations on the nature of basement and sediments provide crucial information on the rift-related history that would not be obtained through the restoration of compressional structures only. Within each unit, key outcrops preserving the nature of contact between the basement and the pre-, syn-, and post-hyper-extensional sediments (e.g. Canérot 2008; Jammes *et al.* 2009, Lagabrielle *et al.* 2010; Debroas *et al*. 2010) can be found. In the following, we describe the main tectonic units and associated outcrops going from a lower to an upper structural position with reference to the first phase of compressional deformation (i.e. from more external to more internal positions in the orogen, fig 9 and 10).

The most external position is preserved in the Axial zone and is delimited to the North by the Lakoura thrust system (Teixell 1998) and lateral equivalents. The basement is formed of pre-hyperextension upper crustal rocks directly onlapped by post-hyperextension shallow water platform carbonates (for outcrop observations see fig 10 a/b and location fig 8 and 9). The erosion of the top of the basement and the deposition of post-hyperextension shallow water sediments indicate that over this domain, no accommodation space was formed during syn-hyperextension time. This observation suggests that the crust was not or only slightly thinned indicating that this domain was in a proximal position relative to the basin (Jammes *et al*. 2009).

Fig. 10*: Field photographs and simplified cartoons of key observations from remnants of the former rift system preserved in the Pyrenean nappe stack (location of outcrops fig 8). (a-b) Axial zone unit: (a) Contact between the "Calcaires des Cañons" platform and underlying eroded Palaeozoic basement exposed in the Kalkuetta canyon. (b) Cartoon representing the contact between the post-hyperextensional sediments and basement observed in the Axial zone. (c-g) Bedous-Mendibelza unit: (c-d) Sedimentary cover of the Mendibelza massif (Mendibelza conglomerate formation and interbedded marls and siliciclastic turbiditic deposits), (e) underlying tectono-sedimentary breccia and (f) cataclastic basement. (g) Cartoon illustrating the onlap of syn-hyperextensional sediments onto exhumed pre-hyperextensional upper crustal basement as observed in the Bedous-Mendibelza unit. (h-j) Layens-Labourd unit: (h-i) From left to right, deformation of the granulitic basement (gouges, cataclastic overprint) and contact with the overlying tectono-sedimentary breccia and sedimentary cover in the Abarratia quarry (j) Cartoon showing the exhumed pre-hyperextensional mid to lower crustal basement onlapped by syn-to post-hyperextensional sediments in the Layens-Labourd unit (as previously described by Jammes et al. 2009). (k-m) Sarrance-Mail arrouy unit: (k) Urdach lherzolitic body and (l) syn-rift breccia reworking mantle and basement rocks in the Urdach quarry as exemplified in (m).*


EXTERNAL
(Orogenic unit)
INTERNAL
(a)
NNE
SSW
Calcaires des Cañons plateform
So
Paleozoic basement
~1 m
Post-hyperextension shallow water sediments
erosional contact
Pre-hyperextension upper crustal rocks
Axial zone
(b)
a
Lakoura thrust
(c)
S
N
Mendibelza conglomerate formation
(d)
N
S
Siliciclastic turbidites & marls interbedded
(e)
SE
NW
Tectono-sedimentary breccia
(f)
S
N
Cataclastic basement
Bedous-Mendibelza unit
(g)
c
e
f
d
(h)
NE
SW
Cataclastic basement
gouges
gouges
Granulitic basement
Legend
Post-HE sediments
Syn-HE sediments
Pre-HE upper crustal rocks
Pre-HE lower crustal rocks
Mantle
Sedimentary breccia
Tectonic breccia
Layens-Labourd unit
(j)
h
i
(i)
NW
Increasing cataclastic overprint
Cretaceous sediments
SE
Tectono-sedimentary breccia
Bedding
Sarrance thrust
(k)
ENE
WSW
Urdach lherzolic body
(l)
W
E
Sedimentary breccia
Mantle clasts
Sarrance-Mail arrouy unit
(m)
l
k

The Bedous-Mendibelza and Layens-Labourd units correspond to higher (i.e. more internal) positions and are separated from more internal ones by the South-directed Sarrance thrust (fig 9). Western and eastern structural equivalents are delimited by north-directed thrusts (e.g. north of the Arbailles and eastern part Labourd massifs, fig 9). On the geological section in fig 9 the two units are delimited from each other by a South directed thrust preserved at the base of the Layens, overprinted westwards and eastwards by North directed thrusts (e.g. North of the Mendibelza massif, fig 9). The Bedous-Mendibelza unit is made of thick aggradational sedimentary sequences (~1 km) that are deposited onto pre- hyperextension upper crustal rocks and sediments corresponding to the Mendibelza and Igountze massifs. The syn-hyperextension sequences consist in their southern proximal part of massive conglomerates (Mendibelza formation, fig 10c) and their northwards time equivalent turbidites and marls (fig 10d). The contact between the upper crustal rocks and the onlapping sediments is formed by a cataclastic fault zone (fig 10f) associated with tectono-sedimentary breccias (fig 10e). This contact was interpreted as a top basement detachment fault by Johnson & Hall (1989 a,b) and Miranda-Avilez *et al.* (2005). This interpretation was made in analogy with supra-detachment basin structures described in the Basin and Range. Hence, the progressive thickening and deepening of sediments onlapping onto exhumed basement rocks indicate the progressive creation of accommodation space along active detachment faults (fig 10g).

In contrast, in the Labourd massif area, in the Layens-Labourd unit, granulites derived from the pre-hyperextension mid- to lower crust (Vielzeuf 1984) are onlapped by late syn- to post-hyperextension sediments. Jammes *et al.* (2009) described in detail the nature of the contact between basement and sediments and identified a major brittle fault zone preserving syn-tectonic breccias and onlapping sediments. The authors interpreted the mid- to lower crust rocks as being exhumed along a detachment system (Abarratia quarry, fig 10h/i) and later resedimented into syn- hyperextension sediments (e.g. Bonloc breccia, Claude 1990). The occurrence of exhumed pre- hyperextension mid-to lower crustal rocks in the Labourd massif and their reworking in syn- hyperextension sediments shows that the whole upper crust had to be removed and hence thinned (fig 10j).

The Sarrance-Mail Arrouy unit (fig 9) and its structural equivalents to the west (e.g. Roquiague) correspond to the most internal units, delimited to the North by the Saint Palais thrust (fig 9). Within this unit crustal basement rocks are rare and only represent thin slices (usually less than 100 m thick) associated to mantle outcrops. Key observations come from the Urdach area (fig 10, Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010), where the occurrence of mantle clasts into a sedimentary breccia within the late syn- hyperextension sedimentary cover (fig 10k/l) suggests that at least locally in the Mauléon basin, mantle rocks had to be exhumed to the seafloor during Cretaceous hyperextension. The sediments of this unit are mainly represented by thick accumulations of syn-hyperextension deep water sediments (in the western part of the basin) and pre-hyperextension sediments representing E-W ridges thrust southward (e.g Mail Arrouy and Sarrance) over upper Triassic evaporates, interpreted as extensional allochthons by Jammes *et al.* (2009). The large accumulation of syn- hyperextension sediments (e.g. near Mauléon city, Roux 1983; Fixari 1984; Souquet *et al.* 1985) indicates the creation of large accommodation space. This observation goes along with the occurrence of mantle rocks reworked in the sediments. All these observations suggest that the central part of the Mauléon basin may have been floored at least locally by exhumed mantle rocks (fig 10m).

7. Synthesis: observations and diagnostic elements for offshore and onshore systems

Interpretations of onshore data often use templates derived from present-day margins, while seismic interpretations are often inspired by onshore observations. The use of analogues has a long tradition in Earth Sciences and has been used successfully for the interpretation of proximal rifted margins and oceanic crust. However, in the case of hyper-extended domains, the use of analogues is still hampered by the lack of well-studied examples and access to critical data (such as drill-hole). The aim of this paper is twofold: firstly to describe and characterize quantitatively and qualitatively onshore and offshore observations and secondly propose coherent, semi-quantitative, observation driven interpretations of hyperextended rift systems.

7.1. Offshore vs onshore type observations

The investigation of present-day hyperextended domains mainly relies on indirect geophysical approaches as only a few drill-holes have penetrated these domains worldwide. In contrast, such domains are exposed in internal parts of collisional orogens providing direct access to these domains. However, while offshore the structures are preserved, but cannot be directly observed, onshore they can be directly observed, but are overprinted or disrupted. Thus, the challenge is to define a number of objective first-order observations that enable the development of an observation-driven and coherent interpretation of hyperextended domains (fig 11).

At a seismic scale, objective interpretations rely on the recognition of first order interfaces such as (1) sea level, (2) seafloor (interface between water/top of sedimentary infill), (3) top basement (base of sediments/top of acoustic basement) and (4) Moho (crust/mantle boundary). These first order observations form the input parameters to estimate accommodation space, crustal thickness or crustal thinning as proposed in this study (fig 5, 6, 7b and fig 11). Additional information comes from observations of the depositional geometry and stratigraphic architecture and their relation to structures controlling the formation of accommodation space. These observations provide constraints for the interpretation of extensional systems (low or high-β extensional setting, Wilson *et al.* 2001) (fig 2).

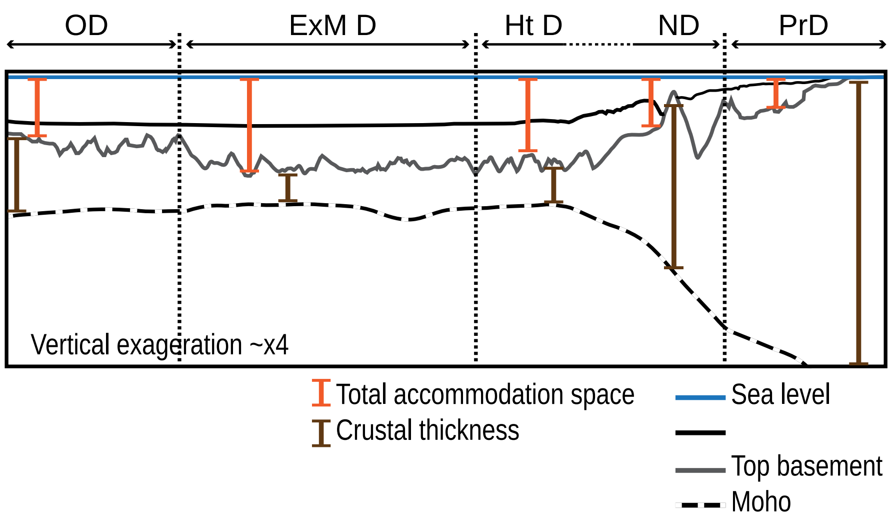
Fig. 11*: Comparison between geological and geophysical approaches. The different types of observation (seismic vs outcrop field observations) result in different diagnostic elements (lower part) that can be used to define structural domains at rifted continental margins or in their fossil analogues (central part). PrD: proximal domain, ND: Necking domain, Ht D: Hyperthinned domain, ExM D: Exhumed Mantle Domain, OD: Oceanic Domain.*

GEOPHYSICAL APPROACH

Seismic observations:

- sea level
- seafloor (water/sediment interface)
- top basement (sediments/acoustic basement interface)
- seismic Moho (Vp=8 km/s)

Domain definition at present-day marine systems



Diagnostic elements:

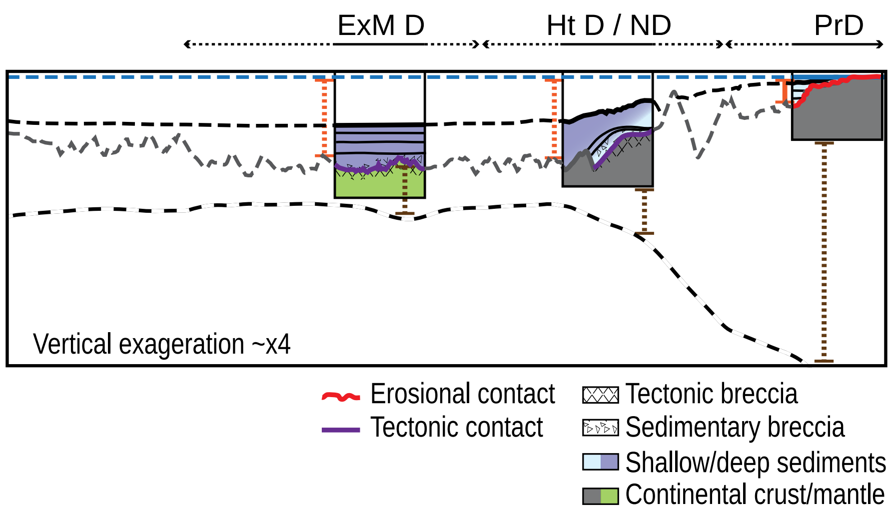
- accommodation space
- crustal thickness/thinning
- extensional setting (high/low-ß settings)

GEOLOGICAL APPROACH

Outcrop observations:

- nature and thickness of sediments
- nature of basement rocks
- nature of contact between sediments and basement rocks

Domain definition in onland deformed analogues



Diagnostic elements:

- paleo-water depth and aggradation
- variation of basement nature
- nature of top basement (tectonic, erosional, magmatic)

At a drill hole and outcrop scale, direct observations are limited in scale but provide high resolution and the possibility of direct sampling. As a consequence large-scale geometries and/or relationships are difficult to be captured, but direct observation enables the assessment of the nature of sediment and basement rocks as well as of the contact separating the two that can be tectonic, erosional, depositional or magmatic. These observations can then be interpreted in order to determine the depositional environment and aggradation history of sediments that provide insights on paleo-water depth and accommodation space creation. Changes in the nature of basement rocks and in particular information on their original position before rifting and the time/temperature/pressure path of these rocks to the surface are key data that provide hints for crustal thinning and the thermal structure associated with this process. The identification of outcrops preserving primary relationships between sediments and basement rocks are of critical importance to identify rift-related structures and distinguish extensional settings (Wilson *et al*. 2001).

7.2. Domain determination: diagnostic elements

In this paper, we have focused on the Bay of Biscay and Western Pyrenees where seismic images, drill-hole data and field exposures exist for one and the same rift system. This data set enables the application of indirect geophysical observations (Western Approaches margin, fig 4, 5, 6), geological observations (Mauléon basin, fig 8, 9, 10) or both type of approaches (Parentis basin, fig 7) to investigate different stages of this rift system. Based on results from these examples, we are now able to point out diagnostic observations and elements that can be used to define different domains across rift sections as well as to propose geologically more coherent and physically better constrained interpretations.

7.2.1. Offshore geophysical observations: defining limits of domains (fig 11)

In order to characterize the Western Approach margin and to estimate crustal thinning and the development of the accommodation space (during rifting) we defined the main interfaces (sea-level/seafloor/base sediments/Moho), their morphology and structures affecting them that lead to the creation of accommodation space. As exemplified by the Norgasis 11-12 section, the quantification of crustal thickness and thinning factor allows three domains to be distinguished: 1) the proximal domain that is weakly thinned, 2) the necking and hyper-thinned domain where most of the thinning occurs and 3) the exhumed mantle or oceanic domain where complete thinning of the continental crust is achieved (figs 1 and 5 and fig 11). In magma-poor rifted margins, accommodation space is to first order controlled by crustal thickness, sedimentary load and post-breakup thermal subsidence. The total accommodation space generally increases oceanward, as continental crustal thinning increases (fig 6). The distinction between the exhumed mantle (OCT) and oceanic domains corresponds to a morphological boundary and step in basement topography (figs 4, 5, 6) with smaller values of accommodation space in the oceanic domain. Assumptions on extensional settings (high vs low-β extensional setting, Wilson *et al.* 2001) depend on basin architecture ("hyperextended sag" basin or half graben type basin, fig 2). Low-β extensional settings can be observed throughout the margin (e.g. Norgasis 11-12, Parentis basin), whereas high-β extensional settings are observed where major thinning occurs: in the necking domain, over hyper-thinned crust or in the exhumed mantle domain (fig 4 and fig 7).

7.2.2. Geological observations: definition of limits and domains (fig 11)

In the case of the Mauléon basin, only remnants of the former hyper-extended rift system can be observed. The interpretation relies on geological observations on the nature of sediment, basement and of their interface as well as on their position within the thrust stack. Estimates of the total accommodation space created during rifting rely on combined observations from post-rift depositional environment and aggradation history (sedimentary thickness). Indeed only considering depositional environment could be misleading, even if post-rift shallow marine sediments are mainly observed in proximal domains, in the cases of a large aggradation during rifting, shallow marine sediment could mask large creation of accommodation space. Basement nature changes with crustal thinning but it cannot be used as a stand-alone criterion to define domains, as only the occurrence of mantle rocks can indicate a complete crustal thinning. Indeed continental crust (upper and lower crust) can be encountered in all domains (except from the oceanic domain) either in a footwall position or as extensional allochthons in a hanging-wall position in the exhumed mantle domain. Finally, the nature of the contact between the basement and overlying sedimentary cover is critical as it enables to identify fault structures (i.e. extensional settings) but also basement erosion and magmatic additions. The identification of top basement detachment faults is critical as in the case of hyperextended rift systems it occurs in the hyperextended domain within the necking or hyper-thinned and exhumed mantle domains. Sub-areal erosion of basement during the rift stage indicates that there was no creation of accommodation space and is characteristic of the proximal domain, but may also occur in the necking domain during early stages of the rift evolution (Péron-Pinvidic & Manatschal 2009). Magmatic additions are dominant in the oceanic domain, but may also occur in the hyperthinned or exhumed mantle domains. None of the criteria presented above can be used as stand-alone criteria to unambiguously replace an outcrop back into its rift context. Only by combining the different observations more precise assumptions about the domains can be done.

7.2.3. Additional constraints

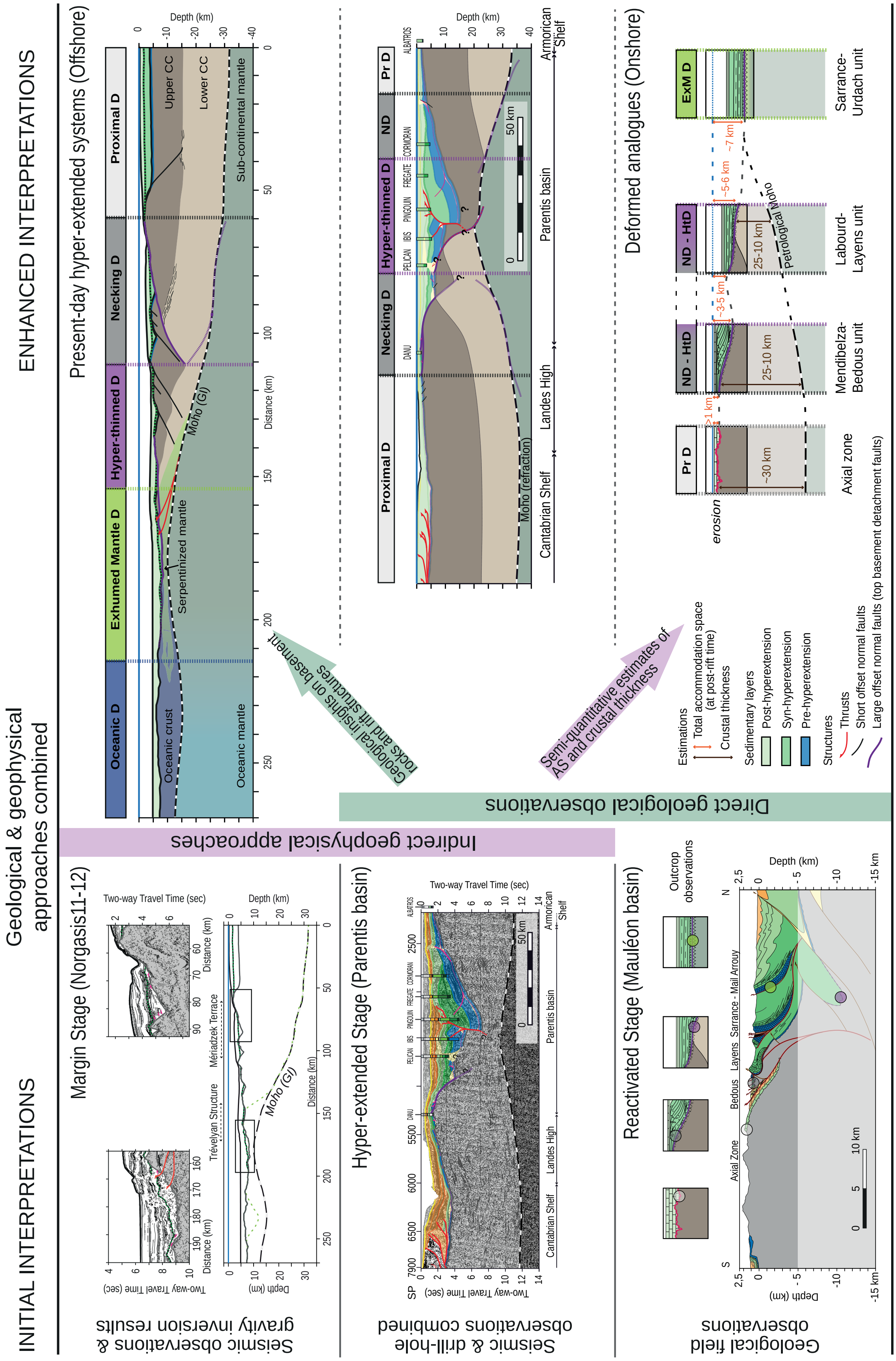
Relying only on the direct and indirect approaches proposed in this study, the distinction between the necking and hyperthinned domain is not straightforward. Following the work of Sutra *et al.* (2013), the distinction between the necking domain and hyperthinned continental crust is based on the existence or removal of ductile layers. Deformation in the necking domain is decoupled at mid-crustal ductile levels, while deformation in the hyper-thinned crust is coupled at the scale of the crust (faults can cut through the completely embrittled crust and penetrate into mantle, Pérez-Gussinyé *et al.* 2003). Sutra *et al.* (2013) showed that the change from decoupled to coupled deformation is accompanied by polyphase deformation and a change in the age of the syn-tectonic sediments getting younger in the hyperextended domain going oceanwards. This tectonic migration of the deformation is reported both offshore on the Iberian margin and also in onshore remnants preserved in the Alps (Mohn *et al.* 2010, Masini *et al.* 2013). In the Bay of Biscay the deeper parts of the margin have not been drilled, but it is likely that also in this basin the syn-tectonic sediments observed on the proximal margin are older than the one overlying the distal margin (3B layer described by Thinon *et al.* 2002).

8. Implications of a combined geological/geophysical approach for the interpretation of offshore and onshore section

In the previous descriptions of the three sections, indirect and direct approaches were applied to describe the different rift domains (fig 12). The Parentis basin was the only example for which both geological and indirect geophysical quantitative approaches could be used to interpret the structure and evolution of the basin. In this section, we will propose interpretations for the Western Approach margin and the Mauléon paleo-rift system based on the combination of geological insights and quantitative estimations of accommodation space and crustal thickness derived from the study of the three sections (fig 12).

8.1. The Parentis basin: combined approaches

The combination of drill-hole observations with reflection and refraction seismic data and depth conversion enabled us to describe three different domains on the ECORS Bay of Biscay section. As previously proposed by Jammes *et al.* (2010a; 2010b; 2010c), the Parentis basin appears strongly asymmetric with a northern sag basin architecture, where domains are transitional and a more complex southern geometry. To the south, the proximal domain over the Cantabrian shelf and Landes high was weakly thinned during hyperextension as indicated by the absence of syn- hyperextension sediments and the overlying post-hyperextension shallow water platform. Only a few very minor normal faults can be observed. North of the Landes High, thinning of the crust is initiated in the necking domain, where a top basement detachment fault was drilled eastward of the section (the Saint Girons en Marensin and Contis boreholes, Jammes *et al.* 2010b), structurally delimiting the hyper-thinned domain to the south. The poor imaging of the ECORS Bay of Biscay profile north of the Landes High can be related to important salt tectonics that may result in the decoupling of pre-hyperextension sediments from the underlying basements and in diapir formation (e.g. Mathieu 1986; Biteau *et al.* 2006). Crustal thinning continues to the north and is interpreted as resulting from another top basement detachment fault rooting towards the Ibis fault corresponding to a crustal thickness discontinuity between the northern and southern parts of the basin (fig 12).


INITIAL INTERPRETATIONS
Geological & geophysical approaches combined
ENHANCED INTERPRETATIONS
Seismic observations & gravity inversion results
Margin Stage (Norgasis11-12)
Present-day hyper-extended systems (Offshore)
Seismic & drill-hole observations combined
Hyper-extended Stage (Parentis basin)
Geological field observations
Reactivated Stage (Mauléon basin)
Deformed analogues (Onshore)
Indirect geophysical approaches
Direct geological observations
Geological insights on basement rocks and rift structures
Semi-quantitative estimates of AS and crustal thickness
Oceanic D
Exhumed Mantle D
Hyper-thinned D
Necking D
Proximal D
Pr D
ND
ND - HtD
ExM D
Upper CC
Lower CC
Sub-continental mantle
Serpentinized mantle
Oceanic crust
Oceanic mantle
Moho (GI)
Moho (refraction)
Petrological Moho
Cantabrian Shelf
Landes High
Parentis basin
Armorican Shelf
Axial zone
Mendibelza-Bedous unit
Labourd-Layens unit
Sarrance-Urdach unit
Estimations
Total accommodation space (at post-rift time)
Crustal thickness
Sedimentary layers
Post-hyperextension
Syn-hyperextension
Pre-hyperextension
Structures
Thrusts
Short offset normal faults
Large offset normal faults (top basement detachment faults)
Outcrop observations
Two-way Travel Time (sec)
Depth (km)
Distance (km)
Mériadzek Terrace
Trévelyan Structure
erosion
~30 km
25-10 km
>1 km
~3-5 km
~5-6 km
~7 km
Axial Zone
Bedous
Layens
Sarrance - Mail Arrouy
S
N

8.2. The Western-Approach margin: insights from outcrop observations

Based on indirect approaches, i.e. estimations of crustal thinning and accommodation space as well as on basin morphology, several domains were successfully defined. Nevertheless as exemplified by the Parentis example, geological insights can greatly improve the usual low resolution of seismic interpretation (fig 12). On the Norgasis 11-12 section, the basement morphology is ill-defined in the proximal domain, nevertheless one major normal fault can be observed and since no major thinning seems to be associated to it, it is interpreted to be decoupled at mid-crustal levels. Thinning is initiated over the Meriadsek terrace, where a high-β extensional setting can be identified (fig 4b, fig 12) and interpreted as a top basement detachment fault structuring the necking domain. The transition to hyper-thinned crust is not evident based on thinning arguments only, but further interesting observations come from the description of the syn-rift stratigraphic architecture overlying the two domains. Over the Meriadzek terrace, in the necking domain, sediments within the basins show evidence for syn-to post-tectonic geometries, whereas southward in the hyperthinned domain the sediments of the same age are tilted in a pre-tectonic position. This change of syn-rift stratigraphic architecture may be related to a migration of the deformation from the necking domain, where deformation is decoupled at the scale of the crust, to the hyperthinned crust where the whole crust is coupled (Pérez-Gussinyé *et al.* 2003; Sutra & Manatschal 2012). The transition to the exhumed mantle domain corresponds to complete thinning of the continental crust (fig 5d). Gravity inversion predicts an inflection of the gravity Moho on the depth section (fig 5 and fig 12). The distal high-β extensional setting identified south of Trevelyan is suggested to represent a top-basement detachment fault responsible for the exhumation of mantle rock overlain by the youngest syn-rift sediments (Aptian in age, Thinon *et al.* 2002). Finally, the transition to the oceanic domain is characterized by a ramp of the top-basement morphology and the direct onlapping of the post-rift sediments. This transition corresponds to the progressive magmatic overprint of the exhumed mantle domain (fig 12).

Fig. 12*: Philosophy of the geological/geophysical approach developed in this study. Left side: Initial interpretations proposed for the sections based either on geophysical approaches (Norgasis 11-12) or geological observations (Mauléon basin). The approach developed in this study relies on the Parentis example for which both geological and geophysical approaches could be applied directly. Right side: contribution of a combined geological/geophysical approach for offshore seismic interpretations and studies in deformed analogues. The Parentis and Norgasis examples provide estimations of accommodation space (AS) and crustal thickness for the Mauléon example. The Parentis and Mauléon example inspire geological insights for the Norgasis 11-12 example.*

8.3. The Mauléon basin: semi-quantitative estimates of accommodation space and crustal thickness

The identification of rift domains in the Mauléon basin relies on the identification of outcrops of remnants of rift domains and their position within coherent tectonic units. However, in spite of the high resolution that outcrop observations can give, the compressional reactivation has completely obliterated the former crustal architecture. In the same way, as the Parentis basin provided geological insights to the Norgasis section, it can offer first-order semi-quantitative estimates of crustal thickness and accommodation space for the remnants of the former rift system preserved in the various Pyrenean nappes in the Mauléon basin. From the geological sections only, three domains were characterized; a proximal domain (axial zone), a domain were upper and mid-to lower crustal rocks were exhumed (Mendibelza massif, Labourd massif) and a domain with exhumed mantle. In the proximal domain, the erosion of the basement overlain by post hyper-extension shallow-water sediments indicates that there was no accommodation space during hyper-extension; hence, this domain was probably not or only very weakly thinned, corresponding to a 30-35 km thick crust (as observed in the Parentis basin, fig 7b and 12). The observation of exhumed upper and mid-to lower crustal rocks respectively in the Mendibelza and Labourd massif indicate the existence of high-β extensional settings. Additionally, the presence of proximal (e.g. the Mendibelza conglomerates) to deeper water sediments (e.g. the Flysch Noir deposits, Souquet *et al.* 1985) indicate the progressive creation of accommodation space that could be related to the thinning of the crust as observed in the necking domain or in the case of the hyperthinned domain. Finally, remnants of exhumed mantle were also observed and described by Jammes *et al.* (2009). The occurrence of mantle outcrops preserving only a few hundred meters thick crust and their reworking in syn- hyperextension sediments indicate that at least locally the thinning of the continental crust was complete. Using quantifications of accommodation space at break-up time (Norgasis 11-12 section, fig 5), a water depth close to 3 to 4 km could be expected for sediment deposition in the exhumed mantle domain, increasing to 6-7 km at early post-rift time (fig12).

9. Discussion

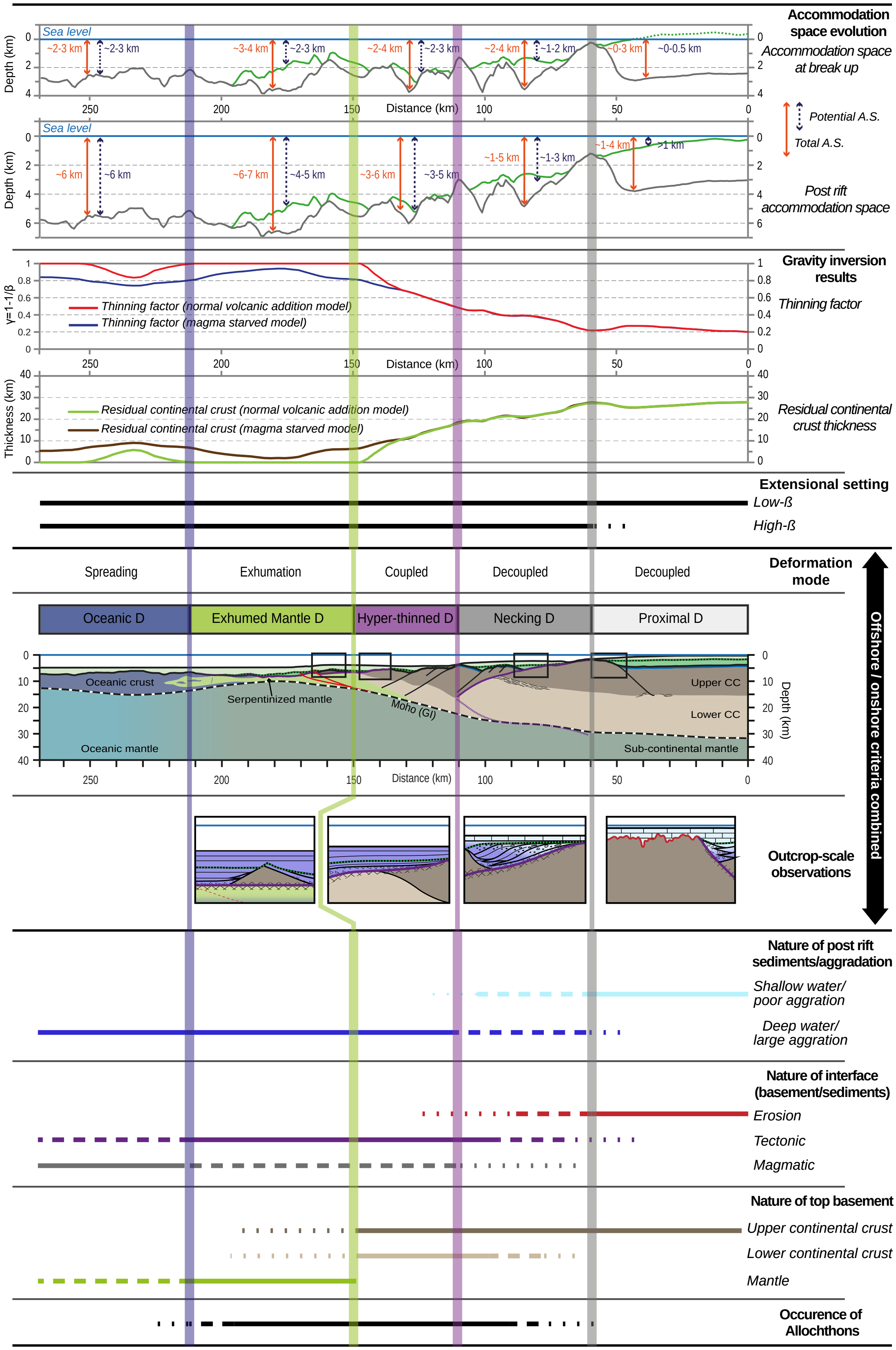
9.1. Towards a reconciliation of observations from present-day and fossil analogues for rift domain definition (fig 13)

The investigation of different case studies using geophysical approaches, geological observations and a combination of both (fig 12) enabled us to propose a self-consistent approach to define rift domains that is both observation driven and semi-quantitative (fig 13). Indirect quantification based on the Norgasis 11-12 profiles are proposed in the upper part of the figure, whereas the lower most part presents an overview of key geological observations relying on observations from onshore study (mainly the Mauléon basin). Diagnostic elements (from direct and indirect datasets) that enable to characterise domains have already been discussed, however a critical point is that unambiguous characterisation of domains relies on the combination of observations. We believe that this synthesis of direct and indirect observations can be used to improve the interpretation of domains both in present-day rifted margins (e.g. the Western approach section, fig 12 and central part of fig. 13) or in collisional orogens that contain remnants of rift domains (e.g. the Mauléon basin, fig 12).

9.2. Major limitations

The Bay of Biscay and Western–Pyrenees case study described in this paper can be considered as representative of an equilibrated magma-poor hyperextended rift system, i.e. magmatic additions are of subordinate importance during the hyperextensional episode (e.g: Montadert *et al.* 1979b; de Charpal *et al.* 1978; Avedik *et al.* 1982; Le Pichon & Barbier 1987; Thinon *et al*. 2003). In magma-poor thermally equilibrated margins, there is a first-order relationship between "rift related residual crustal thickness", "rift related accommodation space creation" and the nature of extensional setting (low- or high-β extensional settings of Wilson *et al.* 2001). Therefore, observations and interpretations resulting from direct and indirect approaches discussed in this study cannot be applied in a straightforward way to all rifted margin settings (e.f. magma rich margins).

Indeed, in the case of magma-rich systems (e.g. conjugate margins in the Northeastern



Accommodation space evolution
Accommodation space at break up
Potential A.S.
Total A.S.
Post rift accommodation space
Sea level
Depth (km)
Distance (km)
~2-3 km
~3-4 km
~2-4 km
~0-3 km
~0-0.5 km
~1-2 km
~6 km
~6-7 km
~4-5 km
~3-6 km
~3-5 km
~1-5 km
~1-3 km
~1-4 km
>1 km
Gravity inversion results
Thinning factor
γ=1-1/β
Thinning factor (normal volcanic addition model)
Thinning factor (magma starved model)
Thickness (km)
Residual continental crust (normal volcanic addition model)
Residual continental crust (magma starved model)
Residual continental crust thickness
Extensional setting
Low-ß
High-ß
Spreading
Exhumation
Coupled
Decoupled
Decoupled
Deformation mode
Oceanic D
Exhumed Mantle D
Hyper-thinned D
Necking D
Proximal D
Oceanic crust
Serpentinized mantle
Moho (GI)
Upper CC
Lower CC
Oceanic mantle
Sub-continental mantle
Offshore / onshore criteria combined
Outcrop-scale observations
Nature of post rift sediments/aggradation
Shallow water/ poor aggration
Deep water/ large aggration
Nature of interface (basement/sediments)
Erosion
Tectonic
Magmatic
Nature of top basement
Upper continental crust
Lower continental crust
Mantle
Occurence of Allochthons

and Southern South Atlantic: for a review of volcanic margins see Geoffroy 2005 and reference therein) crustal thickness evolution during rifting is not only a function of thinning processes but also depends on magmatic additions that can be difficult to quantify. As a result, the relationship between accommodation space and crustal thinning is not straightforward and is more complex to evaluate. Additionally, late magmatic additions to the system may mask and obliterate rift related structures.

Similarly, in the case of thermally un-equilibrated margins or in the case of margins that show a strong dynamic topography, the relation between creation of accommodation space and crustal thinning may be more complex. Indeed, in order to quantify the creation of accommodation space during rifting (as in fig 6), it is necessary to estimate the dynamic topography (table 2) or it may results in completely underestimated (dynamic uplift) or overestimated values (dynamic subsidence).

Depth-dependant thinning processes are described to occur at many rifted margins worldwide (e.g. Royen & Keen 1980; Driscoll & Karner 1998; Roberts *et al.* 1997; Davis & Kusznir 2004; Kusznir & Karner 2007). In this situation, extension and thinning of rifted continental margins is partitioned with depth, i.e. the extension measured from fault heaves cannot explain the overall crustal thinning observed at the margin (Kusznir & Karner 2007). In this case, there is not necessarily a simple direct link between crustal thinning and the accommodation space created through extensional settings (low and high -β extensional settings) during the formation of the margin.

In this study we consider in-sequence rift systems, implying that the deformation is polyphase and is progressively migrating towards the location of final breakup (e. g.: Péron-Pinvidic & Manatschal 2009). Therefore, the domains described here (fig 1) represent genetic domains, i.e. recording different time steps that formed during the evolution of one rift system. Hence, in the example of early magmatic addition during rifting, some domains may not be observed (e.g. domain of exhumed mantle) or masked by late magmatic additions.

As exemplified by the Mauléon case, the reconstruction of the rift related history in

***Fig. 13**: Conclusive diagram synthetizing the geophysical quantitative (upper part) and geological qualitative (lower part) characteristics of structural domains at continental rifted margins as proposed in this study. In the central part this geological/geophysical approach is applied to aid the interpretation of an example of continental rifted margin section (Norgasis 11-12 profile) and to suggest additional geological insights at an outcrop scale (based on geological observations). This combined approach can therefore be used at a first order to suggest geological insights on rift structures, basement and sedimentary rock nature for seismic interpretations, but also to estimate crustal thickness and accommodation space in onshore fossil remnants.*

deformed analogues depends on the ability to distinguish compressive structures from the rift-related ones, but it is also necessary to unravel the different sequences of the deformation history. The restoration of the system prior to the first phase of deformation will define the initial sampling of the rift system. However, the restorations of the compressional structures only provide the original position of units relatively to each other as the horizontal movement along the thrust systems cannot be quantified precisely.

10. Conclusion

The aim of this study was to reconcile offshore and onshore type observations to further characterize structural domains at present-day continental rifted margins and in fossil examples exposed in collisional orogens. We have used the Bay of Biscay and Western Pyrenees as a natural laboratory where the access to seismically imaged, drilled and exposed fossil parts of one and the same hyperextended rift system is granted. We focused on offshore and onshore examples (Western Approach margin, Parentis and Mauléon basins) relying on the interpretation of different type of datasets with different resolutions and observational scales (from outcrop to seismic scale). For the Western Approach margin and Parentis basin examples, we use geophysical quantitative techniques (gravity inversion and flexural backstripping) combined with seismic interpretations. We estimate accommodation space, crustal thickness and lithosphere thinning and identify two types of extensional settings (low and high-β settings, Wilson *et al*. 2001). Based on the onshore Mauléon basin and drilled parts of the Parentis basin, we focus on key outcrops to describe the nature of sedimentary and basement rock and of their interface (stratigraphic, tectonic or magmatic). These observations provide necessary diagnostic elements to characterize qualitatively and quantitatively 5 domains: proximal, necking, hyperthinned, exhumed mantle and oceanic. Finally, we believe that this geological/geophysical approach illustrated in figure 13 can be used in a more general way to (1) place onshore fossil remnants into a rifted margin context resulting in first order predictions of the crustal architecture (2) to suggest geological insights on rift structures, and the nature of basement and sediments for seismic interpretations of continental rifted margins where no drill-hole data are available.

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CHAPITRE 2

Le deuxième chapitre présente une nouvelle carte et plusieurs coupes géologiques d'échelle crustale qui illustrent l'évolution 3D de l'architecture des domaines de rift liés à l'ouverture du Golfe de Gascogne et partiellement intégrés à l'orogène Pyrénéenne. L'identification de domaines de rift comparables à terre et en mer résulte d'une approche multidisciplinaire qui associe des interprétations sismiques et des résultats d'inversion gravimétrique à des observations de géologie de terrain (*chapitre 1*). L'objectif de ce chapitre est double : caractériser *l'évolution spatiale et temporelle des systèmes de rift* « *hyperamincis* » mais également *différentes étapes de leur réactivation*.

Les nouvelles observations et interprétations proposées dans ce travail révèlent une architecture de rift complexe caractérisée par *plusieurs systèmes de rift spatialement distincts, préservés à des stades d'évolution différents*. Les *mécanismes d'hyper-extension* qui caractérisent les systèmes de rift « Golfe de Gascogne-Parentis » et « Basque-Cantabre-Pyrénées » sont *diachrones*. Les variations latérales d'architecture observées semblent contrôlées par une importante segmentation partiellement héritée de la structuration pré-rift. La restauration du système à la fin de l'hyper-extension, avant l'initiation de la convergence met ainsi en avant l'existence de zones de transfert orientées NE-SO dans le système « Basque-Cantabre-Pyrénées ».

L'architecture 3D complexe du rift permet d'expliquer la réactivation hétérogène observée sur l'ensemble du domaine. Plusieurs étapes de la déformation compressive ont pu être identifiées et mises en relation avec l'architecture initiale du rift. Le domaine de *manteau exhumé* et *serpentinisé* représente une zone de faiblesse où la *réactivation* peut être *initiée*. Cette déformation se propage ensuite au domaine de croûte « *hyper-amincie* » qui va former un *prisme d'accrétion* ou être *subducté* lors de l'arrivée de la marge conjuguée. L'architecture finale de la chaîne résulte de la collision des anciens domaines *proximaux* et de *necking* (peu amincis lors du rifting) et qui vont représenter deux *buttoirs* contrôlant la déformation des domaines internes de l'orogène.

Les résultats de ce travail soulignent l'importance de l'architecture 3D des anciens domaines de rift hyper-amincis pour mieux contraindre la formation des orogènes de collision comme les Pyrénées.

Les cartes d'inversion gravimétrique et conversions profondeurs des coupes sismiques proposées dans ce chapitre ont été effectuées au cours de plusieurs séjours à l'Université de Liverpool (collaboration avec Nick Kusznir).

Cette partie du travail de thèse a été soumise à la revue *Tectonics*.

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Formation and deformation of hyperextended rift systems: insights from rift domain mapping in the Bay of Biscay-Pyrenees

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Abstract

The Bay of Biscay and Pyrenees correspond to a Lower Cretaceous rift system including both oceanic and hyperextended rift domains. The transition from preserved oceanic and rift domains in the West to their complete inversion in the East enables to study the progressive reactivation of a hyperextended rift system. We use seismic interpretations, gravity inversion results and field mapping to identify and map former rift domains and their subsequent reactivation. We propose a new map and sections across the system illustrating the progressive integration of the rift domains into the orogen. This study aims to provide insights on the formation of hyperextended rift systems and discuss their role during reactivation. Two spatially and temporally distinct rift systems can be distinguished: the Bay of Biscay–Parentis and the Pyrenean–Basque–Cantabrian rifts. While the offshore Bay of Biscay represent a former mature oceanic domain, the fossil remnants of hyperextended domains preserved onshore in the Pyrenean–Cantabrian orogen record distributed extensional deformation partitioned between strongly segmented rift basins. Reactivation initiated in the exhumed mantle domain before it affected the hyperthinned domains both of which accommodated most of the shortening. The final architecture of the orogen is acquired once the conjugate necking domains became involved in collisional processes. The complex 3D architecture of the initial rift system may partly explain the heterogeneous reactivation of the overall system. These results have important implications for the formation and reactivation of hyperextended rift systems and for the restoration of the Bay of Biscay and Pyrenean domains.

1. Introduction

The description of repeated opening and closing of oceanic basins also referred to as the "Wilson Cycle" (Wilson 1966) represents one of the main achievements of the plate tectonic theory. Plate tectonic cycle evolution suggests that mountain belts build on the former site of conjugate rifted margins and intervening oceanic domains. Therefore, the understanding of the formation and deformation of rift systems is critical to further understand plate tectonics. Over the past decades, the development of high resolution long offset seismic reflection techniques improved the imaging of the crustal architecture of rifted continental margins. These geophysical data combined with drill hole observations show that many rifted continental margins are formed by hyperextended domains consisting in extremely thinned continental crust and/or exhumed sub-continental mantle (e.g. Iberian margin: Boillot *et al.* 1987; Péron-Pinvidic & Manatschal 2009; Exmouth plateau: Driscoll & Karner 1998; West-African margin: Contrucci *et al.* 2004; Aslanian *et al.* 2009; mid-Norwegian margin: Osmundsen & Ebbing 2008).

In the meantime, on land studies in mountain belts showed that remnants of hyperextended domains could also be identified in internal parts of collisional orogens (e.g. Alps: Manatschal 2004; Mohn *et al.* 2010; Masini *et al.* 2012; Pyrenees: Lagabrielle & Bodinier 2008; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Clerc *et al.* 2012, 2013; Caledonides: Andersen *et al.* 2012). In spite of these discoveries, most studies on the formation of collisional orogens have not integrated yet the complex pre-collisional rift architecture; neglecting the implications of the former rift-related thinning and underestimating the role of rift inheritance.

In this study, we focus on the Bay of Biscay and Pyrenees one of the rare examples where both the formation and progressive reactivation of rifted margins can be investigated in one and the same system. This area corresponds to a polyphased Triassic to Early Cretaceous rift system leading to the formation of hyperextended basins and ultimately oceanic crust in the western Bay of Biscay (e.g. Vergés & García-Senz 2001 and reference therein). The late Cretaceous to Cenozoic convergence between the European and Iberian plates led to the heterogeneous deformation of the rift system, illustrated by changes in compressional architecture along strike (e.g. Muñoz 2002). Reactivation was relatively moderate along the southern Cantabrian margin (e.g. Pulgar *et al.* 1996; Gallastegui *et al.* 2002; Roca *et al.* 2011). In contrast the former hyperextended rift basins from the Pyrenean–Cantabrian domain (e.g. the Mauléon and Basque–Cantabrian basins Pedreira *et al.* 2007; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Roca *et al.* 2011) are completely inverted and integrated into the orogenic system. At present the compressional units characterizing the Pyrenean orogen are well-known (e.g. Mattauer 1968; Choukroune & Séguret 1973; Mattauer & Henry 1974; Choukroune 1992). In contrast, the rift-related paleogeographic domains remain poorly or only locally defined and their architecture is mainly described on 2D sections (e.g. Jammes *et al.* 2009; Lagabrielle *et al.* 2010). The lack of a well-defined and uniform definition of pre-compressional rift domains across the overall system results in controversial interpretations of the continuity of some pre-compressional structures and of their role during the rifting episode (e.g. the North Pyrenean fault; Canérot *et al.* 2001; Canérot 2008).

The aim of this work is twofold (1) to illustrate the complex 3D evolution of strain partitioning and architecture within hyperextended rift systems and (2) to investigate the progressive role of rift inheritance from the initiation of reactivation to continental collision. We apply for the first time a new multidisciplinary approach designed to identify diagnostic elements defining similar rift domains in offshore and onshore settings (Tugend *et al. submitted*). We combine gravity inversion results and seismic interpretations with field observations to propose a new map of rift domains from the offshore Bay of Biscay to their onshore fossil equivalents preserved in the Pyrenean orogen.

The mapping approach used in our study may be used in other orogenic systems and may bring new insights on the interpretation of the architecture of collisional orogens as well as on the restoration of the spatial and temporal evolution of fossil rift systems.

2. Geological framework

The Bay of Biscay corresponds to a V–shaped oceanic basin opened westwards towards the Atlantic Ocean. Located between France and Spain (respectively in the European and Iberian plates), it is boarded to the North by the Western Approach and Armorican margins and to the South by the North Iberian margin (fig 1). The eastern termination is characterized by several Mesozoic rift basins recording geophysical and/or geological evidence for extreme crustal thinning both offshore (e.g. the Parentis basin: Pinet *et al.* 1987; Bois & Gariel 1994; Tomassino & Marillier 1997; Jammes *et al.* 2010a, 2010c) and onshore (e.g. the Aulus basin: Lagabrielle & Bodinier 2008; Lagabrielle *et al.* 2010; Clerc *et al.* 2012; the Arzacq–Mauléon basin: Daignières *et al.* 1994; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010 and the Basque–Cantabrian basin: Pedreira *et al.* 2007, Roca *et al.* 2011).

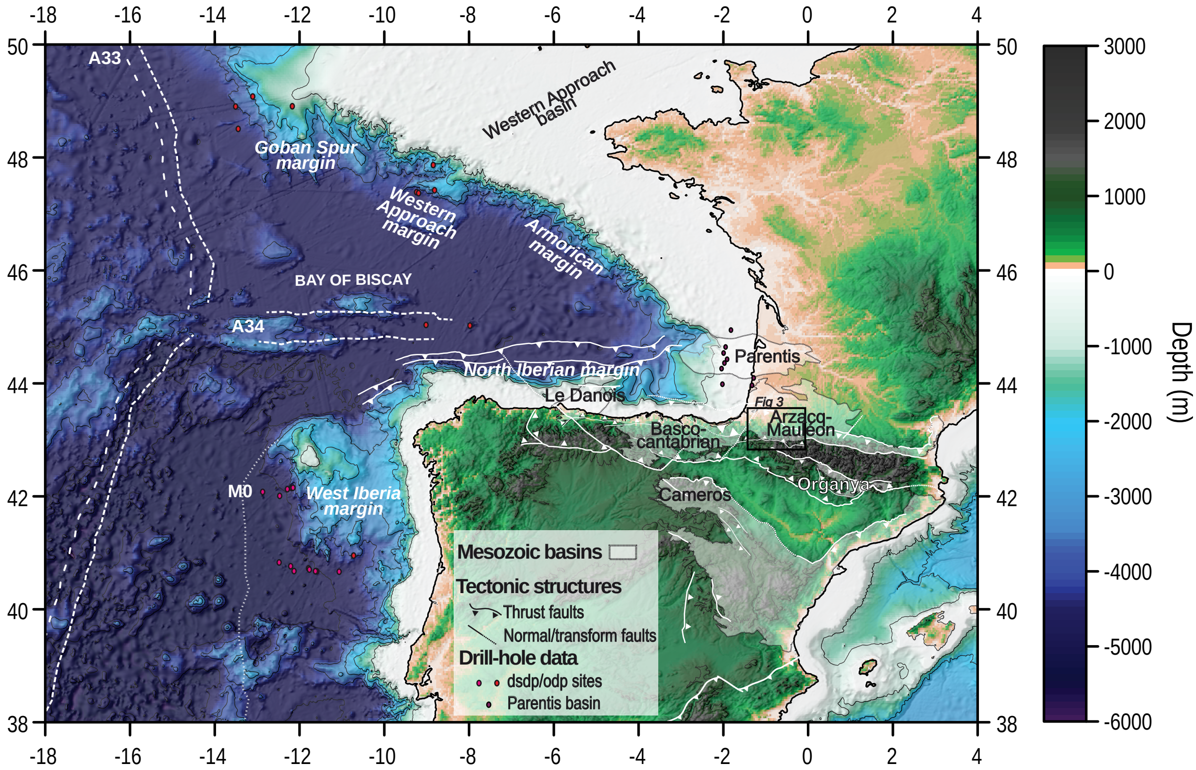


Fig. 1: *Bathymetric map of the Bay of Biscay and Pyrenees showing the major tectonic structures and the main Mesozoic basins. Magnetic anomalies are based on Sibuet et al. 2004.*

2.1. Tectonic and sedimentary evolution of the Bay of Biscay and Pyrenean domains

The Bay of Biscay and Pyrenean domains correspond to a strongly structured region that is the result of consecutive extensive and compressive tectonic cycles initiating in the early Palaeozoic. The formation of the Variscan orogen during the Carboniferous results from the collision of the Laurussia and Gondwana continental margins and intervening oceanic domains and micro-continents such as Armorica (for a review see Matte 1991, 2001). The post-Variscan evolution is related to the emplacement of strike-slip or transform faults throughout the Late Carboniferous to Early-Permian (e.g. Toulouse fault, Cévennes fault, Ventaniella fault, North Pyrenean fault) that durably structured the lithosphere. Two main scenarios are proposed for this post-Variscan evolution: one that includes a compressional episode independent of Variscan tectonics (Arthaud & Matte 1975, 1997) and a second that proposes an extensional to transtensional phase related to the post-orogenic collapse of the chain (Burg *et al.* 1994a, 1994b).

The following Triassic to Jurassic rift events resulted in the formation of intra-continental basins bounded by NE–SW trending normal faults that may partly reactivate Palaeozoic structures (e.g. Aquitaine basin: Curnelle et al. 1982). They are filled by thick sequences of siliciclastic, carbonate and evaporitic Triassic sediments (Germanic Facies: Curnelle 1983; Fréchengues 1993). In the most subsiding parts of the basins, deposits belonging to the Keuper formation are associated with tholeitic magmatism (e.g. Montadert & Winnock 1971; Winnock 1971; Rossi et al. 2003). The lattermost Triassic and Jurassic marine transgression led to the formation of a carbonate platform whose lateral extent is poorly constrained.

A major change occurred in the Late Jurassic to Early Cretaceous related to the northward propagation of the Atlantic rifting. The deposition of marine sediments in the Western Bay of Biscay is contemporaneous with the E–W reorganisation of the depocenters in the present-day Arzacq, Tarbes and Parentis basins (BRGM 1974; Biteau *et al.* 2006). Extreme crustal thinning in the Bay of Biscay resulted in continental break-up and seafloor spreading initiation during latest Aptian to early Albian time (e.g. Montadert *et al.* 1979b; Boillot 1984). At this stage, crustal and mantle exhumation in onshore rift basins (e.g. Basque–Cantabrian and Arzacq–Mauléon basins: Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Roca *et al.* 2011) is related to an acceleration of subsidence. This event is recorded by the deposition of deep marine sediments (Ducasse & Velasque 1988) intruded by alkaline magmatic rocks during Late Aptian to Early Santonian (~113 to 85 Ma; Lamolda *et al.* 1983; Montigny *et al.* 1986).

Onset of compressional deformation is recorded in Santonian sediments by a regional unconformity (Garrido-Megías & Rios 1972; McClay *et al.* 2004) and a weak reactivation in the Northern Bay of Biscay (Thinon *et al.* 2001). The major collision phase is reached during the Eocene and continues until the end of Oligocene (Muñoz 2002; Vergés *et al.* 2002) with the generalized uplift of the chain and formation of foreland basins.

Because this paper is mainly focused on the formation and reactivation of hyperextended basins, we will sub-divide the sedimentary deposits into pre-, syn-, and post-hyperextension and syn- to post-compressional sequences.

2.2. A controversial Late Jurassic to Early Cretaceous plate kinematic context

At present, there is no consensus on the Late Jurassic to Early Cretaceous kinematic evolution leading to the formation of the Bay of Biscay. An opening in a back-arc setting is proposed by Sibuet *et al.* (2004) and Vissers & Meijers (2012). Other hypotheses suggest a left-lateral strike-slip to transtensional deformation accommodated along the North Pyrenean fault or within pull-apart basins (e.g. Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Choukroune & Mattauer 1978).

Most of the plate kinematic models proposed rely on the identification and restoration of magnetic anomalies of the M-series in the North Atlantic between Iberia and Newfoundland (e.g. Olivet 1996; Sibuet *et al.* 2004). These restorations assume minor pre-breakup movements. This hypothesis is questioned by the discovery of hyperextended domains that can be mapped over hundreds of kilometres continentward of the first unambiguous magnetic anomaly related to breakup (Tucholke *et al.* 2007 or Péron-Pinvidic & Manatschal 2009). Controversies also arise from the interpretation of the age and nature of magnetic anomalies in the Bay of Biscay and the southern North Atlantic in general. The M-series (M3-M0, 126 to 118.5 Ma) identified in the Iberia–Newfoundland and Bay of Biscay margins (e.g. Sibuet et *al.* 2004) have been reinterpreted as either related to mantle exhumation (Sibuet *et al.* 2007) or magmatic underplating (Bronner *et al.* 2011 and references therein). Therefore these anomalies do not necessarily represent isochrones. These discoveries ask for a revision of existing plate kinematic models depending on the restoration of the M-series magnetic anomalies (see Bronner *et al.* 2011, 2012 and Tucholke & Sibuet 2012).

Based on these new discoveries combined with field observations, new plate kinematic scenarios were proposed. Some authors suggested that the left lateral displacement of the Iberian plate is already initiated in the Late Jurassic resulting in a transtensional setting along the European and Iberian plate (e.g. Wortmann *et al.* 2001; Schettino & Scotese 2002; Canérot 2008, Jammes *et al.* 2009, 2010a). Indirect evidence for this pre-Aptian movement comes from the thick Late Jurassic to Early Cretaceous sedimentary sequences in the Parentis and Cameros basins (e.g. Salas & Casas 1993; Mas *et al.* 1993; Jammes *et al.* 2010a). A major break occurred in Aptian time (Olivet 1996) as a result of the counter-clockwise rotation of Iberia (Gong *et al.* 2008; Jammes *et al.* 2009, 2010a). As a consequence of this rotation, a NNE–SSW to NE–SW extension is initiated in rift basins preserved onshore (e.g. Arzacq–Mauléon basin; Jammes *et al.* 2009; Lagabrielle *et al.* 2010) as emphasized by the NE–SW segmentation observed (e.g. Pamplona, Toulouse, Cevennes faults). These NE–SW transfer faults cross the Iberian-European plate boundary and are sealed by Albian sediments (Pamplona fault, between the Basque–Cantabrian and Arzacq–Mauléon basins; Razin 1989; Claude 1990). In this paper, we will show that the complex polyphased history strongly influenced the 3D architecture of the Bay of Biscay-Parentis and Pyrenean-Basque-Cantabrian rift systems.

2.3. Pyrenean reactivation and shortening estimations

The paleogeographic evolution of the Bay of Biscay and Iberia–Newfoundland rifted margins is relatively well-constrained after the quiet magnetic period of the Cretaceous and the identification of the magnetic anomaly 34 (83 Ma, Santonian) (e.g. Roest & Srivastava 1991; Rosenbaum *et al.* 2002). The northward movement of the African plate during the Late Cretaceous results in compression initiation in the Bay of Biscay and Pyrenees. Although shortening direction is reasonably coherent between different kinematic models (N–S to NE–SW; respectively Roest & Srivastava 1991; Rosenbaum *et al.* 2002), the total amount of Late Cretaceous shortening relative to Cenozoic shortening is still a matter of debate. At the scale of the Bay of Biscay and Pyrenean–Cantabrian orogen, the transition from an embryonic subduction in the North Iberian margin to a continental collision in the Pyrenees is interpreted as the result of increase of convergence from west to east. Based on magnetic anomaly restorations, Rosenbaum *et al.* (2002) estimated about 144 km of shortening in the Western Pyrenees compared to 206 km in the Eastern Pyrenees. Classical palinspastic restorations of constructed sections across the Pyrenees usually propose restorations of the thick-skin deformation observed in the Axial zone (e.g. Teixell 1998; Muñoz 2002). The results of this study suggest that this approach is unable to decipher the overall amount of convergence accommodated in the Pyrenean domain.

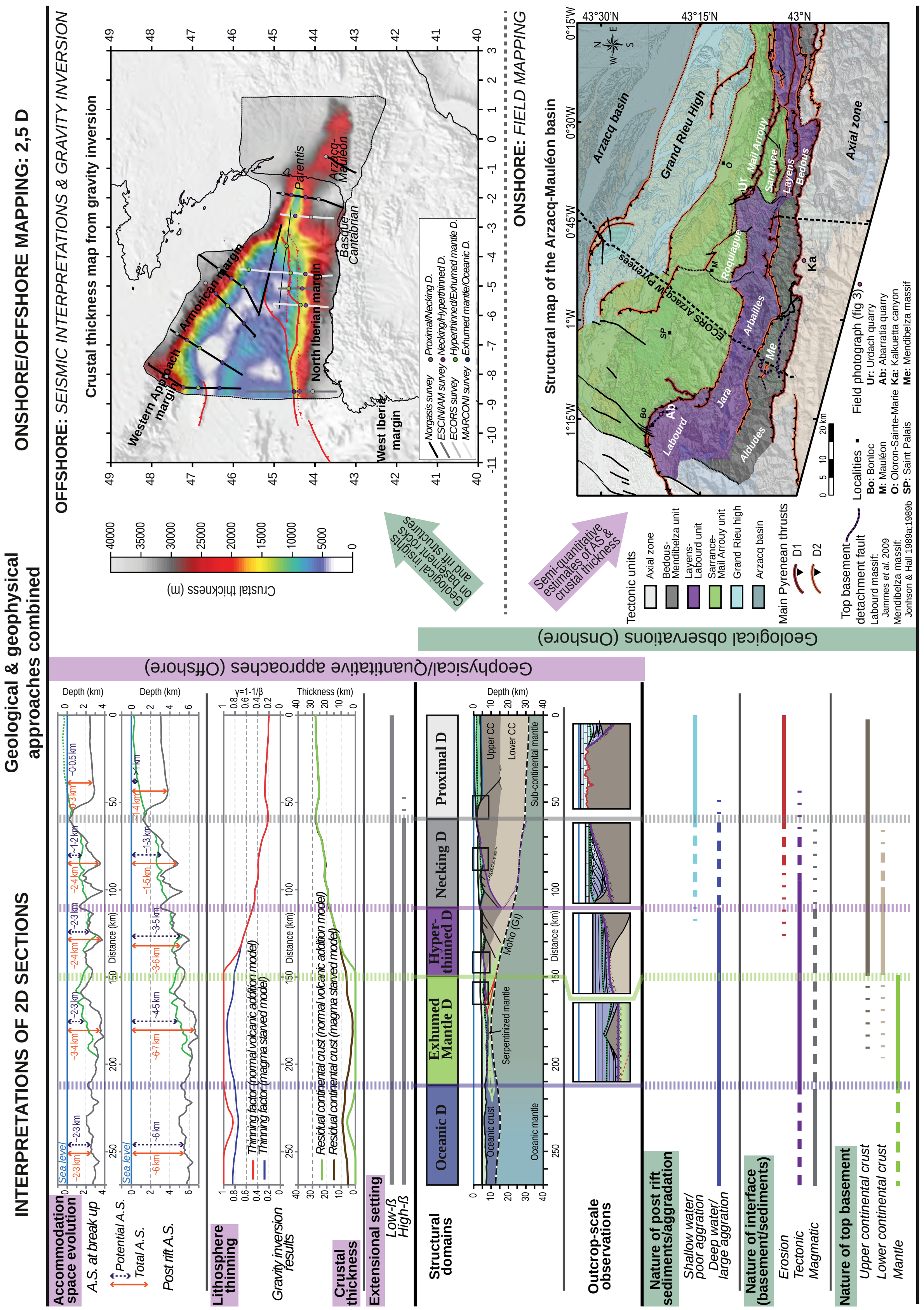
3. Mapping hyperextended domains: combining onshore and offshore observations

3.1. Hyperextended rift domains: terminology and identification

In spite of variable magmatic, structural and sedimentary evolutions, depending on the amount and rates of extension, thermal history, but also on the inheritance, most magma-poor rifted margins share comparable large-scale architectures (Reston 2009; Péron-Pinvidic *et al.* 2013). Using first-order geological and/or geophysical observations, distinctive structural rift domains can be identified. From continent to ocean the following domains can be distinguished: the proximal, necking, hyperthinned, exhumed mantle and oceanic domains (fig 2). Extensional deformation related to the formation of these domains is polyphased, progressively localizing and migrating towards the location of final break-up (e.g. Péron-Pinvidic & Manatschal 2009; Sutra *et al.* 2013). Therefore, these domains reflect successive steps in the formation of magma-poor rifted margins, suggesting that they may also correspond to genetic domains.

We use the approach developed by Tugend *et al. (submitted,* see chapter 1) enabling the characterization and identification of comparable rift domains in present-day magma-poor rifted margins and their fossil analogues preserved in collisional orogens (fig 2). Offshore, quantitative techniques provide estimations of accommodation space, crustal thickness and lithosphere thinning while seismic interpretations enable the recognition of extensional settings (low- and high-β settings; Wilson *et al.* 2001). Onshore, the identification relies on the description of key outcrops preserving the nature of sedimentary and basement rocks and of their interface. This geological/geophysical approach can be further used as an interface between onshore and offshore observations and to suggest analogies. Offshore seismic interpretations can take advantage of onshore observations on the nature of sediment, basement and of their interface. The large scale geometry and stratigraphic architecture imaged offshore can be used to restore onshore fossil remnants back into a rifted margin context (fig 2).

Fig. 2: *Philosophy of the onshore/offshore approach applied in this study. Left side: synthetic diagram combining geophysical (upper part) and geological (lower part) diagnostic elements of structural/ genetic rift domains at continental rifted margins (modified after Tugend et al. subm). The terminology used in this study is indicated in the central part. Right side: this approach is used to map rifted domains offshore (gravity inversion results and seismic interpretations) and onshore (structural map of the Mauléon basin and of the southern part of the Arzacq basin and field observations).*

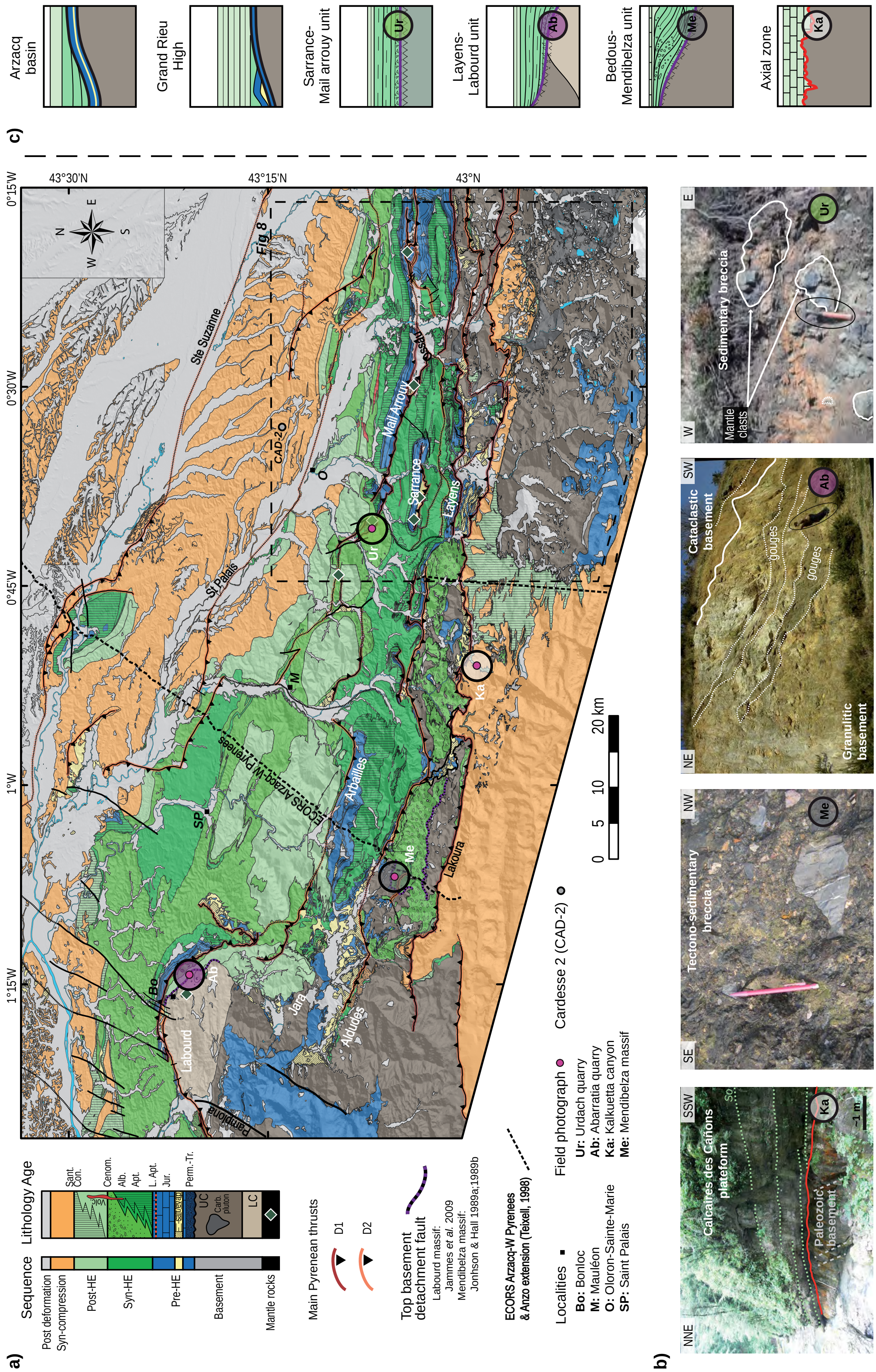


This combined approach is applied for the first time to map the overall rift system from the offshore Bay of Biscay to its onshore fossil analogues preserved in the Pyrenees (including the Basque–Cantabrian basin). Interpretations of structural domains along offshore seismic reflection profiles are combined with 3D mapping of crustal thickness, lithosphere thinning and Moho depth resulting from gravity inversion. Onshore mapping relies on the identification of remnants of the rift system preserved within well-defined compressive tectonic units.

3.2. Onshore: Mapping remnants of the rift system

The mapping of rift domains in onshore deformed analogues depends on the ability to decipher the deformation and sedimentary history. The definition of compressive units is based on the identification of first order compressive structures corresponding to the first generation of thrusts. These coherent tectonic units are usually formed by major thrust contacts, but are often only weakly deformed internally. The methodology used to map rift domains onshore is here exemplified by the example of the Arzacq–Mauléon basin. A summary of main observations is given below and summarized in fig 3 (detailed descriptions of field observations in Jammes *et al.* 2009; Tugend *et al. submitted*, see chapter 1).

Fig 3: *(a) Geological map of the Mauléon basin and the Southern part of the Arzacq basin. Lithologies and age of sequences are synthetized in a log. (b) Field photographs of the main observations. Ka: Kalkuetta canyon, contact between the "Calcaires des Cañons" platform and underlying eroded Palaeozoic basement (0°50'59,72"W/42°59'17,31"N). Me: Mendibelza massif, tectono-sedimentary breccia at the contact with the top basement detachment fault (1°6'55,78"W/43°5'40,16"N). Ar: Abarratia quarry, deformation of the granulitic basement (gouges, cataclastic overprint) charactering the top basement detachment fault described by Jammes et al. 2009 (1°14'17,69"W/43°20'55,64"N). Ur: Urdach quarry, syn-rift breccia reworking mantle and basement rocks (0°40'39,62W/43°7'10,91"N). (c) Simplified cartoons of the key observations from remnants of the former rift system preserved in the Pyrenean nappe stack. Axial zone unit: remnant of the proximal domain. Bedous-Mendibelza unit: remnant of the Necking domain. Layens-Labourd unit: remnant of the hyperthinned domain. Sarrance-Mail arrouy unit: remnant of the exhumed mantle domain. Grand Rieu High: remnant of the hyperthinned domain. Arzacq basin: transition from proximal to necking domain. The map is simplified after the BRGM (1/50000) geological map of: Arthez de Béarn, Argelès-Gazost, Gavarnie, Hasparren, Iholdy, Laruns-Somport, Larrau, Lourdes, Oloron-Sainte-Marie, Orthez, Mauléon-Licharre, Morlaàs, Pau, Saint-Jean-Pied-de-Port, Tardets (see references in the bibliography, authorization n°: 2011-071)*


a)
Sequence
Lithology Age
Post deformation
Syn-compression
Post-HE
Syn-HE
Pre-HE
Basement
Mantle rocks
Sant.
Con.
Cenom.
Alb.
Apt.
L. Apt.
Jur.
Perm.-Tr.
UC
Carb. pluton
LC
Main Pyrenean thrusts
D1
D2
Top basement detachment fault
Labourd massif:
Jammes et al. 2009
Mendibelza massif:
Johnson & Hall 1989a;1989b
ECORS Arzacq-W Pyrenees & Anzo extension (Teixell, 1998)
Localities
Bo: Bonloc
M: Mauléon
O: Oloron-Sainte-Marie
SP: Saint Palais
Field photograph
Ur: Urdach quarry
Ab: Abarratia quarry
Ka: Kalkuetta canyon
Me: Mendibelza massif
Cardesse 2 (CAD-2)
0 5 10 20 km
43°30'N
43°15'N
43°N
0°15'W
0°30'W
0°45'W
1°W
1°15'W
Ste Suzanne
St Palais
Fig 8
CAD-2
Mail Arrouy
Sarrance
Layens
Arbailles
ECORS Arzacq-W Pyrenees
Lakoura
Labourd
Jara
Aldudes
Pamplona
c)
Arzacq basin
Grand Rieu High
Sarrance-Mail arrouy unit
Layens-Labourd unit
Bedous-Mendibelza unit
Axial zone
b)
Sedimentary breccia
Mantle clasts
Cataclastic basement
gouges
Granulitic basement
Tectono-sedimentary breccia
Calcaires des Cañons plateform
Paleozoic basement
~1 m

3.2.1. Defining coherent tectonic units: history of compressional deformation

In the Arzacq–Mauléon basin different phases of deformation can be identified resulting in two opposite directions of thrusting. The shortening related to the initiation of Pyrenean compression is accommodated by Late Cretaceous south-directed thrust systems (e.g. Sarrance, Mail Arrouy or Lakoura thrust systems; Teixell 1998; figs 2 and 3). This initial sampling of the distal parts of the former rift system is locally overprinted by a second phase of deformation resulting in north-directed thrusts (e.g. the Saint Palais, Sainte Suzannne and Ossau thrusts; Canérot *et al.* 2001) and south-directed back-thrusting in the Axial zone (e.g. Gavarnie, Guarga thrust systems; Teixell 1998). As a result of this structural analysis several tectonic units can be defined within the Arzacq–Mauléon basin (fig 2). Within each unit key outcrops preserving primary contacts between basement and the pre-, syn-, and post-hyperextensional sediments can be found (e.g. Canérot 2008; Jammes *et al.* 2009, Lagabrielle *et al.* 2010; Debroas *et al.* 2010).

3.2.2. Outcrops preserving remnants of the former rift system

Remnants of rift domains can be defined in the Arzacq-Mauléon basin (fig 2 and 3) from South to North.

In the axial zone (southernmost external Pyrenean unit) – former proximal rift domain, upper crustal rocks from the basement are eroded and directly onlapped by shallow marine post-hyperextensional sediments (fig 3) suggesting no or only a weak rift-related thinning.

In the Bedous–Mendibelza unit – former necking domain, the exhumed basement of the Mendibelza and Igounze massifs (Johnson & Hall 1989a, 1989b; Miranda-Avilez *et al.* 2005) is onlapped by syn-hyperextension sequences progressively thickening and deepening northwards. This domain marks a progressive transition from shallow to deep-water sediments (conglomerates from the Mendibelza formation, passing laterally to turbidites and marls) as a consequence of the initiation of crustal thinning.

In the Layens–Labourd unit – former hyperthinned domain, granulites derived from the pre-hyperextension middle crust (Vielzeuf 1984) are exhumed in the Labourd massif (Jammes *et al.* 2009). They are onlapped by late syn-to post-hyperextension sediments reworking pieces of exhumed granulite (Bonloc breccias; Claude 1990). Observations from this domain indicate that at least locally the whole upper crust was removed and consequently the crust had to be thinned.

In the Sarrance–Mail Arrouy unit (internal orogenic unit) –former exhumed mantle domain, numerous outcrops of mantle rocks are observed, some of which are reworked in Cenomanian breccias (fig 3; e.g. Urdach breccias: Jammes *et al.* 2009; Lagabrielle *et al.* 2010;

Debroas *et al.* 2010). Syn-hyperextension sediments are near the city of Mauléon more than 7 km thick (Roux 1983; Fixari 1984; Souquet *et al.* 1985). At least locally, mantle rocks have been exhumed resulting from a complete thinning of the continental crust.

In the Grand Rieu high unit – former hyperthinned domain, a thick post-hyperextension sequence may directly onlap onto basement as indicated by the Cardesse 2 drill-hole (location fig 3; BRGM 1974; Serrano *et al.* 2006). This local observation may reflect a large sedimentary aggradation consecutive to post-hyperextension thermal subsidence in a former hyperthinned domain (fig 2).

In the Arzacq basin –former proximal to necking domain: the progressive thinning of the continental crust (from about 25 km to the north to about 20 km below the Grand Rieu high; Daignières *et al.* 1982, 1994) is concomitant with a southward thickening of syn-hyperextension sequences (Biteau *et al.* 2006). This progressive creation of accommodation space may indicate the progressive transition from a proximal to a necking domain (from north to south).

The mapping of former rift domains onshore relies on the exposed surface observations (here locally supported by drill-hole data) and reflects the integration of the former rift into the orogenic system.

3.3. Offshore: seismic interpretations and gravity inversion mapping

Gravity inversion results have already been used to characterize and identify structural domains on 2D sections at rifted continental margins (fig 2; Tugend *et al., submitted*, see chapter 1). In this paper, the aim is to associate this technique with local seismic interpretations to map and unravel the spatial evolution of structural/genetic rift domains from the offshore Bay of Biscay to the onshore Mesozoic Arzacq basin buried under post-hyperextension sediments and the thick syn-compressional sequences of the Tertiary Aquitaine basin.

3.3.1. Gravity inversion: scheme and datasets

Gravity inversion is based on public domain data (fig 4): free air gravity (Sandwell & Smith 2009), bathymetry (Smith & Sandwell 1997) and oceanic isochrones (Müller *et al.* 1997). Information on sedimentary thickness comes from a compilation of offshore seismic interpretations combined with the depth to basement map of the Aquitaine basin from Serrano *et al.* (2006) (fig 4). Seismic reflection data are derived from different surveys: the Norgasis survey (Avedik *et al.* 1993, 1996; Thinon 1999), the ECORS Bay of Biscay section (Pinet *et al*. 1987; Bois & Gariel 1994), the IAM 12 and the ESCIN 4 seismic lines (e.g. Banda *et al.* 1995; Gallastegui *et al.* 2002; Gallart *et al.* 2004).

Moho depth, crustal thickness, continental lithosphere thinning factor and residual continental crust maps (fig 5) were produced by gravity inversion. This technique includes a thermal gravity anomaly correction and a parameterization of decompression melting to predict volcanic additions (detailed scheme described by Greenhalgh & Kusznir 2007; Chappell & Kusznir 2008; Alvey *et al.* 2008; parameters for this study are presented in table 1). A compaction controlled density-depth relationship is applied to sedimentary sequences.

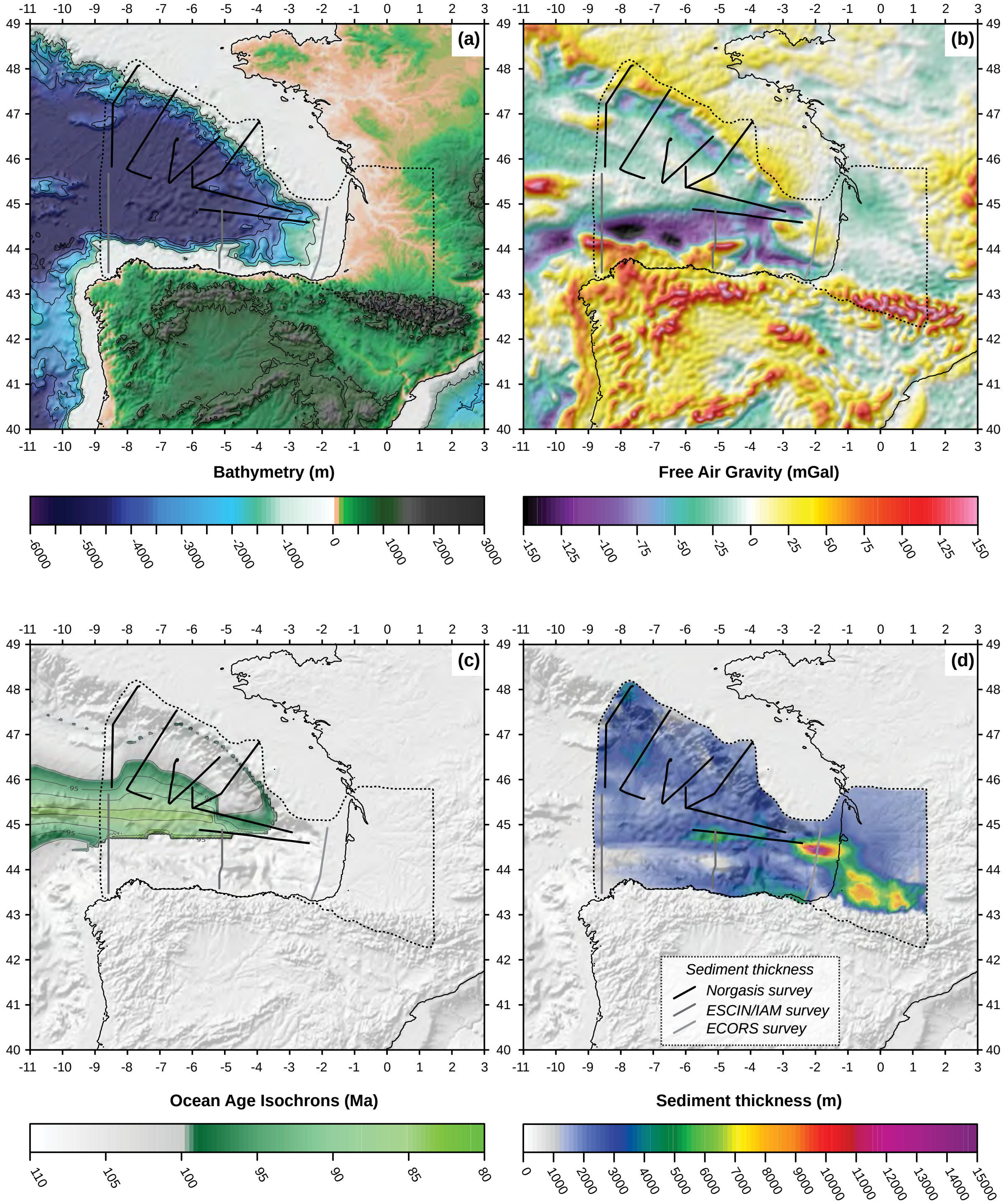


Fig 4: *Raw data used for the gravity inversion (a) Bathymetry and topography (Smith & Sandwell 1997) (b) Free-air gravity (Sandwell & Smith 2009) (c) Oceanic age isochrones (Müller et al. 1997) (d) Sediment thickness is derived from offshore seismic interpretations (Raw seismic lines from the Norgasis survey: Thinon 1999; Avedik et al. 1993; 1996; ECORS Bay of Biscay: Pinet et al. 1987; Bois & Gariel 1994; IAM 12 and ESCIN 4 seismic lines: Gallart et al. 2004; Gallastegui et al. 2002; Banda et al. 1995) and the depth to basement map of the Aquitaine basin from Serrano et al. 2006.*

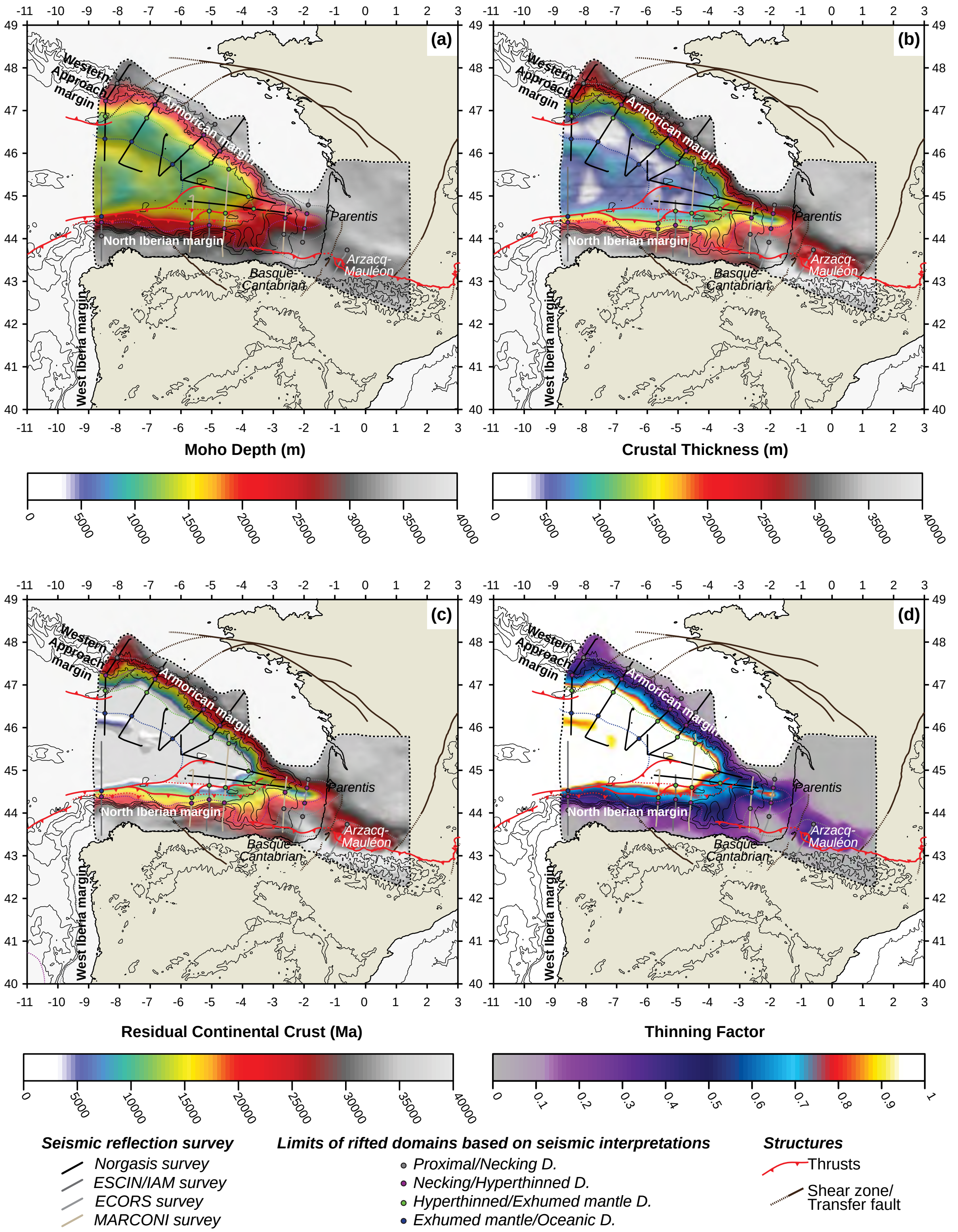


Fig 5: *(a) Moho depth, (b) crustal thickness, (c) residual continental crust and (d) continental lithosphere thinning factor maps resulting from gravity inversion. The limits of rifted domains determined from seismic interpretations are also indicated.*

Table 1: *Parameters used for gravity inversion*

Parameters	Value and reference dataset
Critical thinning factor (γ)	0,7
Reference crustal thickness	40 km
Break up age	110 Ma
Volcanic additions	
Standart thickness	7 km
Magma-starved	0 km

3.3.2. 3D mapping derived from gravity inversion and seismic interpretations

The maps resulting from gravity inversion presented in fig 5 were computed assuming a “normal” volcanic addition (corresponding to a standard oceanic crust thickness of about 7 km; White *et al.* 1992). This solution assumes the occurrence of decompression melting during rifting once the lithosphere is thinned below a critical value ($\gamma = 0.7$; table 1), producing magmatic additions to the extended continental crust. In the case of a magma-poor rifted margin as proposed for the Bay of Biscay (e.g. De Charpal *et al.* 1978; Montadert *et al.* 1979b; Avedik *et al.* 1982; Le Pichon & Barbier 1987; Thinon *et al*. 2003) the transition zone between unequivocal oceanic and continental crusts may be interpreted as exhumed mantle (fig 2). The characterisation of structural domains on 2D sections (fig 2) partly relies on diagnostic elements determined from gravity inversion results (Tugend *et al. submitted*, see chapter 1). The combined analysis of Moho depth, crustal thickness and lithosphere thinning maps (fig 4) can be further used to delimit offshore structural domains and identify their evolution in 3D.

The proximal domain presents relatively constant values of lithosphere thinning (fig 5d) and a smooth Moho topography (fig 5a), characterizing its limited crustal and lithosphere thinning. The rapid increase in lithosphere thinning values coincides with a Moho rise delimiting the proximal from the necking domain (fig 5a/d). The transition from the necking to hyperthinned domain can be identified on seismic sections (fig 2; e.g. Sutra *et al.* 2013) and corresponds in this example to thinning values between 0.5 to 0.6 and crustal thicknesses of about 10 to 15 km (fig 5b/c). The beginning of the exhumed mantle domain can be interpreted at a first order at the termination of the continental crust (fig 5c). Moho topography derived from gravity inversion (Greenhalgh & Kusznir 2007; Chappell & Kusznir 2008; Alvey *et al.* 2008) is slightly curved through this domain enabling the shallowest Moho depth to be reached (fig 5a). Gravity inversion provides indirect observations compatible with the existence of an exhumed mantle domain in the Bay of Biscay. This hypothesis is further supported by reflection and refraction seismic interpretations (e.g. Fernández-Viejo *et al.* 1998; Thinon *et al.* 2003; Ruiz 2007) and by analogy with the existence of such a domain onshore (Arzacq–Mauléon; fig 3). The oceanic domain is marked by a deepening of Moho depth (fig 5a) and a thickening of the crust at the oceanward edge of the exhumed mantle domain (fig 5b). Residual patches of thicker crust are observed within the oceanic domain (fig 5 c/d) and can be interpreted as a locally overthickened oceanic crust (in comparison with the 7 km thick volcanic addition used for gravity inversion; table 1).

The resulting maps show the present day distribution of oceanic and continental crust as well as a possible extent of exhumed mantle in the Bay of Biscay. In areas that underwent reactivation and crustal thickening (e.g. the North Iberian margin), these maps need to be interpreted carefully and are further constrained by seismic interpretations. Onshore, in the Western Pyrenees, the former rift-related thinning of the northern Arzacq basin can still be deduced from lithosphere thinning and crustal thickness maps (fig 5).

4. Map of the Bay of Biscay–Pyrenean rift system (fig 6)

Offshore and onshore observations are combined to recognize, describe and map structural/genetic rift domains and their compressional overprint from the Bay of Biscay to the Pyrenean-Cantabrian orogen (fig 6). We apply the terminology and geological/geophysical approach described in the previous section and synthetized in fig 2.

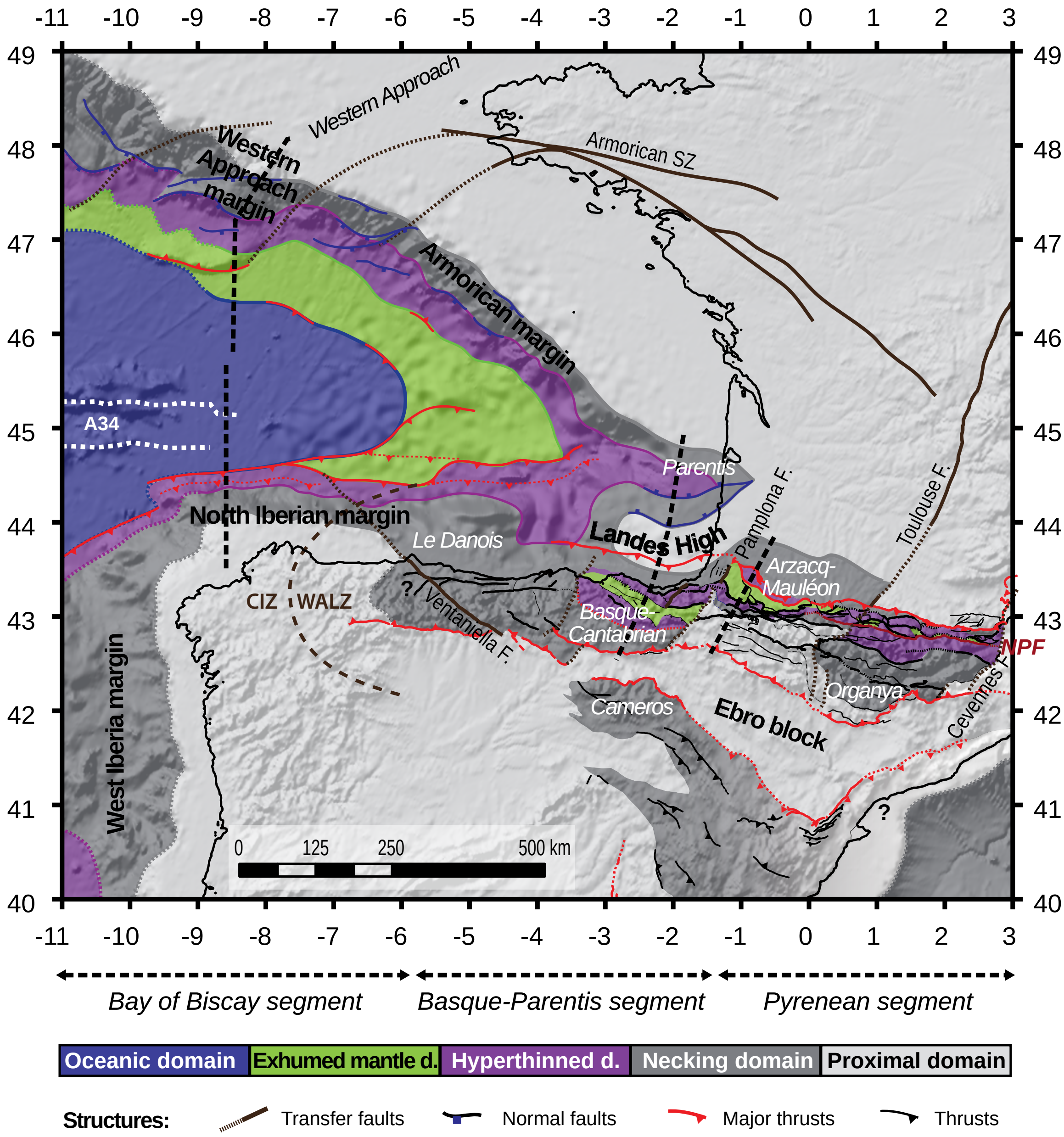


***Fig 6:** Map of the rift domains preserved in the Bay of Biscay and their fossil analogues from the Pyrenean domain. Extensional rift structures and thrust faults from the Armorican and Western Approach margins are based on the work of Thinon 1999 and own observations. CIZ: Central Iberian zone. WALZ: West Asturian-Leonese zone. NPF: North Pyrenean fault. The location of the geological sections from fig 7 is indicated in dashed line.*

4.1. Definition of rift domains

4.1.1. Proximal domain

Proximal domains are characterized by only weak to no rift-related lithosphere and crustal thinning (fig 5). Creation of accommodation space is restricted to graben and half-graben basins characterizing low-β extensional systems (fig 2). The sedimentary record of this domain may include sub-aerial exposure, continental to shallow water sedimentary systems, and no or minor aggradation of post-rift sequences (figs 2 and 3).

Offshore, this domain corresponds to the continental shelves (e.g. the Armorican platform, the western part of the North Iberian shelf, near Ortegal Spur, Landes high; fig6). Onshore, it can be mapped in external parts of the orogen and may be partly buried below Tertiary foreland basins such as the Duero basin (south of the Basque–Cantabrian orogen), the Jaca and Ebro basins south of the Pyrenees or in the northern part of the Aquitaine basin (fig 6).

4.1.2. Necking domain

Lithosphere and crustal thinning are initiated in the necking domain (fig 5). The oceanward increasing accommodation space and deepening of the top basement coincides with the ascending Moho topography defining a taper geometry (Osmundsen & Redfield 2011). The necking domain is limited oceanward by the coupling point ("taper break" in Osmundsen & Redfield 2011) interpreted as the transition from decoupled to coupled deformation at a crustal scale (Sutra *et al.* 2013; "crustal embrittlement" in Pérez-Gussinyé & Reston 2001; Reston & Pérez-Gussinyé 2007). Tilted block geometries and/or exhumation surfaces related to top basement detachment faulting can be observed (fig 2). The sedimentary architecture can often be defined by offlapping packages and onlapping backwards onto the basement of the necking domain (outcrop scale observations; fig 2). The progressive creation of accommodation space can be recorded by deltaic sedimentary systems or slope facies including gravitational systems. A transition from shallow to deeper marine environments can also be observed in underfilled basins (fig 2).

In the Bay of Biscay, the necking domain corresponds to the continental slope of the Western Approach (over the Meriadsek Terrace), Armorican and North Iberian margins (including the "Le Danois basin" in the eastern part; fig 6). The southern part of the Basque–Cantabrian basin, the Organya and Perdraforca basins (as suggested by the restorations of Muñoz 1992; Vergés & Garcia-Senz 2001; Lagabrielle *et al.* 2010) and most of the Arzacq basin may represent fossil analogues of this domain (fig 6). Narrow remnants of this domain can also be mapped in basins from the eastern and western Pyrenees (respectively south of the North Pyrenean Frontal thrust and north of the Lakoura thrust).

4.1.3. Hyperthinned domain

Continental crust is typically less than 15 to 10 km thick and associated with a large accommodation space (figs 2 and 5). The complete crustal embrittlement (Pérez-Gussinyé & Reston 2001; Reston & Pérez-Gussinyé 2007) of the hyperthinned domain is accompanied by a polyphase deformation indicated by the rejuvenation of syn-tectonic sediments oceanwards (e.g. Péron-Pinvidic *et al.* 2007, Sutra *et al.* 2013). Both low and high-β extensional settings may be observed and are characterized by half-graben and hyperextended sag basin architectures (fig 2 in Tugend *et al. submitted*, see chapter 1). Top basement detachment faults commonly lead to local exhumation of mid to lower crustal levels as observed in onshore fossil remnants (e.g. Manatschal 2004; fig 2). The infilling history of this domain mainly depends on sedimentary supply and sedimentary sources but is usually associated to thick aggradational sequences and/or deep marine sedimentary sequences (fig 2).

The occurrence of extremely thinned continental crust was already recognized in the Parentis basin (e.g. Pinet *et al.* 1987; Bois & Gariel 1994; Tomassino & Marillier 1997) and at the toe of the continental slope of the Armorican and Western Approach margins ("neck area" in Thinon *et al.* 2003). In the southern Bay of Biscay, this domain is integrated into the accretionary prism of the North Iberian margin, as indicated by refraction velocities (e.g. Ruiz 1997; Gallart *et al*. 2004; Roca *et al.* 2011; Fernández-Viejo *et al.* 2012). Onshore in the Pyrenees (fig6), this domain is characterized by numerous granulitic rocks (e.g. in the Labourd, Trois Seigneurs, Agly massif) interpreted as mid to lower crustal rocks (e.g. Vielzeuf 1984) exhumed during Cretaceous rifting (e.g. Jammes *et al.* 2009; Vauchez *et al.* 2013).

4.1.4. Exhumed mantle domain

The beginning of the exhumed mantle domain can be observed once the complete thinning of the continental crust is achieved (fig 5c/d). Local remnants of continental crust may be preserved as extensional allochthons on top of the exhumed mantle basement of this domain (outcrop scale observation; fig 2). The seismic reflection pattern and velocity structure described from suspected and drilled exhumed mantle domains contrasts with the adjacent hyperthinned and oceanic domains. Seismic reflection observations show a complex set of deep intra-basement reflections (e.g. Armorican margin: Thinon *et al.* 2003; Iberian margin: Pickup *et al.* 1996; Dean *et al.* 2001; Reston & McDermott 2011; Sutra *et al.* 2013). Refraction studies indicate a downward gradient of velocities commonly interpreted as the progressive decrease in serpentinization with depth (e.g. Minshull 2009; Reston 2009, 2010). Magmatic additions can also be observed within this domain, as indicated by the local occurrence of syn- to post-hyperextension alkaline magmatism in onshore remnants of exhumed mantle in the Pyrenees. Offshore, volcanic bodies and volcanoclastic sediments are interpreted on the Armorican margin (Thinon *et al.* 2003). This magmatic overprint may progressively become more important toward the oceanic domain. Hyperextended sag basins are often observed in the exhumed mantle domain, sometimes extending to the hyperthinned and necking domains in the case of overfilled basins (e.g. the Angola margin; Unternehr *et al.* 2010).

The existence of exhumed mantle has already been proposed to floor the Armorican basin, (Thinon *et al.* 2003) and suggested in the South Iberian margin (Roca *et al.* 2011). Gravity inversion results combined with seismic interpretations enable us to refine the eastern termination towards the hyperthinned Parentis basin (fig 6). Onshore, this domain is characterized by the occurrence of mantle bodies and Mesozoic metamorphic rocks included in tectonic units near the North Pyrenean fault in the Pyrenees (fig 6; e.g. Monchoux 1970; Fabriès *et al.* 1991, 1998; Lagabrielle & Bodinier 2008; Lagabrielle *et al.* 2010; Clerc *et al.* 2012, 2013) or within the so-called Nappe des Marbres in the Basque–Cantabrian basin (fig 6; Lamare 1936; Mendia & Gil-Ibarguchi 1991).

4.1.5. Oceanic domain

The transition to the oceanic domain is characterized by a deepening of Moho depth and the ramp morphology of the basement referred to as outer high (Péron-Pinvidic & Manatschal 2010; Péron-Pinvidic *et al.* 2013). Refraction data across this basement high in the Bay of Biscay indicate the presence of an underplated high velocity body originating in the exhumed mantle domain (Thinon *et al.* 2003). This velocity structure may suggest a possible overlap between the exhumed mantle and oceanic domains (see also Roca *et al.* 2011). It may be interpreted as an underplated gabbroic body similarly to the interpretation of Bronner *et al.* (2011) for the Iberian–Newfoundland rifted margins. It may also represent the oceanward termination of serpentinized mantle (Thinon *et al.* 2003) that is overprinted by the first oceanic crust. Both interpretations suggest a possible overlap or transition zone between the exhumed mantle and oceanic domains. A specific seismic pattern is also observed, characterized by diffractive and irregular reflections in the upper basement (Thinon *et al.* 2003) and locally strong reflections parallel to the top basement at deeper levels (Sutra *et al.* 2013). The first onlapping and overlying sediments are of post-rift age. In the central part of the Bay of Biscay, one drill-hole (DSDP site 118, leg 12; Laughton *et al.* 1972) reached the oceanic basement. This domain is characterized by E–W to ESE–WNW trending magnetic anomalies (Matthews & Williams 1968) attributed to the magnetic anomaly 34 (e.g. Srivastava *et al.* 1990). The absence of the magnetic anomaly 33 (Cande & Kristoffersen 1977) suggests that seafloor spreading stopped after anomaly 34 and before anomaly 33, corresponding to Santonian–Early Campanian time (e.g. Montadert *et al.* 1979a; 1979b). No remnants of oceanic crust (ophiolites) can be observed onshore in the Pyrenean or Basque Cantabrian basins (Lagabrielle & Bodinier 2008).

4.2. Key structures and lateral extend of domains

The definition of rift domains also relies on the identification of key structures as they represent markers of the deformation during the different extensional and compressional events.

4.2.1. Compressional systems

The overall WNW to ESE trend of compressional structures (foreland basins, boundary between internal and external domains) results from the Iberian/European convergence. Compressional deformation is important in the Southern Bay of Biscay and controls the evolution of the North Iberian frontal thrust system (e.g. Derégnancourt & Boillot 1982; Roca *et al.* 2011; Fernández-Viejo *et al.* 2011, 2012). In its eastern termination it represents the boundary between continental and oceanic domains without evidence for oceanic subduction (Ayarza *et al.* 2004; refraction data along the IAM 12 line; Fernández-Viejo *et al.* 1998). To the West, it delimits the accretionary prism and reactivation in the exhumed mantle domain (Roca *et al.* 2011; Fernández-Viejo *et al.* 2012). Locally NW–SE trending thrusts are observed in the Northern Bay of Biscay mainly located at domain transitions (Thinon *et al.* 2001).

Most of the compressive structures can be observed onshore. The North and South Pyrenean frontal thrusts delimit the almost E–W trending Pyrenean–Cantabrian orogen from the adjacent foreland basins. Onshore remnants of rift domains are delimited by thrusts systems. Among them, the North Pyrenean fault (e.g. Mattauer 1968; Choukroune & Mattauer 1978) and Basque–Cantabrian equivalent, the Leiza fault (Boillot *et al.* 1973; Choukroune 1976; Rat 1988; Combes *et al.* 1998; Mathey *et al.* 1999) can be mapped discontinuously.

4.2.2. Extensional faults

NW–SE to E–W trending extensional structures can be mapped in the Northern Bay of Biscay (Derégnancourt & Boillot 1982; Thinon 1999; Thinon *et al.* 2003) related to the rifting phase in the Bay of Biscay. Top basement detachment faults may initiate at the boundary between the proximal and necking domains (e.g. Southern boundary of the Parentis basin, Jammes *et al.* 2010b, 2010c). The hyperthinned domain of the Armorican margin is mainly characterized by rare small offset normal faults (low-β extensional settings; fig 2) as described by Thinon (1999) and Thinon *et al.* (2003). Remnants of extensional rift structures and top basement extensional detachment faults are described locally onshore in the Mauléon basin (e.g. Canérot *et al.* 1978; Canérot 1989; Johnson & Hall 1989a, 1989b; Jammes *et al.* 2009; Masini 2011), in the Aulus basin (e.g. Lagabrielle *et al.* 2010) or in the Basque–Cantabrian basin (e.g. Tavani & Muñoz 2012; Bodego & Agirrezabala 2013).

4.2.3. Transfer/transform faults

A striking observation resulting from this mapping is the importance of transfer faults. These structures strongly segment offshore and onshore rift systems. The segmentation of the Northern Biscay margin is characterized by NE–SW transfer faults (Derégnancourt & Boillot 1982; Thinon *et al.* 2003) that may be partly related to the Variscan structuration observed onshore (Thinon 1999). The offshore prolongation over the continental shelf is underlined by magnetic and gravity trends and is suggested to be partly controlled by the Armorican shear zone (e.g. Martínez-Catalán *et al.* 2012; Ballèvre *et al.* 2012 and reference therein). The influence of the inferred structures can be observed on the segmentation of the limit between necking and exhumed mantle domains between the Goban Spur, Western Approach and Armorican margins (fig 6).

Second order inherited Varican structures locally break the NW–SE straight orientation of the southern part of the Armorican margin (Thinon 1999), but the control of these structures on the subsequent rift structures is complex and not directly observable. In contrast, the South Iberian margin is segmented by NW–SE structures as for instance the Ventaniella fault that records a Permo-Carboniferous post-Variscan evolution (Burg *et al.* 1994a, 1994b).

Fossil remnants of the rift system preserved in the Pyrenean-Cantabrian orogen are segmented by NNE–SSW to NE–SW trending structures. These structures have a Permo-Carboniferous origin and formed as strike-slip or transfer faults (Arthaud & Matte 1975, 1997; Burg *et al.* 1994a, 1994b). Locally they show evidence of a later reactivation as normal faults during the formation of Permo-Triassic intra-continental basins (Curnelle *et al.* 1982). To the east, the Cevennes fault system delimits the Pyrenean from the Provencal domain (fig 6); to the south, it also may partly control the shape of the South Pyrenean compressional thrust system (Vergés *et al.* 2002; Muñoz 1992).

The Toulouse fault, also referred to as the Villefrance or Sillon Houiller fault, can be mapped from the Massif Central region to the northern Aquitaine basin (fig 6). The southern continuation towards the Pyrenees delimits the eastern termination of the Arzacq–Aquitaine basin from the Occitan high (Curnelle *et al.* 1982). This structure can be identified on gravity inversion maps as a major crustal discontinuity (fig 5). In the Axial zone, it may be mapped as the eastern boundary of a large Carboniferous basin and further to the south to the west of the Organya basin (fig 6) in the South Pyrenean zone (segmentation also described by Muñoz 1992; Vergés & García-Senz 2001).

The Pamplona fault delimits the Mauléon and the Basque–Cantabrian basins (Turner 1996). Although this structure is not exposed continuously, it controls the Mesozoic sedimentation (Larrasoaña *et al.* 2003) and represents a major crustal offset observed from gravity modelling (Bouguer anomaly map; Jammes *et al.* 2010c).

The segmentation of the Basque–Cantabrian basin is well-defined from its sub-division into an eastern "Alava" and western "Peri-Asturian" domain (e.g. Rat 1988; García-Mondéjar *et al.* 1996). This NE–SW structuration terminates towards the Landes High to the north. The western termination of Basque–Cantabrian basin is poorly defined due the Cenozoic tectonic uplift and erosion of Mesozoic sediments south of the "Le Danois basin" and north of the Duero basin (Gallastegui *et al.* 2002; Pedreira *et al.* 2003, 2007; Alonso *et al.* 2007). In this area, Roca *et al.* (2011) proposed the existence of the Santander soft transfer zone that would relay the compressional front of the Pyrenees to the north into the Bay of Biscay. This transfer zone may also be suggested from gravity inversion maps (fig 5) relying the necking domain of the South Iberian margin to the South at the eastern termination of the "Le Danois" basin (fig 6).

4.3. Two distinct rift systems

The previous definition of rift domains, associated structures and segmentation enables us to distinguish two extensional systems: (1) the Bay of Biscay–Parentis rift in the NW and (2) the Pyrenean–Basque–Cantabrian rift to the SE. Both preserve a specific spatial and temporal evolution.

4.3.1 The Bay of Biscay–Parentis rift

The V–shaped Bay of Biscay–Parentis rift system preserves different stages of the lithosphere thinning process ranging from hyperthinned basins in the East to oceanic domains in the West. The Parentis basin is interpreted to show a progressive lateral decrease in extension going eastward (Jammes *et al.* 2010c). Onshore, a crustal discontinuity is observed (fig 5) that may represent the northern continuation of the Pamplona fault (e.g. the Bordeaux fault; Lefort *et al.* 1997). This structure corresponds to the eastern termination of the Parentis basin. The relatively sharp necking and hyperthinned domains of the northern Parentis basin can be mapped in continuity westwards into the Armorican margin. To the south, the Landes High forms a weakly thinned crustal block connecting south-westwards to the necking domain of the North Iberian margin.

The initiation of extensional to transtensional deformation in the Bay of Biscay rift system is difficult to date with precision and relies on local observations. Evidence for a pre-Cretaceous extensional deformation may be found in the Parentis basin recorded by a late Jurassic subsidence (Brunet 1994) and a thickening of pre-Cretaceous sequence toward the centre of the basin (e.g. Ibis fault of Bois & Gariel 1994; Jammes *et al.* 2010a, 2010b, 2010c). Hyperextensional processes may have taken place from Early Cretaceous (Berriasian to Barremian) to at least Aptian time as recorded in the Parentis basin and in the North Biscay margin by the creation of massive accommodation space (Thinon *et al.* 2002, 2003; Tugend *et al. submitted*, see chapter 1). The beginning of sea floor spreading is dated at the Aptian-Albian boundary (Montadert *et al.* 1979b) but may not be synchronous through the overall Bay of Biscay.

The major change in rift domain architecture between the Western Approach and Armorican margins occurs across the Armorican shear zone corresponding to an inherited transfer/transform fault. A similar change of the overall domain architecture can be observed between the eastern and western North Iberian margin west of the “Le Danois basin”.

4.3.2. The Pyrenean–Basque–Cantabrian rift

The Landes High forms a weakly thinned crustal block delimiting the Bay of Biscay–Parentis from the Pyrenean–Basque–Cantabrian rift systems (figs 5 and 6; see also Roca *et al.* 2011). Onshore, the overall crustal architecture of the different domains is not preserved. Only remnants included in the compressional nappe stack can be observed. Ophiolites sampling former remnants of oceanic crust are not observed within the Pyrenean–Basque–Cantabrian rift system (Lagabrielle & Bodinier 2008). The only Cretaceous magmatic additions observed are Late Aptian to Early Santonian (~113 to 85 Ma; Lamolda *et al.* 1983; Montigny *et al.* 1986) alkaline sills and flows preserved into remnants of an exhumed mantle domain or the overlying sediments. This observation suggests that this extensional system did not evolve into a mature oceanic domain. Furthermore, mantle outcrops are usually associated with small fragments of continental crust (at least in the Western Pyrenees: Debroas *et al.* 2010; and Basque–Cantabrian basin: Mendia & Gil-Ibarguchi 1991) indicating that probably only local windows of exhumed mantle existed.

The increasing subsidence recorded in the onshore Arzacq and Tarbes basins (Désegaulx & Brunet 1990) during the Late Jurassic to Early Cretaceous may indirectly date the onset of transtensional deformation (Canérot 2008; Jammes *et al.* 2009). Hyperextension only initiated in the Late Aptian to Albian time in the Pyrenean–Basque–Cantabrian rift (Jammes *et al.* 2009; Lagabrielle *et al.* 2010) coinciding with the beginning of sea-floor spreading in the Bay of Biscay (Montadert *et al.* 1979b). The sedimentary record preserved in the Pyrenean–Cantabrian orogen indicates a progressive deepening of the basin from Late Aptian to Mid-Albian (Pyrenees: Debroas 1987, 1990; Basque–Cantabrian basin: García-Mondéjar *et al.* 1996, 2005). Mantle exhumation processes may have lasted until Early Cenomanian as suggested by the reworking of mantle rocks in Mid-Albian to Cenomanian breccias (fig 3; Urdach breccia: Jammes et al. 2009; Debroas *et al.* 2010; Lagabrielle *et al.* 2010; Aulus breccia: Clerc *et al.* 2012).

In spite of the well-defined NE–SW segmentation, an overall lateral continuity of hyperextended domains can be observed throughout the Pyrenean orogen (fig 6). The Pamplona fault represents a major crustal discontinuity (Turner 1996; Larrasoaña *et al.* 2003; Jammes *et al.* 2010c) delimiting the Basque–Cantabrian from the Arzacq–Mauléon basin.

5. Structural evolution: insights from geological sections

The mapping of the rift domains across the Bay of Biscay–Pyrenees enables us to characterize the progressive compressional overprint of the former rift systems. To the west, the initiation of reactivation can be investigated to the east, a complete reactivation of rift structures is observed. In order to decipher the possible interaction between the two extensional rift systems during the Late Cretaceous to Oligocene compression, three segments are distinguished (fig 6 and 7): (1) a western "Bay of Biscay segment"; (2) a central "Basque–Parentis segment" preserving heterogeneously reactivated hyperextended domains and (3) an eastern "Pyrenean segment" that is completely inverted, exemplified by the Mauléon–Arzacq basin.

5.1. Western Bay of Biscay segment

5.2.1. Constraints from seismic interpretations and gravity inversion

The western Bay of Biscay segment initiates at the western termination of the onshore Pyrenean–Basque–Cantabrian rift system. Located at the junction with the Atlantic Ocean, it preserves the ultimate stage of lithosphere thinning processes in the Bay of Biscay.

In order to characterize the architecture of this domain, two conjugate geological sections are proposed, extending from the proximal to oceanic domains in the Western Approach margin to the North (fig 7a and 8) and across the North Iberian margin to the South (fig 7a and 8). The sedimentary and basement architecture of these sections is derived from the interpretation of the Norgasis 11-12 seismic sections (fig 8; Avedik *et al.* 1993, 1996; Thinon 1999) and IAM 12 line (fig 8; Banda *et al.* 1995; Alvarez-Marron *et al.* 1997), respectively extending in the Western Approach and North Iberian margins. Moho depth is determined from gravity inversion results (fig 5; parameters are given in table 1) combined with refraction data under the proximal domain of the North Iberian margin (Fernández-Viejo *et al.* 1998).

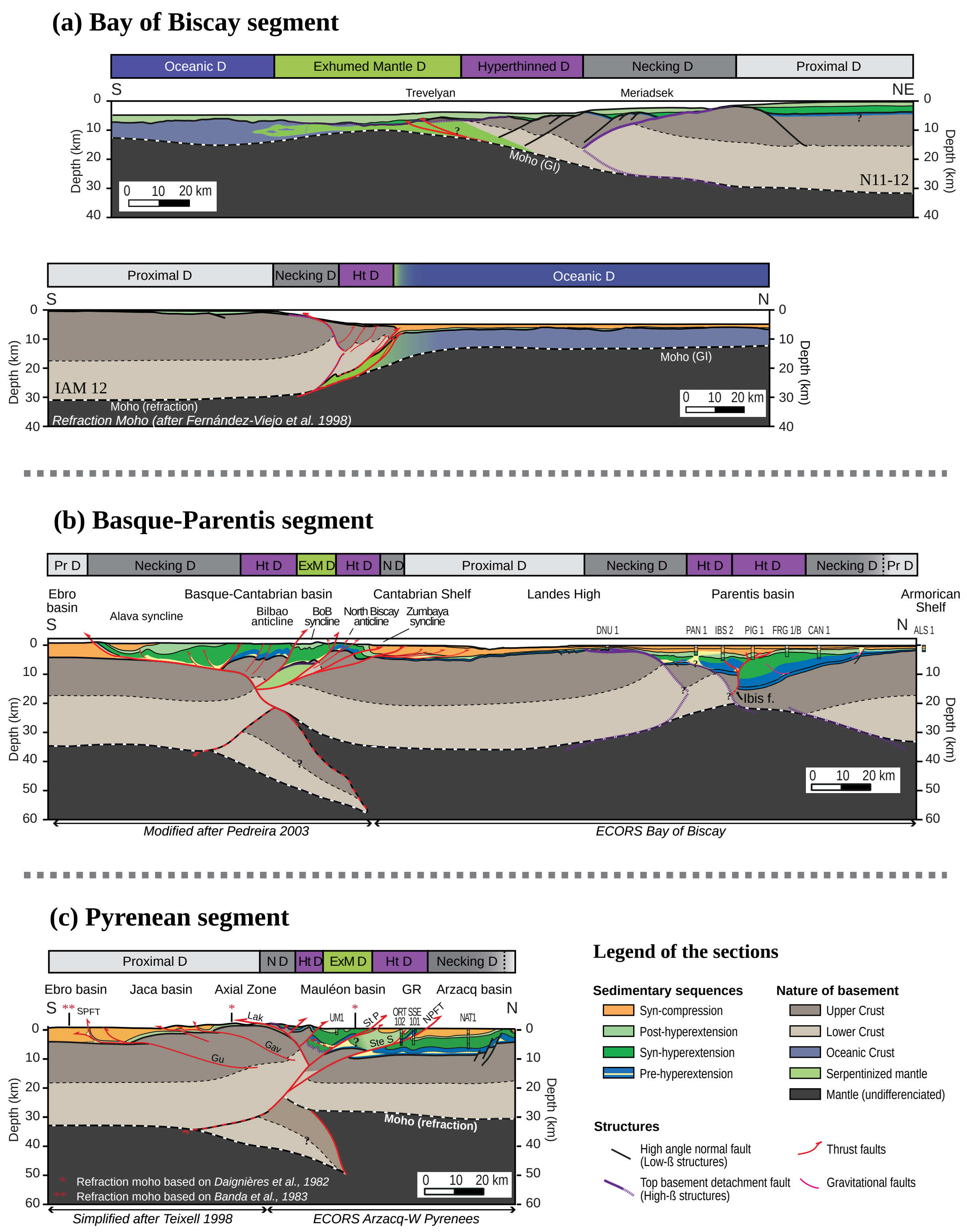


Fig 7: *Geological sections, location fig 6 (a) Bay of Biscay segment: the architecture is based on the Norgasis 11-12 and IAM 12 seismic profiles (see also fig 8). (b) Basque-Parentis segment: the section is a composite between the depth interpretation of the ECORS Bay of Biscay seismic profile and the gravity modelling proposed by Pedreira 2003 for the Basque-Cantabrian basin. (c) Pyrenean segment: the section is based on the interpretation of the ECORS Arzacq-W Pyrenees and the southern extension is modified after Teixell 1998.*

5.2.2. Architecture of the Western Approach margin

The rift-related crustal architecture of the Western Approach margin is well-preserved. Detailed descriptions of seismic observations supporting the interpretation of this section are proposed by Tugend *et al.* (*submitted*, see chapter 1) and summarized below. The proximal domain crossing the Western Approach basin is characterized by its half graben-type basin architecture (fig 7a and 8). Deformation is interpreted to decouple at mid-crustal levels. No drill hole reached the basement of the necking domain, but based on the overlying sedimentary architecture (fig 8) and onshore analogy (fig2; Tugend et al. *submitted,* see chapter 1) it is suggested to be structured by a conjugate set of top basement detachment faults delimiting a crustal necking zone (Mohn *et al.* 2012; Sutra *et al.* 2013). The hyperthinned domain is characterized by a titled block architecture limited by short-offset normal faults dipping oceanward (fig 7a and 8). These structures are rooting onto the S reflector (fig 8; De Charpal *et al.* 1978; Montadert *et al.* 1979b; Thinon 1999) interpreted as the crust-mantle boundary. The southern edge of this domain is proposed to be structured by a top basement detachment fault exhuming lower crust and mantle to the seafloor during the final stage of rifting. The occurrence of an extensional allochthon overlaying the exhumed mantle is proposed. It is delimited at its base by the S reflector on seismic sections (fig 8; Thinon 1999). The transition to the oceanic domain may not be a sharp boundary and may result from a progressive magmatic overprint of the exhumed mantle domain as suggested by refraction data (Thinon *et al.* 2003). The only important compressional reactivation is observed close to the transition between the exhumed mantle and hyperthinned domains, forming the Trevelyan structure (Debyser *et al.* 1971; Derégnaucourt & Boillot 1982; Thinon *et al.* 2001). The associated thrust fault system is interpreted to root in the serpentinized upper mantle (fig 7a and 8).

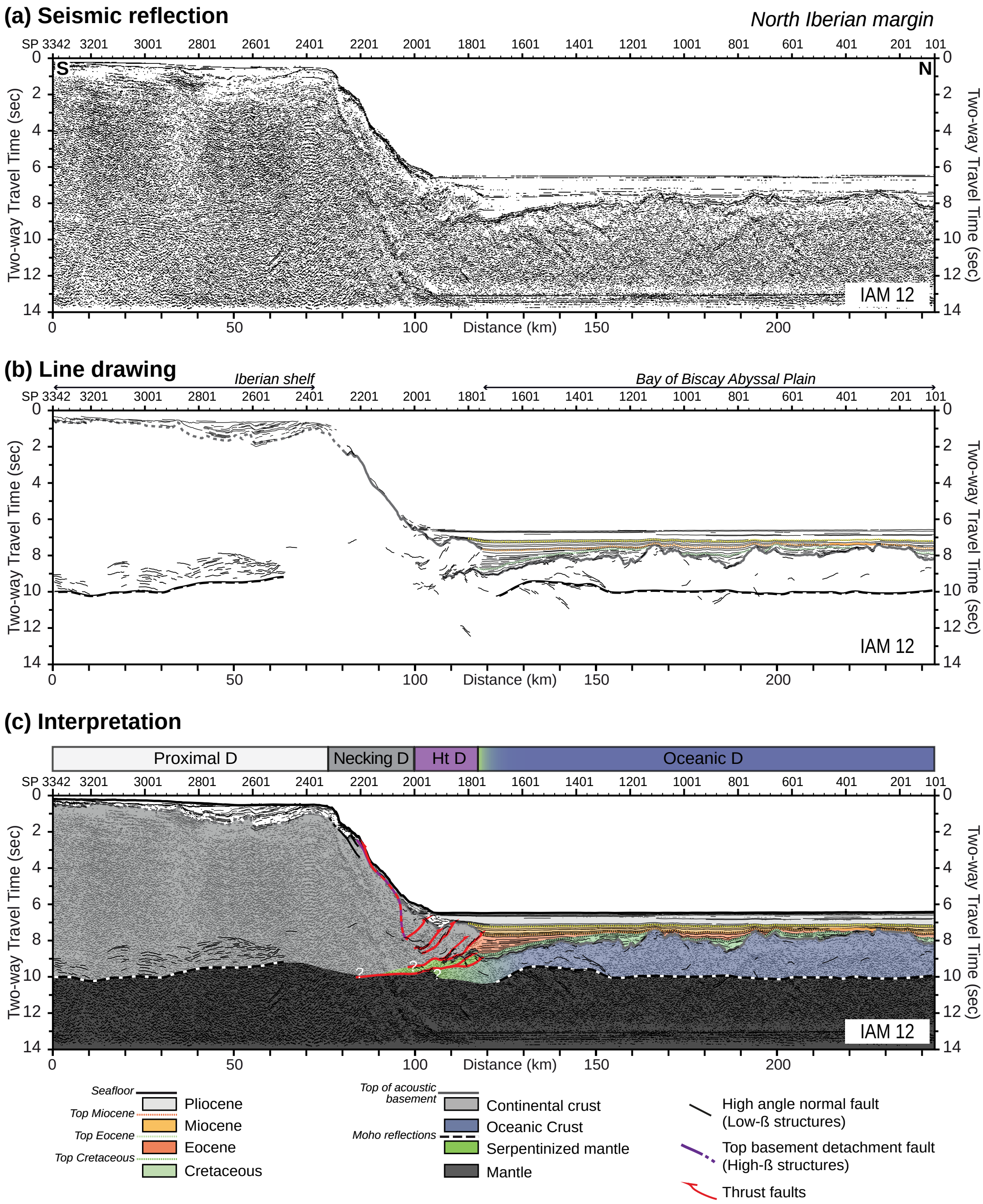
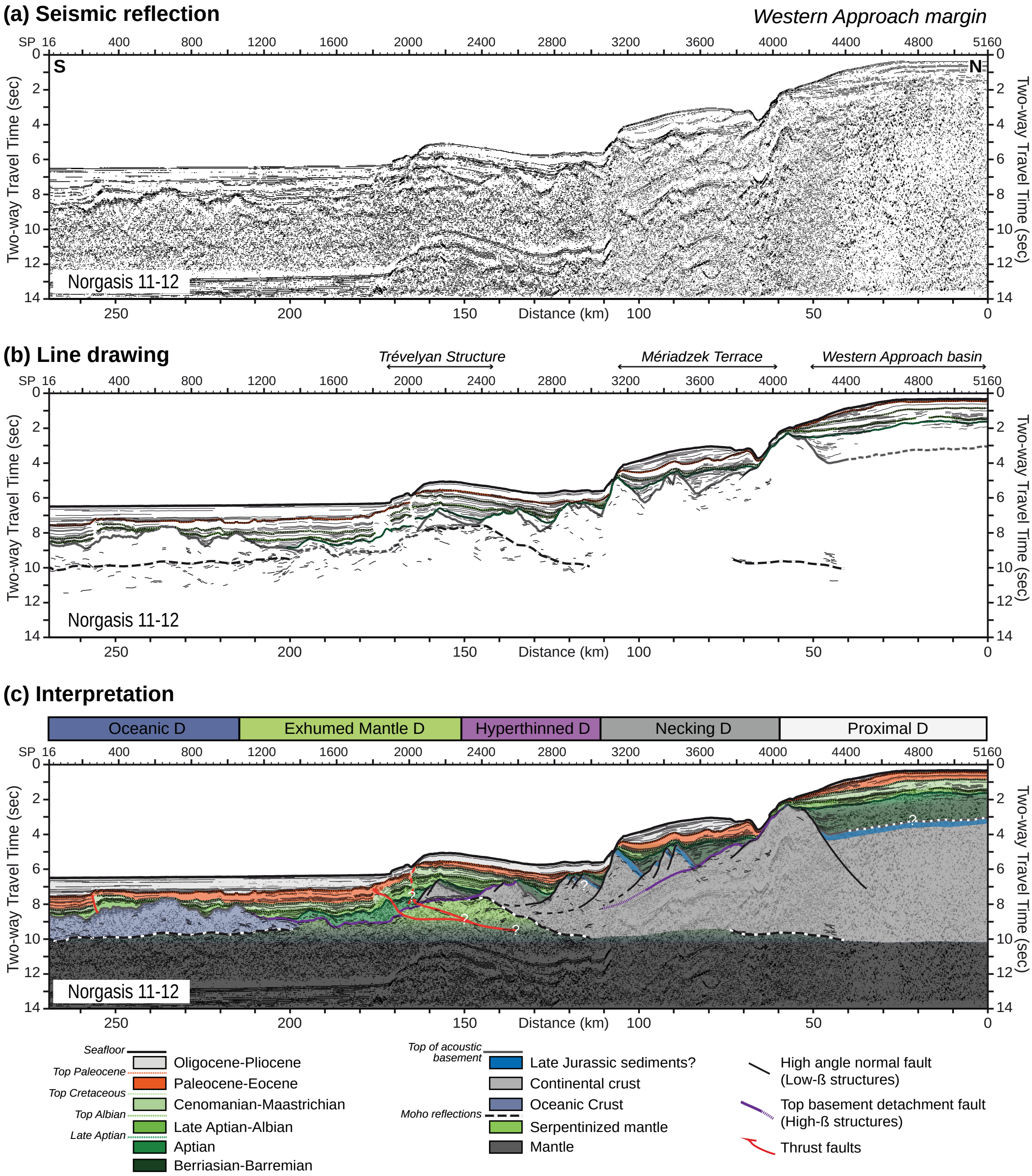


Fig 8: *(a) Seismic reflection (b) line drawing and (c) interpretation proposed in this study for the Norgasis 11 and 12 seismic sections along the Western Approach margin (to the right) and of the IAM 12 seismic line along the North Iberian margin (to the left). The location of the seismic lines is the same as for fig 7a.*



(a) Seismic reflection
Western Approach margin
SP 16 400 800 1200 1600 2000 2400 2800 3200 3600 4000 4400 4800 5160
S
N
Two-way Travel Time (sec)
Norgasis 11-12
Distance (km)
250 200 150 100 50 0
(b) Line drawing
Trévelyan Structure
Mériadzek Terrace
Western Approach basin
(c) Interpretation
Oceanic D
Exhumed Mantle D
Hyperthinned D
Necking D
Proximal D
Seafloor
Top Paleocene
Top Cretaceous
Top Albian
Late Aptian
Oligocene-Pliocene
Paleocene-Eocene
Cenomanian-Maastrichian
Late Aptian-Albian
Aptian
Berriasian-Barremian
Top of acoustic basement
Moho reflections
Late Jurassic sediments?
Continental crust
Oceanic Crust
Serpentinized mantle
Mantle
High angle normal fault
(Low-ß structures)
Top basement detachment fault
(High-ß structures)
Thrust faults

5.2.2. Reactivation of the North Iberian margin

The distal part of the North Iberian margin was interpreted in the past as an accretionary prism related to the formation of an oceanic subduction (e.g. Boillot 1984; Alvarez-Marron *et al.* 1997). Based on the refraction results any important underthrusting of oceanic crust below the South Iberian margin (Fernández-Viejo *et al.* 1998; Ruiz 2007) is precluded. Therefore, many authors suggested a complete or partial overprint of the former rift architecture (Gallastegui *et al.* 2002; Fernández-Viejo & Gallastegui 2005; Alonso *et al.* 2007; Roca *et al.* 2011; Fernández-Viejo *et al.* 2012). The identification of the rift domains (fig 6) enables us to propose a new interpretation that characterizes the deformation in each rift domain clearly unraveling the role of the rift inherited architecture. No major evidence of reactivation can be observed from the proximal and necking domains preserving their rift-related architecture delimited oceanwards by a crustal necking zone similar to the structures proposed for the Western Approach margin. The accretionary wedge characterizing the distal part of the North Iberian margin (fig 8) is interpreted as the former hyperthinned domain that has been reactivated (fig 7a). Refraction velocities are compatible with the hypothesis of a deformed wedge made of continental crust and sediments rather than of sediments only (Ruiz 2007; Fernández-Viejo *et al.* 2012). The existence of an exhumed mantle domain is proposed northwards and eastwards and may have also occurred in this part of the Bay of Biscay. We suggest that it may be partly underthrust below the hyperthinned domain during the compressional overprint and formation of a Tertiary flexural basin on top of the oceanic domain (fig 7a and 8; Alvarez-Marron *et al.* 1997). In our interpretation the exhumed serpentinized mantle represents a low friction surface between the hyperthinned domain and the oceanic domain that is only weakly reactivated (fig 7a).

5.2. Eastern Bay of Biscay: Parentis and Basque–Cantabrian segment

5.2.1. Constraints from geological and geophysical datasets

In the Parentis and Basque–Cantabrian segment, both preserved and reactivated rift structures can be investigated. The eastern Bay of Biscay is characterized by its progressive termination into the hyperthinned Parentis basin. Eastwards this segment terminates towards the Pamplona fault delimiting the Basque–Cantabrian basin from the Western Pyrenees.

The geological section shown in fig 7b is based on the interpretation of the sedimentary and top basement architecture of the ECORS Bay of Biscay section (Pinet *et al.* 1987; Bois & Gariel 1994) located in the Parentis basin. The southward extension into the Basque–Cantabrian basin is based on published field studies supported by local sub-surface data (IGME 1987). Moho depth is based on refraction studies in the Parentis basin (Tomassino & Marillier 1997). In the Basque–Cantabrian basin, the crustal architecture is based on one of the possible solutions derived from gravity and magnetic modelling proposed by Pedreira *et al.* (2007).

5.2.2. The rift architecture preserved in the Parentis basin

The Parentis basin is only weakly reactivated in its central part and the overall rift-related sedimentary and crustal architectures are relatively well-preserved (fig 7b). The architecture of the basin is strongly asymmetric unravelling a major crustal discontinuity towards the emplacement of the so-called sub-vertical Ibis fault (Jammes *et al.* 2010a, 2010b, 2010c). The northern geometry is characterized by the progressive transition from a proximal to a hyperthinned domain without any major rift structure and the complete record of the pre-, syn-, and post-hyperextension sedimentary sequences. The southern part of the basin is more complex. The necking domain is structured by a conjugate system of detachment faults. Evidence for such structures is based on drill-hole observations (the Saint Girons en Marensin and Contis boreholes; Jammes *et al.* 2010b). The structural limit between the northern and southern part of the hyperthinned domain of the Parentis basin is also interpreted as a top basement detachment fault rooting towards the Ibis fault and associated with salt tectonics and diapir formation (e.g. Mathieu 1986; Biteau & Canérot 2007; Jammes *et al.* 2010b). The Landes High and the Cantabrian shelf belonging to the proximal domain represent a weakly thinned crustal block and form a major boundary between the Parentis and Basque–Cantabrian basin (see also Roca *et al.* 2011).

5.2.3. Rift-remnants and reactivation in the Basque–Cantabrian basin

In contrast the Basque–Cantabrian basin to the South is completely inverted (fig 7b). Sedimentary sequences are decoupled on upper Triassic evaporates (Cámara 1997) and the basement is buried under thick sedimentary sequences. Field and sub-surface observations provide good constraints on the stratigraphic evolution of the Mesozoic sediments (e.g. Soler *et al.* 1981; García-Mondéjar *et al.* 1985; Rat 1988; IGME 1987) and on locally preserved rift related-structures (e.g. García-Mondéjar *et al.* 1996; Tavani & Muñoz 2012; Bodego & Agirrezabala 2013). Similarly as for the northern Landes high, the Rioja shelf to the South (buried under the Ebro basin) is representative of a proximal domain, characterized by the deposition of thin Late Cretaceous post-hyperextension shallow marine sediments (e.g. Cámara 1997). The sedimentary sequences of the conjugate necking domains are preserved in the Zumbaya and Alava synclines (respectively in the northern and southern part of the basin). These units are characterized by thickening of syn- (Late Aptian to Cenomanian) to post-hyperextensional sequences toward the basin axis and locally absent pre-hyperextensional sequences (Castillo 5; IGME 1987). The North Biscay and Bilbao anticline are mapped as remnants of hyperthinned domains and characterized by thick pre-hyperextension sediments. The Bay of Biscay syncline is mapped as the western continuation of the “Nappe des Marbres” locally sampling mantle rocks and associated high temperature low pressure metamorphic marbles of Mesozoic age (Mendia & Gil-Ibarguchi 1991) representative of a former exhumed mantle domain. An important gravity anomaly is centered on the basin and interpreted to result from the stacking of lower crustal rocks (Pedreira *et al.* 2003, 2007) or mantle rocks emplaced at shallow levels during compression. In analogy with the interpretation proposed for the Labourd anomaly (Jammes *et al.* 2010c), we favor the mantle body solution that would be coherent with this unit being a remnant of a former exhumed, partly serpentinized, mantle domain. Nevertheless the lower crustal interpretation proposed by Pedreira *et al.* (2003, 2007) cannot be excluded. Another important observation resulting from the gravity modeling proposed by Pedreira *et al.* (2007) is the occurrence of a large northward dipping crustal root reaching a 55 to 60 km depth and interpreted as the former thinner basement of the Basque–Cantabrian basin (Pedreira 2004; Pedreira *et al.* 2007).

5.3. Arzacq–Mauléon basin: Pyrenean segment

5.3.1. Constraints from field and sub-surface geology

The Pyrenean segment initiates at the eastern termination of the Parentis and Basque–Cantabrian basins and is delimited from the two basins by the Pamplona fault. Therefore this segment only preserves rift remnants derived from the former hyperextended Pyrenean–Basque–Cantabrian rift system.

Insights on the sedimentary and basement architecture come from the reinterpretation of the ECORS Arzacq–W Pyrenees seismic profile (Daignères *et al.* 1994) extended to the South based on the Anso transect proposed by Teixell (1998). Field geology, sub-surface observations (BRGM 1974) and local refraction data (Daignières *et al.* 1982; Banda *et al.* 1983) are combined to further constrain the interpretation proposed in fig 7c.

5.3.2. Compressional architecture of a reactivated hyperextended basin

Arguments for the determination of rift-related remnants in the Arzacq–Mauléon basin have been previously presented and described relying on the identification of key field observations (fig 3). Field observations from the “Béarnais range” area, in the eastern part of the basins (fig 3) are used to further constrain the compressional overprint of former rift domains (figs 7c, 9 and 10). The present-day geometry of the Mauléon basin results from its inversion within a large pop-up structure along north and south vergent thrusts (fig7c and 9).

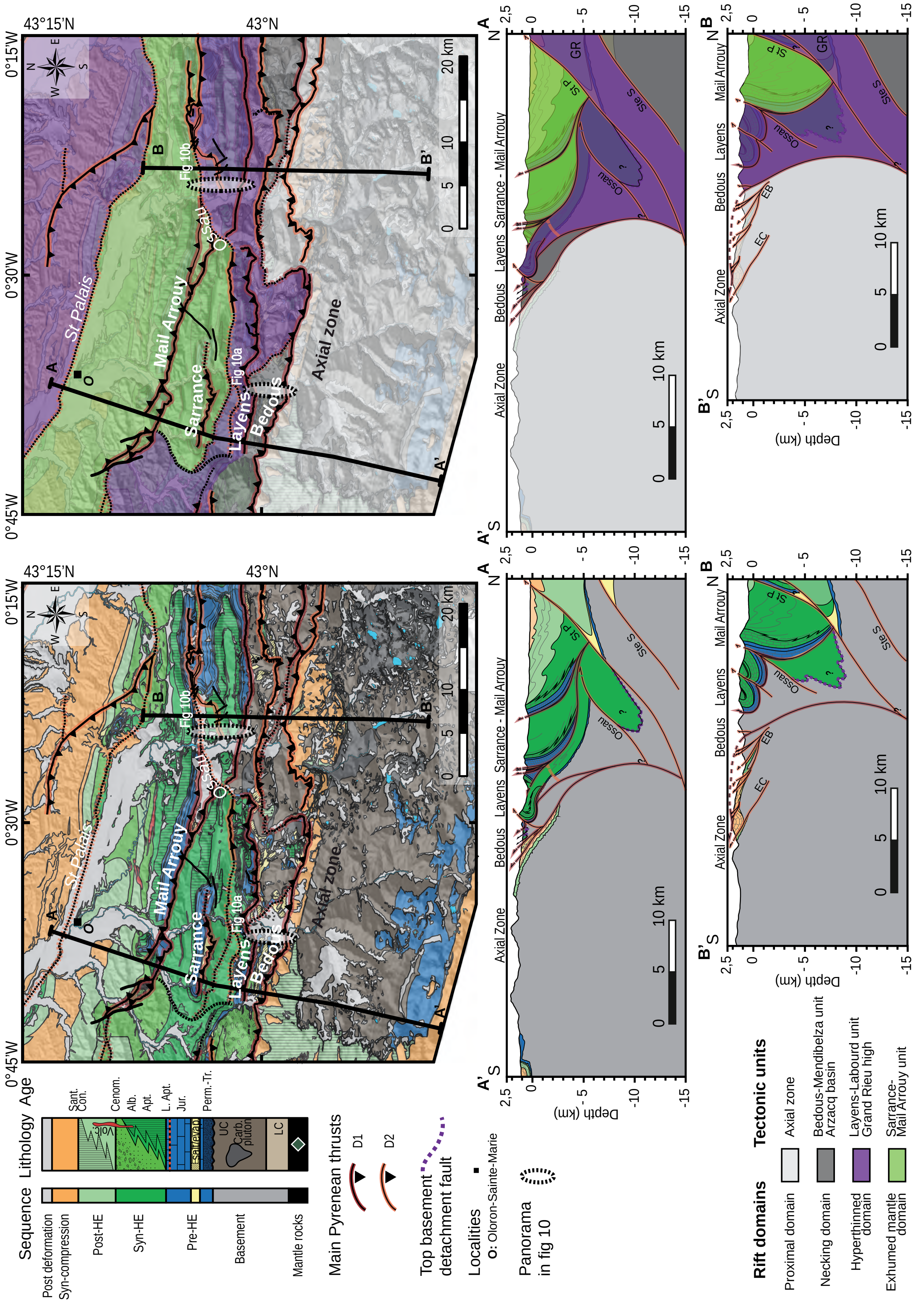
To the south, distal parts of the basin (corresponding to the former hyperextended domain) are thrust over the proximal domain preserved in the axial zone along the south-directed Lakoura thrust system and lateral equivalents (figs 7c, 9 and 10; Muñoz 1992; Teixell 1998). The proximal domain is weakly deformed as a large-scale anticline (geological sections fig 10; see also Teixell 1990) and by local thick-skin structures related to the final stage of deformation (D2 phase; e.g. Guarga and Gavarnie thrusts fig 7c; Eaux-Chaudes thrust fig 9). The sedimentary cover of the axial zone is characterized by thin-skin thrusts decoupling in Upper Triassic evaporates (Teixell 1998; geological section BB’ fig 9).

The transition from the proximal to necking domains (see structural maps; figs 3, 9 and 10) corresponds to a thrust fault characterized by an intense deformation of the sediments close to the contact (e.g. Upper Triassic formation within the Bedous unit; fig 10; Canérot *et al.* 2001; Ternet *et al.* 2004). Locally, this thrust fault is folded by the anticlinal deformation of the axial zone as indicated by the steep northward dipping Cretaceous cover of the axial zone (fig 9 and 10). Remnants of the necking domain are only locally sampled within the Bedous-Mendibelza unit as syncline structures. In contrast, the rift-related sedimentary and basement architecture of the conjugate remnants of the necking domain preserved in the Arzacq basin is interpreted to be well preserved (fig 7c).

The initial contact between units preserving the necking domain and the hyperthinned domain is only locally preserved. In most cases it is overprinted by North-directed thrusts resulting from the second phase of deformation (figs 7c, 9 and 10). The deformation of the hyperthinned and exhumed mantle domains is characterized by the wide-spread occurrence of this second phase of deformation locally inverting and overprinting tectonic units preserving remnants of hyperthinned and exhumed mantle domains (e.g. the Ossau thrust; Canérot *et al.* 2001; figs 7c, 9 and 10). As a result a large part of the hyperthinned and exhumed mantle domains may be underthrust as suggested in the geological sections in figs 7c and 9.

Field observations from the Mauléon basin unravel the contrasting compressional deformation between the different rift domains. The proximal domain is only weakly deformed in contrast to the hyperthinned and exhumed mantle domains that are affected by a stronger compressional overprint.

Fig 9: *Zoom over the "Béarnais Range" area (location fig 3). To the left: geological map and two geological sections (AA' and BB'). To the right: the same map indicating the remnants of rift domains. The associated cross sections illustrate the reactivation of the former rifted domains. EB: Eaux-Bonnes thrust; EC: Eaux-Chaudes thrust; GR: Grand Rieu High; St P: Saint Palais thrust system; Ste S: Sainte Suzanne thrust system.*



Sequence
Lithology
Age
Post deformation
Syn-compression
Post-HE
Syn-HE
Pre-HE
Basement
Mantle rocks
Sant.
Con.
Cenom.
Alb.
Apt.
L. Apt.
Jur.
Perm.-Tr.
Volc.
UC
Carb.
pluton
LC
Main Pyrenean thrusts
D1
D2
Top basement detachment fault
Localities
O: Oloron-Sainte-Marie
Panorama in fig 10
Rift domains
Proximal domain
Necking domain
Hyperthinned domain
Exhumed mantle domain
Tectonic units
Axial zone
Bedous-Mendibelza unit Arzacq basin
Layens-Labourd unit Grand Rieu high
Sarrance-Mail Arrouy unit
43°15'N
43°N
0°45'W
0°30'W
0°15'W
St Palais
Mail Arrouy
Sarrance
Layens
Bedous
Axial zone
Ossau
Fig 10a
Fig 10b
A
A'
B
B'
0 5 10 20 km
Axial Zone
Bedous
Layens
Sarrance - Mail Arrouy
Mail Arrouy
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Sle S
GR
EB
EC
10 km
Depth (km)
2,5
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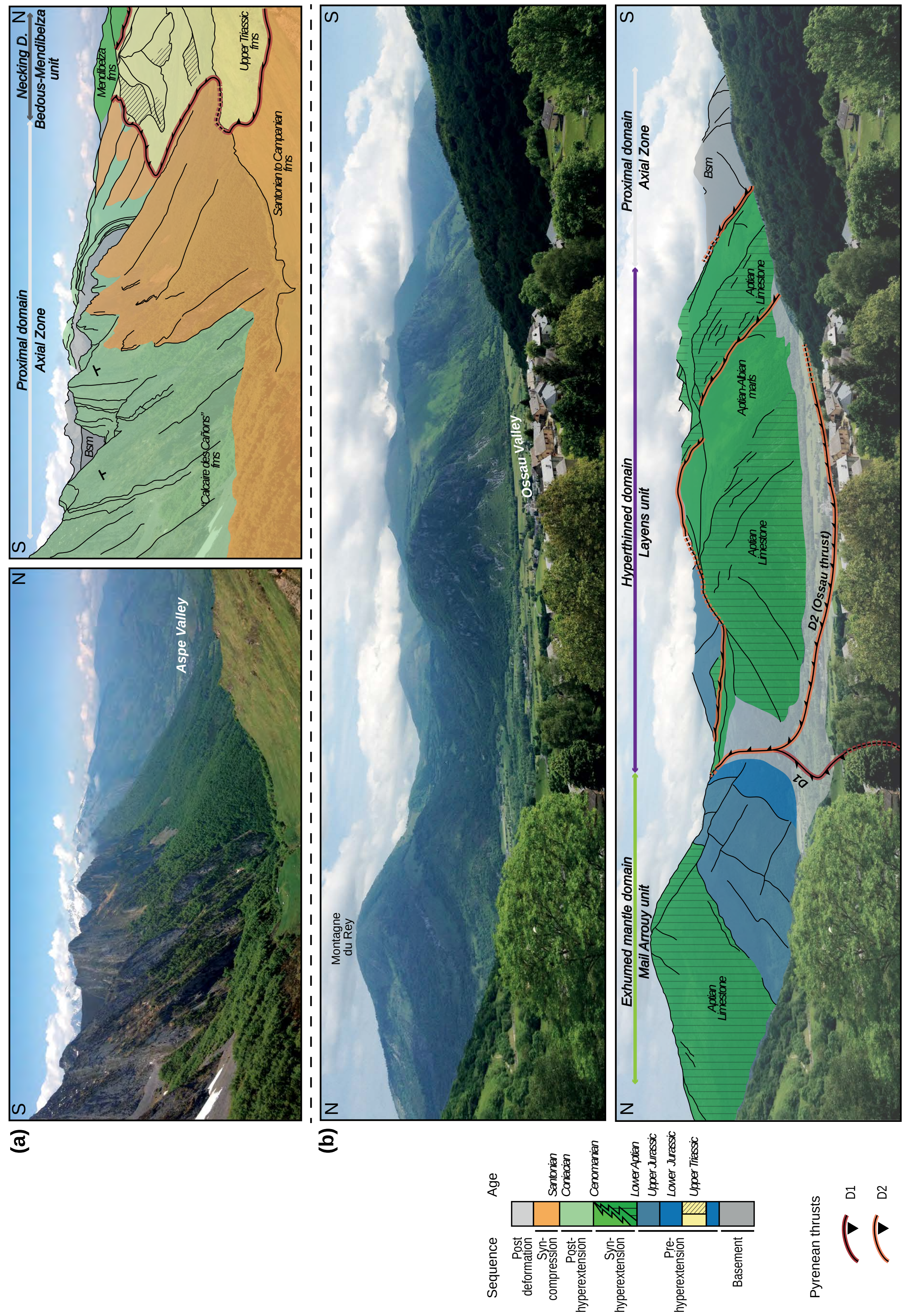

(a)
S
N
Aspe Valley
Proximal domain
Axial Zone
Necking D.
Bedous-Mendibelza unit
Mendibelza fms
Upper Triassic fms
Santonian to Campanian fms
"Calcaire des Cañons" fms
Bsm
(b)
N
S
Montagne du Rey
Ossau Valley
Exhumed mantle domain
Mail Arrouy unit
Hyperthinned domain
Layens unit
Proximal domain
Axial Zone
Aptian Limestone
Aptian-Albian marls
D2 (Ossau thrust)
D1
Bsm
Sequence
Age
Post deformation
Syn-compression
Post-hyperextension
Syn-hyperextension
Pre-hyperextension
Basement
Santonian Coniacian
Cenomanian
Lower Aptian
Upper Jurassic
Lower Jurassic
Upper Triassic
Pyrenean thrusts
D1
D2

6. Discussion

6.1. Spatial and temporal evolution of hyperextended rift systems

6.1.1. The diachronous evolution of the Bay of Biscay–Parentis and Pyrenean–Basque–Cantabrian rift systems

The mapping of rift domains across the Bay of Biscay and Pyrenees unravelled the occurrence of two spatially disconnected rift systems characterized by a different temporal evolution (e.g. García-Mondéjar *et al.* 1996; Thinon 1999; Jammes *et al.* 2009; Roca *et al.* 2011) (fig 11). These basins are strongly segmented and are separated by the "Landes High", a crustal block that was only weakly affected by crustal thinning (see also Roca *et al.* 2011). Its crustal structure as well as its relationship to the surrounding hyperextended domains compares well with continental ribbons described by Péron-Pinvidic & Manatschal (2010) from the North Atlantic (e.g. Flemish Cap, Rockall Bank, Galicia Banks). This Landes High connects westwards to the necking domain of the North Iberian margin (figs 6 and 11).

Both rift systems are characterized by the occurrence of extreme crustal thinning, but the relative timing of hyperextension is different. In the Bay of Biscay, hyperextension initiates during the Early Cretaceous (Berriasian to Barremian) and results in the formation of necking and hyperthinned domains. During the late Aptian, mantle may have been exhumed in the Bay of Biscay as suggested by Thinon *et al.* (2002). In the Parentis basin further to the east extension did not go to mantle exhumation as shown in geological sections (fig 7 and restorations fig 11). The initiation of sea-floor spreading processes in the western Bay of Biscay (around Aptian–Albian time: Montadert *et al.* 1979b; Boillot 1984; Vergés & García-Senz 2001) is synchronous with the northward propagation of the Atlantic Ocean (e.g. Williams 1975; Olivet 1996; Rosenbaum *et al.* 2002). The lateral preservation of different rift stages along the Bay of Biscay–Parentis rift system suggests that this basin formed as a V-shaped basin (e.g. Jammes *et al.* 2010a).

Fig 10: *(a) Panoramic view showing the contact between the proximal (axial zone) and necking domains (Bedous-Mendibelza unit). Note that the contact is steepening eastwards (0°31'45,07"W/42°57'08,21"N). (b) Panoramic view showing the relashionships between the different phases of deformation observed in the hyperthinned (Layens unit) and exhumed mantle domains (Mail Arrouy unit). The initial south directed stacking of units (D1) is overprinted by north directed deformation (D2), exemplified here by the Ossau thrust (0°27'16,21"W/43°03'30,98"). The location of the photographs is indicated on fig 9. The coordinates indicate the location from which the photographs were taken.*

In contrast, in the Pyrenean–Basque–Cantabrian rift system, hyperextension is only documented to initiate during the Late Aptian–Early Albian (Jammes *et al.* 2009) as indicated by the initiation of a rapid subsidence and the occurrence of thick deep water Albian deposits (e.g. Basque–Cantabrian basin: García-Mondéjar *et al.* 1996, 2005; and Pyrenean basins: Debroas 1987, 1990). Local mantle exhumation is indirectly dated by the reworking of mantle rocks in Mid-Albian to Cenomanian breccias (e.g. Urdach breccia in Western Pyrenees: Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010 or Aulus breccias in Eastern Pyrenees: Clerc *et al.* 2012). This age is further supported by a recent review on serpentinisation and magmatic processes and on the formation of ophicalcites across the Pyrenees (Clerc *et al.* 2013). The diachronous evolution observed between the two rift systems may illustrate the eastwards migration of extension in the Pyrenean–Basque–Cantabrian rift system contemporaneous with ongoing seafloor spreading in the western Bay of Biscay.

6.1.2. Lateral evolution of the rift architecture: insights on the role of segmentation and pre-rift inheritance

6.1.2.1. The Bay of Biscay–Parentis system: the role of pre-rift inheritance

Establishing maps showing the lateral distribution of the different rift domains enables the investigation of the spatial evolution of extensional and compressional processes. Along strike, changes in the rift architecture between the Western Approach and the Armorican margins have been previously described by Thinon (1999). The architecture of the Western Approach margin is characterized by extremely thinned tilted blocks and an extensional allochthon overlying exhumed serpentinized mantle. The southern edge of the hyperthinned domain is suggested to be structured by the break-away of a detachment system related to final crustal thinning and mantle exhumation. This architecture is interpreted to characterize distal lower plate margins (Sutra *et al.* 2013), implying that the Western Approach margin may correspond to a lower plate margin (fig 7a and 8). The situation is inverted at the eastern end of the Bay of Biscay where the northern part of the Parentis basin can be interpreted as an upper plate margin (see also Jammes *et al.* 2010a, 2010b, 2010c). This change between upper and lower plate margins seems well-expressed from a mapping point of view (fig 6 and 11). The Western Approach margin is characterized by wide necking and hyperthinned domains, contrasting with sharper domains observed from the Armorican margin to northern Parentis basin. This switch from an upper to a lower plate margin occurs where the Armorican Shear zone intersects with the margin. The shear zone structure is described as a major lithospheric structure delimiting different Variscan domains (e.g. Matte 2001; Martínez-Catalán *et al.* 2012; Ballèvre *et al.* 2012 and reference therein).

In the North Iberian margin, the hyperthinned and exhumed mantle domains are reactivated within north directed thrust systems, precluding the identification of the rift-vergence of the detachment system in seismic sections. However, along the North Iberian margin (exemplified by the IAM 12 section; fig 7a and 8) and the “Le Danois basin”, a change in the size of the necking and hyperthinned domain can be observed in a map view (fig 6). The transition is relatively abrupt and not progressive, probably not resulting from a difference of shortening along strike. This change could be inherited from the former rift architecture, as proposed for the northern Bay of Biscay margin. It may illustrate a former switch between an upper plate margin (to the west) and a lower plate margin (to the east). Furthermore, this change of architecture seems to coincide with the intersection of the rift system with major inherited Variscan structures separating the Central Iberian Zone to the West from the West Asturien Leonese Zone to the East (Martínez-Catalán *et al.* 2012; Ballèvre *et al.* 2012 and references therein).

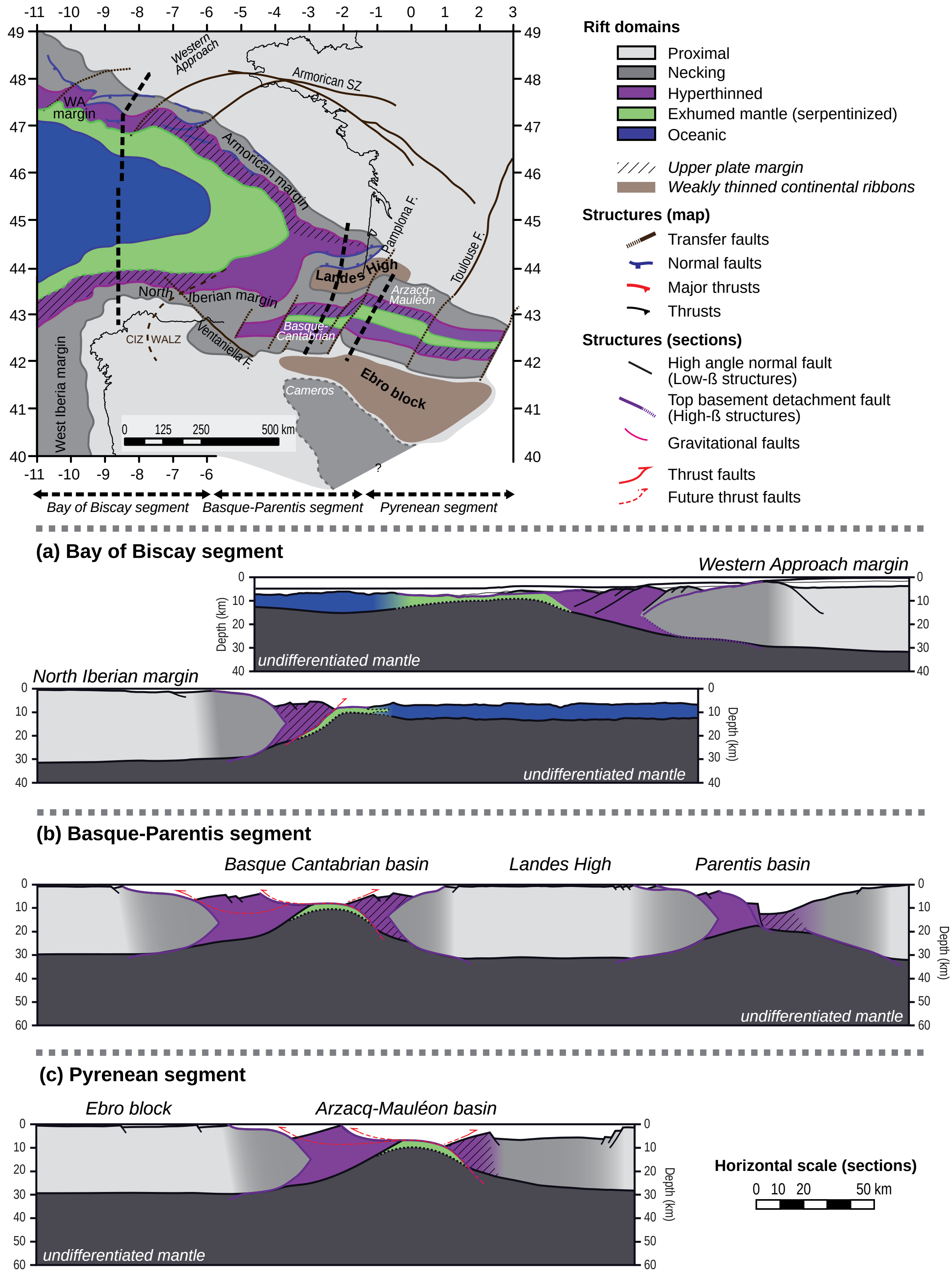
Based on observations from the conjugate margins of the Bay of Biscay, it seems that switches between upper and lower plate architecture can be controlled by the pre-rift history and inherited rheology of the crust as they bound distinct Variscan domains. They impact the architecture of the necking, hyperthinned and exhumed mantle domains. It is therefore likely that the evolution of these domains may be controlled by the inherited rheology (as discussed by Pérez-Gussinyé *et al.* 2003 or Sutra & Manatschal 2012 for the West Iberia margin). It is interesting to note that these inherited structures the does not seem to offset the contact to the first oceanic crust. This observation may suggest that while rift systems may be controlled by crustal and lithospheric inheritance oceanic systems are mainly controlled by deep-seated asthenospheric processes and are therefore insensitive to continental lithospheric inheritance.

6.1.2.2. The Pyrenean–Basque–Cantabrian hyperextended rift system: the importance of segmentation

The Pyrenean–Basque–Cantabrian rift system is characterized by its strong segmentation corresponding to a set of NE–SW to NNE–SSW trending transfer/transform faults regularly spaced across the studied area (e.g. the Santander, Pamplona, Toulouse, Cevennes faults, Jammes *et al.* 2009; Roca *et al.* 2011). They are considered to be lithospheric scale structures partly inherited from the late to post-Variscan evolution (e.g. Arthaud & Matte 1975, 1997; Burg *et al.* 1994a, 1994b). These transfer faults control the large-scale lateral evolution of the Late Aptian–Early Albian hyperextended rift system (see the restoration prior to the onset of compression; fig 12). These faults can either juxtapose rift domains that experienced different amounts of extension or transfer the overall extension to another basin as observed along the Pamplona fault (fig 12). Further to the east a switch in exhumation vergence, similar to the one described in the western Bay of Biscay, occurs along the Toulouse fault controlling a different rift architecture between the western and eastern Pyrenees. This change was already suggested by Jammes *et al.* (2009). It separates north dipping top basement detachment fault systems exhuming mid-crustal granulites and mantle rocks in the Arzacq–Mauléon basin (e.g. Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Masini 2011) from south dipping detachment fault systems also exhuming granulites and mantle in the eastern Pyrenees. This was recently demonstrated by Vauchez *et al.* (2013) for the Agly massif (in the eastern Pyrenees). Vauchez *et al.* (2013) further suggested that a similar exhumation of the North Pyrenean massif in the Eastern Pyrenees is possible as most of them are also characterized by a tectonic contact between mid to lower crustal rocks and Cretaceous sediments.

The suggested change in the rift-related pre-compressional architecture may partly explain the change in the orogen architecture described between the western and eastern Pyrenees (e.g. Muñoz 2002 and reference therein).

Fig 11: *Restoration of the Bay of Biscay and Pyrenean rift systems prior to the onset of compression (before the Santonian). Restored sections are proposed in the (a) Bay of Biscay segment (b) Basque-Parentis segment and (c) Pyrenean segment. WA margin: Western Approach margin.*



Rift domains
Proximal
Necking
Hyperthinned
Exhumed mantle (serpentinized)
Oceanic
Upper plate margin
Weakly thinned continental ribbons
Structures (map)
Transfer faults
Normal faults
Major thrusts
Thrusts
Structures (sections)
High angle normal fault (Low-ß structures)
Top basement detachment fault (High-ß structures)
Gravitational faults
Thrust faults
Future thrust faults
Western Approach
WA margin
Armorican SZ
Armorican margin
North Iberian margin
Landes High
Arzacq-Mauléon
Basque-Cantabrian
Pamplona F.
Toulouse F.
Ventaniella F.
CIZ
WALZ
Ebro block
Cameros
West Iberia margin
0 125 250 500 km
Bay of Biscay segment
Basque-Parentis segment
Pyrenean segment
(a) Bay of Biscay segment
Western Approach margin
undifferentiated mantle
Depth (km)
North Iberian margin
(b) Basque-Parentis segment
Basque Cantabrian basin
Landes High
Parentis basin
(c) Pyrenean segment
Ebro block
Arzacq-Mauléon basin
Horizontal scale (sections)
0 10 20 50 km

6.2. Role of rift inheritance for reactivation initiation and mountain building processes

6.2.1. Role of hyperextension for the heterogeneous reactivation

The Late Cretaceous to Late Oligocene convergence shows a heterogeneous distribution of compressional deformation. Based on the mapping of rift domains (fig 5) and on the restoration of the stage prior to compression (figs 11), we aim to discuss the role of the complex 3D rift architecture for the subsequent reactivation. In the previously described Bay of Biscay and Pyrenean segments, deformation is only accommodated in one of the rift systems (respectively the Bay of Biscay–Parentis and Pyrenean–Basque–Cantabrian rift systems). In the Western Bay of Biscay deformation is localized in between the exhumed mantle and hyperthinned domains (figs 7 and 11) while the oceanic domain is characterized by a diffuse and weak deformation (Thinon *et al.* 2002). In the Pyrenean segment former hyperextended rift basins are completely inverted and integrated in the orogen (fig 6; e.g. Roca *et al.* 2011 and reference therein). One of the major differences between the two settings is related to the initial size, structure and nature of basement in the different rift basins. In the Basque–Parentis segment, the occurrence of the two rift systems will lead to a “competition” during convergence. The Parentis basin is well preserved and only a minor reactivation can be observed in the central part of the basin where the crust is extremely thinned (Jammes *et al.* 2010a, 2010b, 2010c; Tugend *et al. submitted,* see chapter 1). In contrast the Basque–Cantabrian basin is completely inverted (fig 7). It is difficult to understand why the Parentis basin was preserved while the Basque–Cantabrian basin was reactivated. One possible interpretation could be that the Basque–Cantabrian basin was locally floored by exhumed serpentinized mantle (as proposed by Roca *et al.* 2011) and was consequently easier to reactivate. Westwards where mantle may be exhumed, compressional deformation is partitioned between the Bay of Biscay–Parentis and Pyrenean–Basque–Cantabrian rift systems, as indicated by the occurrence of the North Iberian Frontal thrust system (e.g. Derégnancourt & Boillot 1982; Roca *et al.* 2011; Fernández-Viejo *et al.* 2012) and by the tectonic uplift of the western termination of the Basque–Cantabrian basin (Gallastegui *et al.* 2002; Pedreira *et al.* 2003, 2007; Alonso *et al.* 2007).

6.2.2. Conceptual model to explain the evolution from inversion initiation to orogen formation

Thanks to the heterogeneous reactivation of the rift systems, we have access to the different stages of the compressional deformation enabling us to propose a conceptual model to explain the progressive integration of the rift-related domains into the orogen architecture, relying offshore and onshore observations (figs 11, 12 and 13).

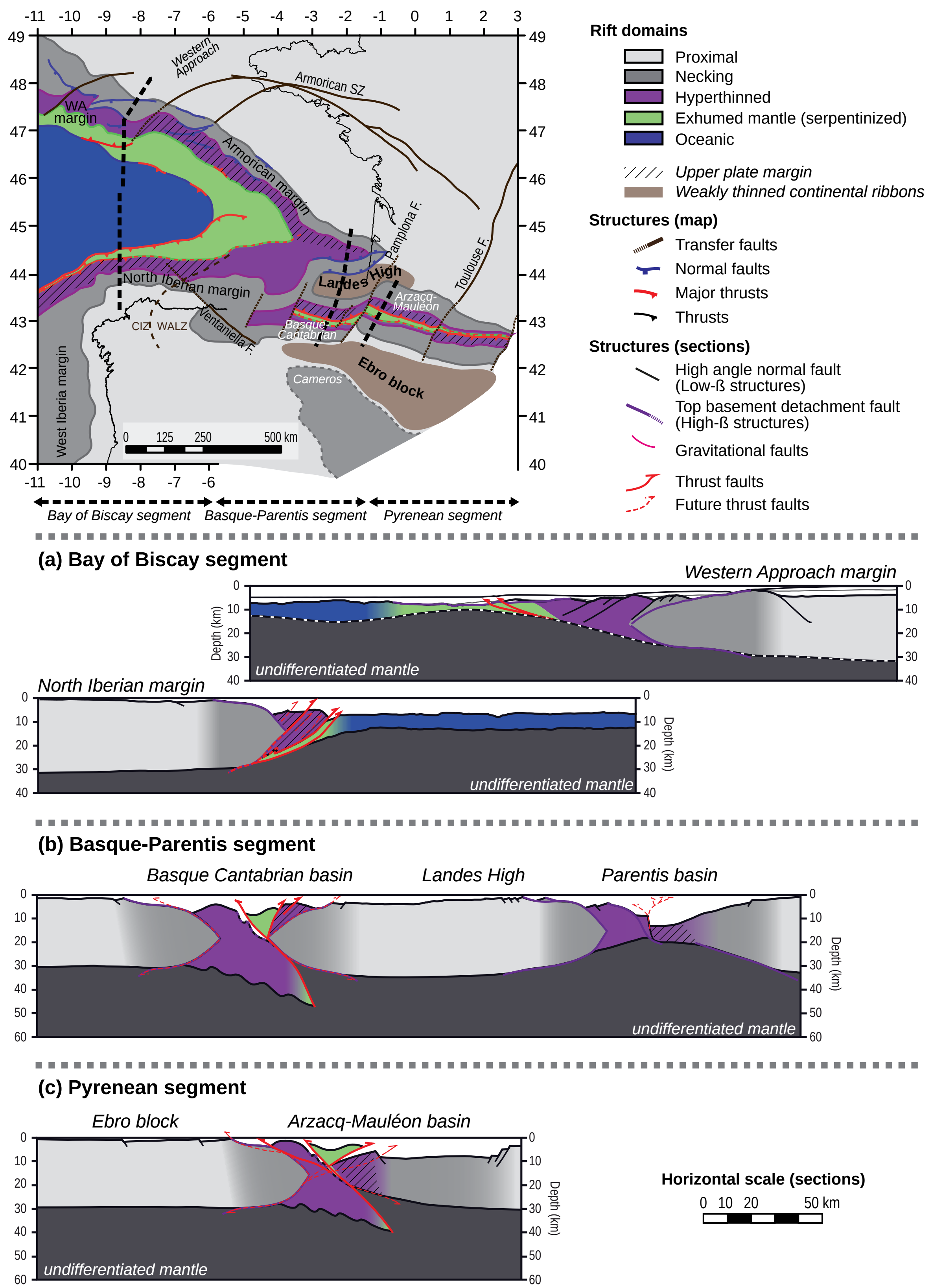
6.2.2.1. Initiation of reactivation: the role of exhumed serpentinized mantle

Only a weak reactivation is described on the Western Approach and Armorican margins mainly at the boundary between the exhumed mantle and hyperthinned domains (Thinon *et al.* 2001). The serpentinisation of exhumed mantle rocks may be responsible for the creation of a weak horizon where compressional deformation may initiate (as previously described for the West Iberian margin by Péron-Pinvidic *et al.* 2008; also discussed by Lundin & Doré 2011). At this stage the former top basement detachment fault may be used as a decoupling layer resulting in the possible sampling of mantle rocks. The former vergence of the detachment system may not play an important role for the onset of reactivation. In the Bay of Biscay, compressional deformation in the exhumed mantle domain can be observed on conjugate margins as exemplified by the Western Approach and North Iberian margin examples (figs 7a, 8, and 11; Thinon *et al.* 2001; Fernández-Viejo *et al.* 2012 and reference therein).

6.2.2.2. Accretionary prism and "subduction" stage: the role of hyperthinned domains

The progressive underthrusting of the former exhumed mantle domain will lead to the propagation of the compressional deformation into the hyperthinned domain. This domain is characterized by a strongly hydrated, brittle crust containing low friction minerals such as clays (Pérez-Gussinyé & Reston 2001; Reston & Pérez-Gussinyé 2007). At this stage the former hyperthinned domain can be progressively deformed resulting in structures similar to those observed in accretionary prisms, however, with a larger component of crustal material. Such structures have been observed along the North Iberian margin (e.g. Fernández-Viejo *et al.* 2012). In the former hyperextended basins, the initially smaller size of the system will lead to the progressive closure of exhumed mantle domain. Subsequently, the convergence of the two conjugate margins may trigger the initiation of a proto-subduction of the former hyperthinned domain as exemplified by the Arzacq–Mauléon and Basque–Cantabrian examples (fig 12). Shortcuts or incisions from new thrust faults may occur into the footwall of the former top basement detachment fault while it is progressively reactivated. They can result in the emplacement of hyperthinned and exhumed mantle rocks within the overlying compressional stack system formed predominantly by hanging wall material of the former rift related detachment faults. An example of a former hyperthinned domain incorporated into an accretionary prism has been described from the SW termination of the Taiwan Island (McIntosh *et al.* 2013).

Fig 12: Accretionary prism and "subduction" stage (Santonian-Campanian to Eocene) of the Bay of Biscay and Pyrenean rift systems. Sections are proposed in the (a) Bay of Biscay segment (b) Basque-Parentis segment and (c) Pyrenean segment. WA margin: Western Approach margin.


Rift domains
Proximal
Necking
Hyperthinned
Exhumed mantle (serpentinized)
Oceanic
Upper plate margin
Weakly thinned continental ribbons
Structures (map)
Transfer faults
Normal faults
Major thrusts
Thrusts
Structures (sections)
High angle normal fault
(Low-ß structures)
Top basement detachment fault
(High-ß structures)
Gravitational faults
Thrust faults
Future thrust faults
Western Approach
WA margin
Armorican SZ
Armorican margin
North Iberian margin
West Iberia margin
CIZ
WALZ
Ventaniella F.
Pamplona F.
Toulouse F.
Landes High
Arzacq-Mauléon
Basque-Cantabrian
Cameros
Ebro block
0 125 250 500 km
Bay of Biscay segment
Basque-Parentis segment
Pyrenean segment
(a) Bay of Biscay segment
Western Approach margin
undifferentiated mantle
Depth (km)
North Iberian margin
(b) Basque-Parentis segment
Basque Cantabrian basin
Landes High
Parentis basin
(c) Pyrenean segment
Ebro block
Arzacq-Mauléon basin
Horizontal scale (sections)
0 10 20 50 km

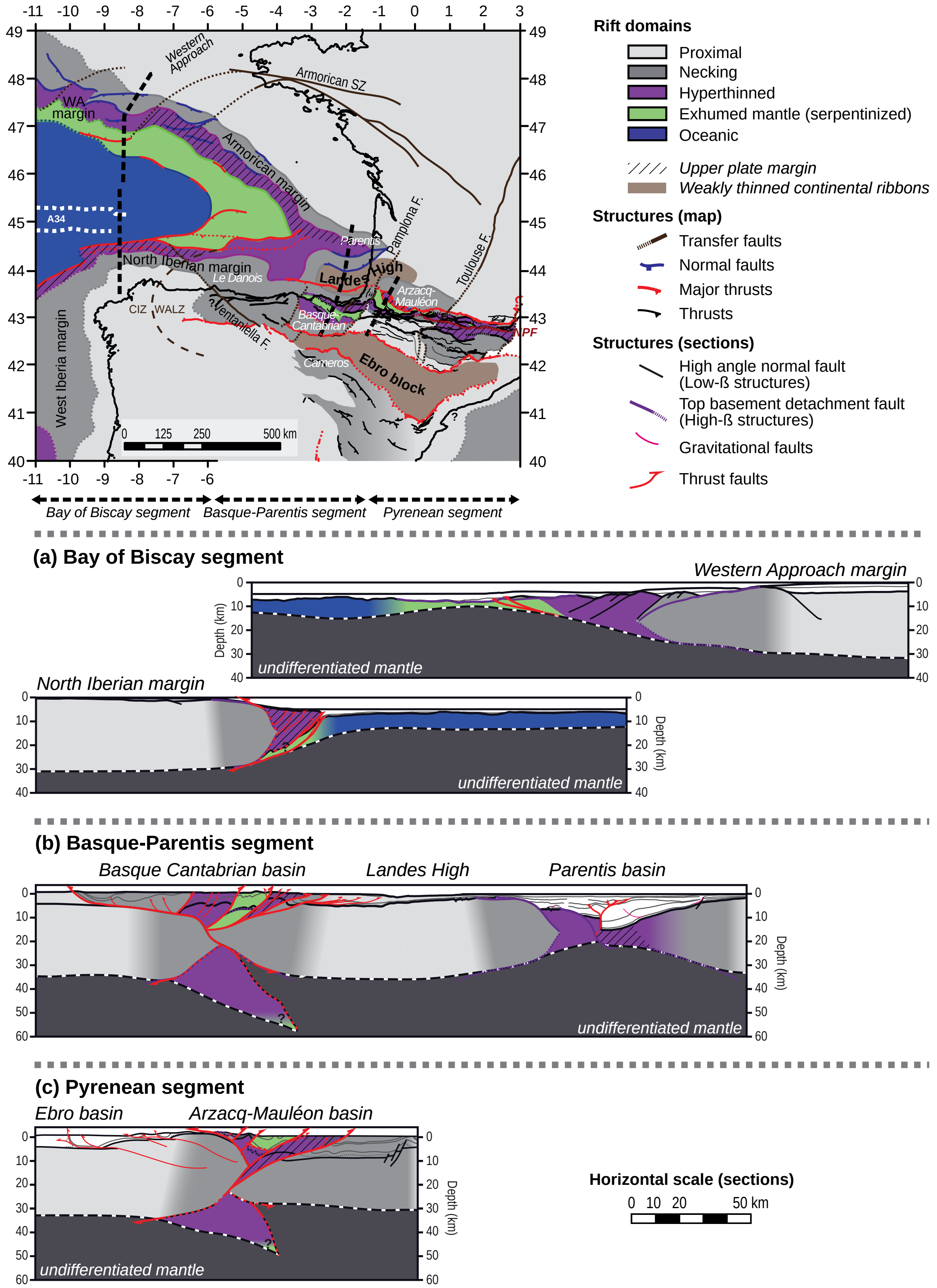
6.2.2.3. Continental collision and orogen formation: the role of the necking domain

In a later stage of convergence, after the early deformation and progressive subduction/ underthrusting of the hyperthinned and exhumed mantle domains, the arrival of the necking and proximal domains result in a major change in the compressional history. This stage is related to a new phase of deformation as observed and described for the Arzacq–Mauléon example. We interpret this second phase of deformation as related to the progressive collision between the former conjugate necking and proximal domains. The former set of conjugate top basement detachment faults interpreted to structure the necking domain (Mohn *et al.* 2012; Sutra *et al.* 2013) may be reactivated during onset of collision. At this stage, the former necking domains can act as buttress (Mohn *et al. submitted*). The collision between these two “buttresses” leads to a change in the mode of deformation. The subduction stage is mainly controlled by asymmetric deformation. In contrast the collision stage may induce symmetric processes including pro and retro thrusting (e.g. Pyrenean segment; figs 7c and 9). In the Pyrenean segment, a new phase of deformation can initiate the pop-up inversion of the Arzacq-Mauléon basin along north vergent thrusts (e.g. the Saint Palais, Sainte Suzannne and Ossau thrusts) and south-directed back-thrusting in the Axial zone (e.g. Gavarnie, Guarga thrust systems; Teixell 1998).

In this new interpretation of the orogen architecture (fig 13) the crustal roots imaged from geophysical methods (e.g. Daignières *et al.* 1994; Roure & Choukroune 1998; Pedreira *et al.* 2003, 2007) represent the subducted hyperthinned and exhumed mantle domain. In the case of the North Iberian margin the necking domain may play a similar role, representing the buttress for the accreted hyperthinned domain. Neves *et al.* (2009) proposed a similar model to explain the distribution of compressional deformation observed in the Tagus Abyssal plain (West Iberian margin) and interpreted the former necking zone as a crustal indentor.

This evolutionary model underlines two major points: (1) most of the deformation is located at the limits of rift domains implying that the former extensional structures separating rift domains were preferentially reactivated during compressional processes (Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Roca *et al.* 2011) and (2) the hyperthinned and exhumed mantle domains accommodate most of the deformation while the necking and proximal domains act as buttresses and consequently accommodate a relatively weak crustal shortening.

Fig 13: *Collision stage of the Bay of Biscay and Pyrenean rift systems. Sections are proposed in the (a) Bay of Biscay segment (b) Basque-Parentis segment and (c) Pyrenean segment to illustrate the buttress role of necking domains. WA margin: Western Approach margin.*


Rift domains
Proximal
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Top basement detachment fault (High-ß structures)
Gravitational faults
Thrust faults
Western Approach
Armorican SZ
WA margin
Armorican margin
A34
Parentis
North Iberian margin
Le Danois
Landes High
Pamplona F.
Toulouse F.
Arzacq-Mauléon
Basque-Cantabrian
NPF
CIZ
WALZ
Ventaniella F.
Cameros
Ebro block
West Iberia margin
0 125 250 500 km
Bay of Biscay segment
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(a) Bay of Biscay segment
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Depth (km)
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(b) Basque-Parentis segment
Basque Cantabrian basin
Landes High
Parentis basin
(c) Pyrenean segment
Ebro basin
Arzacq-Mauléon basin
Horizontal scale (sections)
0 10 20 50 km

6.3 General implications

6.3.1 Partitioning of the deformation and shortening rate estimations

The reinterpretation of the orogen architecture proposed in this study (fig 13) emphasises the role of the former rift domains (and not only of rift-inherited structures) for reactivation initiation and collisional orogen formation (figs 11, 12 and 13). It has additional implications for the restoration of the area and hyperextended rift systems in general. In order to propose a complete restoration of the former rift architecture, it is necessary to quantify the horizontal movement accommodated during the reactivation of rift structures (in particular top basement detachment faults) and to estimate the underthrust/subducted part of the hyperthinned and exhumed mantle domains (figs 12 and 13).

Many restorations are focused on the thick skin deformation observed in external parts of the orogens (e.g. in the Western Pyrenees; Teixell 1996, 1998). In the previous discussion, we showed that at least in the Western Pyrenees, this deformation may be minor and only corresponds to the final stage of compression (figs 12 and 13). Therefore only considering the restoration of structures in the external domains leads to underestimations of the initial shortening related to the underthrusting/subduction of the hyperthinned and mantle domains preserved in the internal parts of the orogen (fig 13).

Restorations of the crustal roots inferred from geophysical methods (e.g. refraction, gravity modelling) are also proposed (e.g. Muñoz 1992; Daignières *et al.* 1994; Pedreira 2004). Although some of these restorations unravelled previously thinned domains, they cannot restore quantitatively the underthrust/subducted portions of the former exhumed mantle domain based on crustal scale geophysical imaging alone.

Finally, in order to estimate the shortening accommodated at the scale of the overall Bay of Biscay and Pyrenees it is necessary to apprehend the partitioning of compressional deformation between two former rift systems as well as their complex 3D rift architecture (Bay of Biscay–Parentis and Pyrenees–Basque–Cantabrian rift systems, see also Roca *et al.* 2011). Therefore, the increasing deformation observed eastwards from the Bay of Biscay to the Pyrenees may not be related to an increasing shortening but related to the heterogeneous reactivation of a complex pattern of rift systems.

6.3.2. Nature of the North Pyrenean fault and plate kinematic implications

The NNE–SSW to NE–SW segmentation characterising the Pyrenean–Basque–Cantabrian rift system (see also Jammes *et al.* 2009; Roca *et al.* 2011) controls the spatial evolution of the rift domains mapped in this study at least from Late Aptian to Early Albian. The relative continuity of these structures across the former rift system preclude any strike-slip or major left lateral movements after Late Aptian time (as previously discussed by Jammes *et al.* 2009, 2010a).

The North Pyrenean fault was interpreted as a major crustal discontinuity defining the plate boundary between the European and the Iberian plates (Mattauer 1968; Choukroune & Mattauer 1978; Choukroune & ECORS Team 1989). This structure was supposed to accommodate a left lateral displacement of the Iberian plate during mid-Cretaceous time (e.g. Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Olivet 1996). However, the nature and lateral continuity of this fault was questioned by several authors (e.g. Canérot *et al.* 2001 and reference therein; Canérot 2008).

In the light of our mapping results, the North Pyrenean fault system coincides with the occurrence of the exhumed mantle domain. In the east this domain corresponds to a narrow, sub-vertical zone that can be mapped as a highly deformed zone (see geological sections in Lagabrielle *et al.* 2010) indirectly also characterized by a HT–LP metamorphism attributed to the extreme rift-related thinning (e.g. Goldberg & Leyreloup 1990). Controversies on the western continuation of this structure (Canérot *et al.* 2001) correspond with the progressive widening of the exhumed mantle domain characterized by a weaker compressional overprint. Therefore, we suggest that the North Pyrenean Fault represents a “suture” zone related to the inversion of the former hyperextended and exhumed mantle domains.

7. Conclusions

The aim of this study was to characterize the spatial and temporal evolution of hyperextended rift systems and their progressive reactivation. We used gravity inversion, seismic interpretation and field mapping to propose a map of the rift domains from the offshore Bay of Biscay to their onshore fossil remnants preserved in the Pyrenees. The key observations and learnings of this work on the formation and deformation of hyperextended systems can be summarized as follow:

(1) Two diachronous and spatially distinct rift systems can be described: the Bay of Biscay–Parentis and the Pyrenean–Basque–Cantabrian rift systems characterized by different evolutionary stages.

(2) These rift systems are characterized by a strong segmentation inherited from the complex structuration of the Variscan orogen and/or from the Permo-Carboniferous post-Variscan history. This segmentation may control lateral variations in the architecture of the rift systems but does not control the formation of the oceanic domain. This last observation may indicate that rift systems are sensitive to inherited structures and rheology while oceanic domains are more likely to be controlled by asthenospheric processes.

(3) The NE–SW to NNE–SSW segmentation observed across the Pyrenean–Basque–Cantabrian rift system preclude the accommodation of an important left lateral movement after Late Aptian to Early Albian time in this rift system as proposed from previous studies.

(4) Different steps in compressional deformation are related to the rift domain inheritance. At first reactivation is initiated in the serpentinized layers of the exhumed mantle domain. The progressive closure of this domain may lead to the propagation of compressional deformation to the hyperthinned domain resulting in the formation of an accretionary prism. In the former rift system preserved onshore, the proximity of the conjugate system enables the underthrusting/subduction of the former hyperthinned domains. The final architecture of the orogen is acquired during the collision of the former necking domains representing conjugate buttresses.

Finally, we believe that the complex evolution of the Bay of Biscay and Pyrenees may be used as an analogue to understand the partitioning of extensional deformation in other V-shaped rift basins. Additionally, the role of rift inheritance for reactivation and orogen formation may provide new insights for the formation of collisional orogens.

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CHAPITRE 3

Le troisième chapitre discute du contexte de formation de la limite de plaque entre Ibérie et Europe. Un modèle d'évolution et de distribution de la déformation extensive est proposé et mis en relation avec l'architecture 3D complexe des systèmes de rift hyper-amincis préservés dans le Golfe de Gascogne et les Pyrénées.

La synthèse de précédentes et nouvelles observations souligne la complexité de la transition entre les plaques Ibérique et Européenne et son évolution polyphasée. Les reconstructions paléogéographiques régionales imposent un mouvement senestre de l'Ibérie dont les étapes, l'amplitude et les implications pour la limite de plaque Ibérie-Europe restent fortement controversées. Les résultats présentés dans ce chapitre suggèrent que le déplacement de l'Ibérie est accommodé par un épisode de rift oblique essentiellement partitionné entre deux systèmes de rift spatialement distincts, orientés NO-SE (les systèmes Golfe de Gascogne–Parentis et Intra–Ibérique). Entre ces deux systèmes de rift, séparés par les blocs d'Ebro et des Landes la déformation semble plus diffuse et partitionnée entre plusieurs bassins étroits et orientés E-O. L'initiation du processus d'accrétion océanique dans le Golfe de Gascogne à l'Apto-Albien est associée à un changement régional des contraintes extensives souligné par la segmentation NE-SW observée dans l'ancien système de rift préservé dans les Pyrénées. Un épisode de déformation extensive diffus est enregistré dans les bassins de rift Pyrénéens et Basque-Cantabre et interprété comme une tentative avortée de localisation de limite de plaque.

Cette interprétation souligne l'architecture et l'évolution complexe de la transition entre Ibérie et Europe et remet en question le rôle de la faille Nord Pyrénéenne comme limite de plaque entre Ibérie et Europe.

Publication 3: in preparation for Geology

FORMATION AND DEFORMATION OF THE BAY OF BISCAY–PYRENEAN DOMAIN: IMPLICATION FOR THE NATURE, KINEMATICS AND TIMING OF THE IBERIAN–EUROPEAN PLATE BOUNDARY

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ABSTRACT

The understanding of how the continental lithosphere thins and breaks up to form a new plate boundary represents a long standing problem in Earth Science. We focus on the European–Iberian plate boundary whose tectonic context is strongly debated.

We propose that the Late Jurassic to Mid Cretaceous left lateral displacement of the Iberian plate is accommodated within a wide zone of deformation mainly partitioned between two rift systems propagating in opposite directions (the Bay of Biscay–Parentis and Iberian Range rift system). This interpretation has some major implications for the evolution of the Pyrenean domain that is no-longer characterized by a major strike slip to transtensional deformation. In contrast, the Pyrenean–Basque–Cantabrian rift system is mainly activated during the initiation of seafloor spreading in the Western Bay of Biscay. The diffuse deformation observed onshore may be interpreted as related to a tentative propagation of a plate boundary that failed to localize. We believe that this work may provide insights on the processes preceding break-up and initiating the formation of segmented and transform rifted continental margins. Additionally, this study may further improve the understanding of the processes leading to the formation and localization of plate boundary.

1. Introduction

The formation, evolution and destruction of plate boundaries represent one of the main challenges in the understanding of plate tectonics. The combination of high-resolution seismic imaging with drill-hole observations unraveled the occurrence of extremely thinned continental crust and exhumed mantle extending over hundreds of kilometers continentward of the first oceanic crust related to lithospheric break-up (e.g. Iberia, Tucholke *et al.* 2007 or Péron-Pinvidic & Manatschal 2009). These hyperextended domains seem to form many continental rifted margins worldwide (Reston 2009 and reference therein), however the implications of this discovery for the initiation and localisation of divergent plate boundaries remains poorly understood.

In this study, we investigate the tectonic setting preserved at the transition between the European and Iberian plates. The initiation of lithosphere thinning processes consecutive to the displacement of the proto-Iberian plate relatively to Europe resulted in the formation of hyperextended rift domains (e.g. the North and South Biscay margins: Thinon *et al.* 2003; Roca *et al.* 2011; in the Pyrenean and Cantabrian domain: Pedreira *et al.* 2007; Lagabrielle & Bodinier 2008; Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Clerc *et al.* 2012) and ultimately oceanic crust in the Western Bay of Biscay in the Mid Cretaceous (Montadert *et al.* 1979; Boillot 1984). The northward movement of the African plate initiated in the Late Cretaceous progressively reactivating the former plate boundary, resulted in the partial closure of the Bay of Biscay and the formation of the Pyrenean orogen.

The tectonic setting related to the thinning and break-up of the continental lithosphere in the Western Bay of Biscay remains strongly debated, resulting in controversial interpretations of the timing, kinematic and location of the boundary between the European and Iberian plates. In this paper, we investigate the spatial and temporal evolution of the rift systems preserved at the transition between Europe and Iberia. Based on these observations, we aim to discuss the strain distribution and partitioning of the deformation at the scale of the plate boundary from its initiation to reactivation. Finally, we believe that results of this work may provide general insights to understand the initial stages of the formation and reactivation of plate boundaries.

2. A controversial plate kinematic context

The structuration of the overall region is strongly controlled by the formation of the Variscan orogen and a Late Carboniferous to Early Permian orogenic collapse (Burg *et al.* 1994a, 1994b). A Triassic to early Jurassic rift event resulted in the formation of intra-continental basins partly reactivating Palaeozoic structures (e.g. Curnelle *et al.* 1982). The kinematics associated to the Late Jurassic to Early Cretaceous rift event resulting in sea-floor spreading initiation in the Western Bay of Biscay is strongly debated. The main disagreements about the movement of the proto-Iberian plate concerns the amount of left lateral displacement, timing of events and interpretation of rotation poles (fig1). As a result, contrasting plate kinematic scenarios are proposed as illustrated by hypotheses ranging from a back-arc setting (e.g. Sibuet *et al.* 2004; Vissers & Mejers 2012) to a strike-slip or transtensional context (e.g. Le Pichon *et al.* 1971; Mattauer & Seguret 1971; Choukroune & Mattauer 1978).

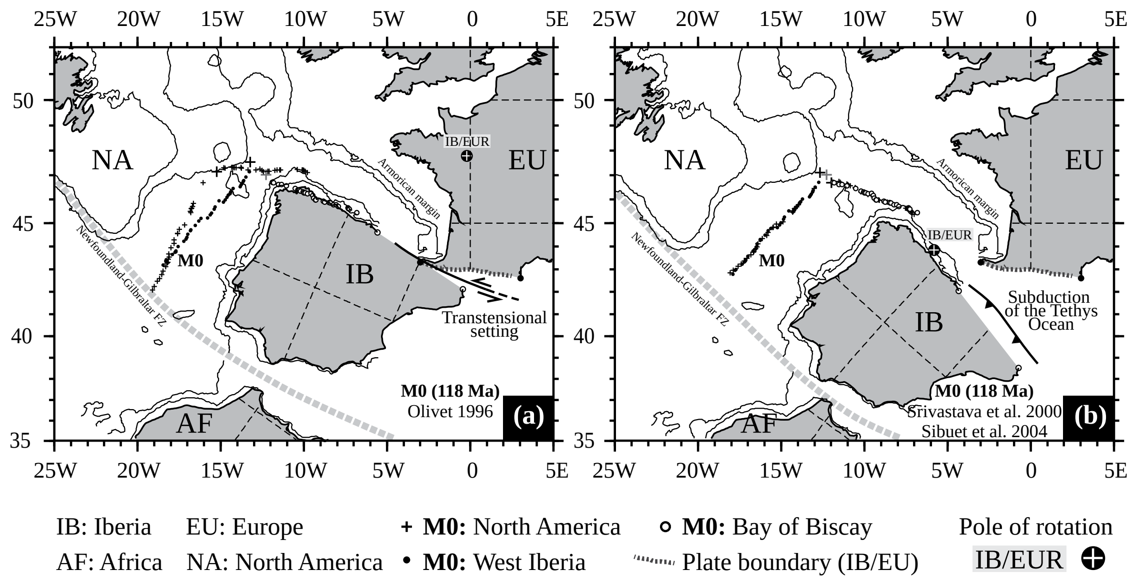


Fig. 1*: Two different plate kinematic restorations at Chron M0 (118 Ma) (a) Olivet 1996 (b) Srivastava et al. 2000; Sibuet et al. 2004. The pole of rotation between Europe and Iberia is different between the two scenarios as well as the implications for the Pyrenean domain (transtensional setting vs subduction). The same restoration as b) is proposed by Vissers & Mejers 2012 at M0 (121.2 Ma).*

Main uncertainties are related to the identification, interpretation and restoration of magnetic anomalies of the M–series in the Bay of Biscay and North Atlantic in general (M3–M0, 126 to 118.5 Ma). Indeed, most restorations based on magnetic anomalies only consider minor pre-break up movements, an assumption that is questioned by the discovery of hyperextended domains that can be mapped over hundreds of kilometres continentward of the first magnetic anomaly related to break-up (Tucholke *et al.* 2007; Péron-Pinvidic & Manatschal 2009). Taking into account pre-breakup movements between the Iberia and Newfoundland margins the left lateral movement of the proto-Iberian plate may have been already initiated in the Late Jurassic resulting in a transtensional setting along the European and Iberian plate (Wortmann *et al.* 2001; Schettino & Scotese 2002; Canérot 2008; Jammes *et al.* 2009; 2010).

Furthermore, recent studies proposed different origins for the formation of the M3 to M0 anomalies in the Iberia–Newfoundland margins. They are either interpreted as related to mantle exhumation (Sibuet *et al.* 2007) or to an excess magmatic event related to lithospheric breakup, overprinting a former exhumed mantle domain (Bronner *et al.* 2011 and references therein). Norton *et al.* (2007) showed large inconsistencies between plate kinematic restorations based on M0 and geological observations. These authors suggested that these anomalies may not represent isochrones and consequently cannot be used for plate kinematic restorations, a point that was further discussed by Bronner *et al.* (2012).

3. Spatial and temporal evolution of rift systems: insights from subsidence and deformation history

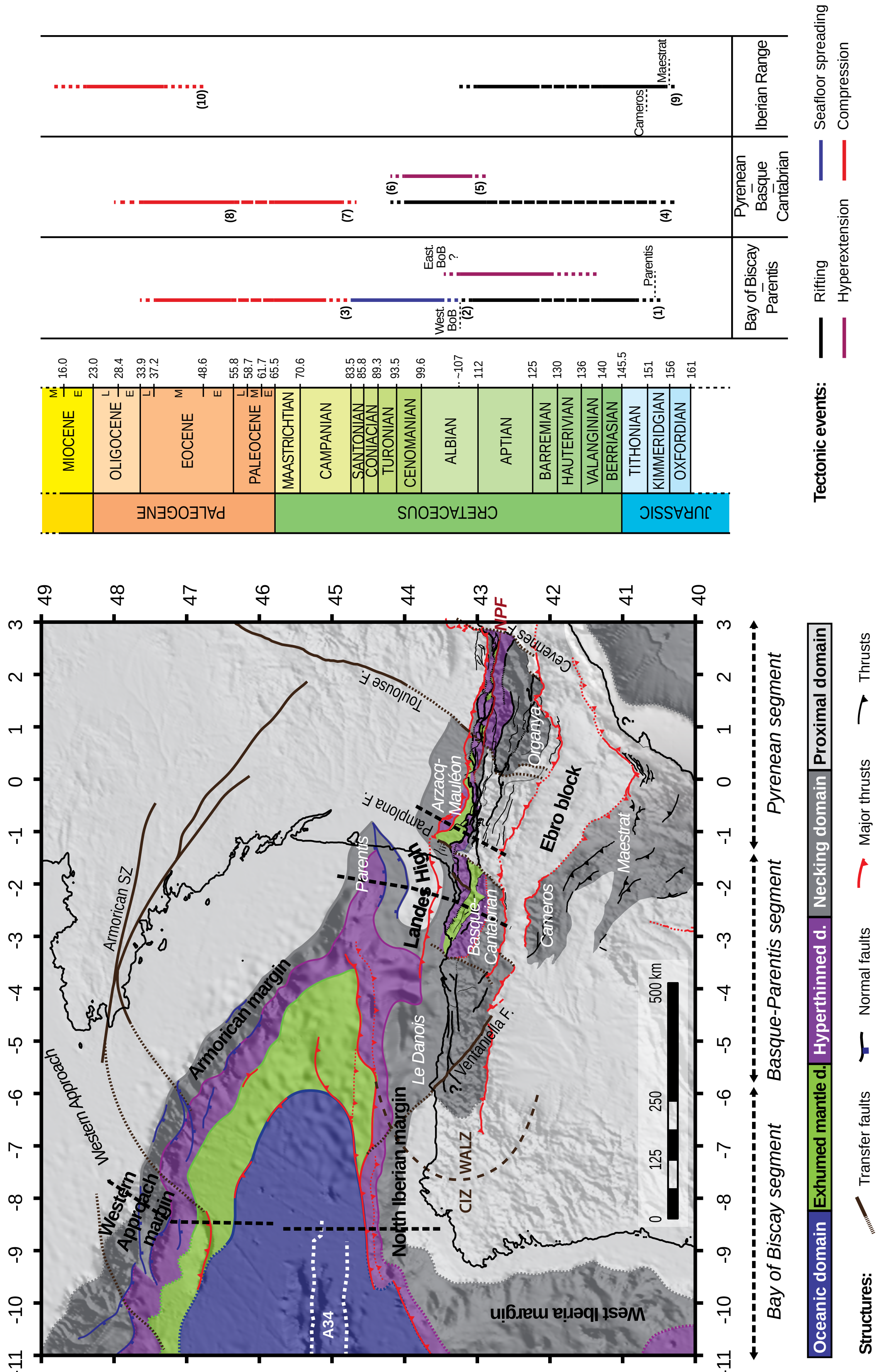
This study relies on the compilation of a new map that illustrates the rift domains and their reactivation along the European–Iberian plate boundary (fig2). The approach underlying this map combines both geophysical and geological observations to characterize the spatial evolution of rift domains from the offshore Bay of Biscay to their deformed analogues preserved onshore along the Cantabrian–Pyrenean mountain range (Tugend *et al.* submitted, see chapter 1). The map documents the spatial evolution of different rift systems showing distinct subsidence and deformation histories during the Latest Jurassic to mid Cretaceous rifting phase. We distinguish the Bay of Biscay–Parentis, the Pyrenean–Basque–Cantabrian and Iberian–Range rift systems (see also Salas & Casas 1993 and reference therein; Vergés & García-Senz 2001; Roca *et al.* 2011).

3.1. Bay of Biscay-Parentis rift system

The Bay of Biscay is formed by different extensional segments including an oceanic domain to the West at the junction with the Atlantic Ocean to an exhumed mantle and hyperthinned domains in the Parentis basin to the East (Pinet *et al.* 1987; Bois & Gariel 1994; Tomassino & Marillier 1997; Jammes *et al.* 2010). NE–SW transfer faults and E–W to NW–SE trending extensional faults can be mapped in the rift system observed along the northern part of the Bay of Biscay (Derégnaucourt & Boillot 1982; Thinon 1999; Thinon *et al.* 2003).

Onset of subsidence is difficult to date in the Bay of Biscay, however, in the Parentis basin, a Late Jurassic subsidence is described (Brunet 1994) and indicated by a thickening of Jurassic sequences towards the centre of the basin (Bois & Gariel 1994; Jammes *et al.* 2010).

Hyperextensional processes may have initiated in the Early Cretaceous (around Berriasian to Barremian) and persisted until Late Aptian time, as suggested by Late Aptian sediments overlying exhumed mantle (3B layer in Thinon *et al.* 2002). In the Western part of the Bay of Biscay, seafloor spreading may have started as early as latest Aptian to earliest Albian time (Montadert *et al.* 1979) simultaneous with onset of seafloor spreading between Iberia and Newfoundland (Tucholke & Sibuet 2007).


Bay of Biscay segment
Basque-Parentis segment
Pyrenean segment
Oceanic domain
Exhumed mantle d.
Hyperthinned d.
Necking domain
Proximal domain
Structures:
Transfer faults
Normal faults
Major thrusts
Thrusts
Tectonic events:
Rifting
Hyperextension
Seafloor spreading
Compression
Bay of Biscay – Parentis
Pyrenean – Basque – Cantabrian
Iberian Range
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
APTIAN
BARREMIAN
HAUTERIVIAN
VALANGINIAN
BERRIASIAN
TITHONIAN
KIMMERIDGIAN
OXFORDIAN
PALEOGENE
CRETACEOUS
JURASSIC
West Iberia margin
North Iberian margin
Armorican margin
Western Approach margin
Western Approach
Armorican SZ
Le Danois
Landes High
Parentis
Basque-Cantabrian
Arzacq-Mauléon
Organya
Ebro block
Cameros
Maestrat
Pamplona F.
Toulouse F.
Cevennes F.
Ventaniella F.
NPF
CIZ
WALZ
A34
500 km

The onset of convergence is recorded in the Bay of Biscay by a Late Cretaceous to Paleocene unconformity sealing a heterogeneous and diffuse deformation (Thinon *et al.* 2001).

3.2. Pyrenean-Basque-Cantabrian rift system

The Pyrenean-Basque-Cantabrian rift system is delimited to the North by the Landes high (see also Roca *et al.* 2011) and to the South by the Ebro block (see also Salas *et al.* 2001; fig 2). The spatial evolution of this hyperextended rift system is controlled by NE–SW transfer faults (see also Jammes *et al.* 2009; Roca *et al.* 2011) relaying rift segments to the North or to the South (ex: the Basque–Cantabrian basin).

In the Late Jurassic (latest Kimmeridgian to Tithonian) an increase in subsidence is recorded in the WNW–ESE directed Arzacq and Tarbes basins north of the Pyrenees (Désegaulx & Brunet 1990) as inferred from the facies and environment distribution proposed by the BRGM (1974). In the Basque Cantabrian basin, the first syn-rift deposits are Late Tithonian in age deposited in elongated grabens bounded by E–W to NW–SE trending basement faults (Martin-Chivelet *et al.* 2002 and reference therein).

The Late Aptian to Early Albian is characterized by a progressive increase in subsidence history persisting until Mid- to Late Albian (e.g. Basque–Cantabrian basin: García-Mondéjar *et al.* 1996, 2005; Martin-Chivelet *et al.* 2002; Pyrenean basins: Debroas 1987, 1990). In the Pyrenean basins, local mantle exhumation may have occurred from Mid-Albian to Early Cenomanian indirectly indicated by the reworking of mantle bodies in Mid Albian to Cenomanian breccias (e.g. Urdach breccia in Western Pyrenees: Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Debroas *et al.* 2010 or Aulus breccias in Eastern Pyrenees: Clerc *et al.* 2012).

Onset of compressional deformation is recorded by a regional uniformity in Santonian sediments (Garrido-Megías & Rios 1972; McClay *et al.* 2004) whereas the main collision phase is Eocene in age (Muñoz 2002; Vergés *et al.* 2002).

Fig. 2: *To the left, map of the rift domains preserved at the transition between the European and Iberian plates. Three rift systems are distinguished: the Bay of Biscay–Parentis, the Pyrenean–Basque–Cantabrian and Intra-Iberian. To the right, deformation history of the different rift systems. (1) Brunet 1994; (2) Montadert et al. 1979; (3) Thinon et al. 2002; (4) Désegaulx & Brunet 1990; Martin-Chivelet et al. 2002 ; (5) Jammes et al. 2009 ; Lagabrielle et al. 2010 ; (6) Clerc et al. 2013 (7) Garrido-Megías & Ríos 1972 ; McClay et al. 2004 ; (8) Muñoz 2002 ; Vergès et al. 2002 ; (9) Salas et al. 2001 ; Capote et al. 2002 ; Mas et al. 2002 ; (10) Salas et al. 2001 ; Capote et al. 2002.*

3.3. Iberian range rift system

This triangular shaped rift system is oriented in a NW–SE direction. In contrast to the other rift systems, no evidence of extreme crustal thinning is reported. Based on the restoration of Salas & Casas 1993, at the end of the Mesozoic rifting the crust was thinned to a maximum of about 22 km. Rift basins are bounded by NW–SE and NE–SW structures interpreted either as transfer or normal faults successively activated (e.g. in the Cameros basin: Platt 1990; Maestrat basin: Soria 1997; Soria *et al.* 2000).

The subsidence history of this area is relatively well-known from the analysis of subsidence curves (Salas & Casas 1993; Mas *et al.* 1993). The onset of rifting is diachronous at the scale of the rift system. In the SE (Maestrat basin), syn-rift subsidence is already initiated in the Late Oxfordian whereas it is delayed until Early Tithonian to the NW in the Cameros basin (Salas *et al.* 2001; Capote *et al.* 2002; Mas *et al.* 2002). At this stage, the rift system is opened towards the Tethysian domain (Mas *et al.* 1993, 2002; Salas & Casa 1993; Salas *et al.* 2001 and reference therein). Around Aptian- to Mid Albian time, tectonic subsidence progressively decreases followed by a Late Cretaceous thermal subsidence (Salas & Casas 1993; Capote *et al.* 2002).

In spite of a minor compressional event locally recorded in Early Albian (Cameros basin, Casas & Gil 1998) most of the compressional deformation initiated in the Middle to Late Eocene and continued until Late Oligocene to Early Miocene (Salas *et al.* 2001; Capote *et al.* 2002 and reference therein).

4. Strain distribution, partionning and evolution of the European-Iberian plate boundary

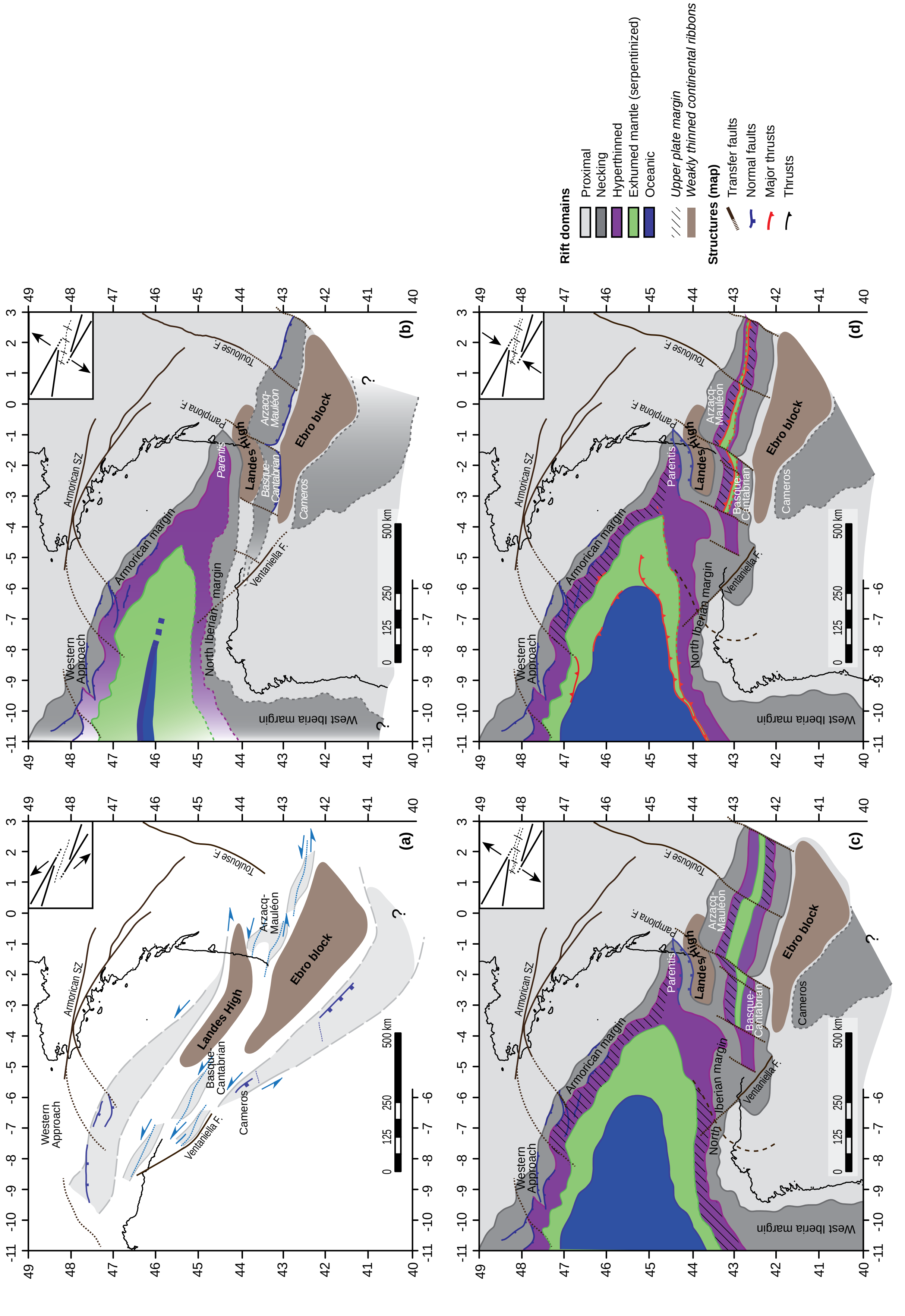
The spatial and temporal evolution of the rift systems enables us to distinguish several phases in the deformation history and discuss the associated strain distribution (fig 3).

4.1. Rift initiation in an oblique setting (Kimmeridgian-Tithonian to Late Aptian-Early Albian)

The late Jurassic to Early Cretaceous displacement of the proto-Iberian plate (e.g. Wortmann *et al.* 2001; Schettino & Scotese 2002; Canérot 2008; Jammes *et al.* 2009; 2010) seems to be accommodated within a wide corridor of deformation including to the north the Bay of Biscay–Parentis and southern Iberian range rift systems (see also Canérot 2008). The NW-SE divergence related to the movement of the proto-Iberian plate (Olivet 1996) is interpreted to result in an oblique rifting episode. The deformation is mainly partitioned between the Bay of Biscay–Parentis and Iberian Range representing two proto V-shaped rift systems that are progressively propagating in opposite directions (fig. 3a).

The Bay of Biscay is opened towards the Atlantic from the Late Jurassic (Durand-Delga, 1973) and may be propagating towards the SE as suggested by the less mature evolution of the Parentis basin. In contrast, the Iberian Range rift system seems to be opened towards the Tethyian domain (Mas *et al.* 1993, 2002; Salas & Casas 1993) and propagating towards the NW. In that case, the Ventaniella fault at the termination of the rift system may represent the trace of the tentative propagation towards the northwest. This structure only accommodated a small displacement. The oblique rift setting of the two systems may partly explain the similar segmentation characterized by rift structures and transfer faults oriented NW–SE to NE–SW. The two rift systems are spatially disconnected separated by the Landes High and Ebro block that represent weakly thinned continental ribbons probably characterized by an old and rigid basement inherited from the pre-rift history (e.g. Matte 1991; Lefort *et al.* 1997).

This interpretation contrasts with previous models that accommodate the deformation within a transtensional corridor or along the North Pyrenean fault (e.g. Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Choukroune & Mattauer 1978) in the Pyrenean–Basque–Cantabrian rift system. We suggest that only a minor E–W deformation is accommodated within individualized rift basins of the future Pyrenean–Basque–Cantabrian rift system that may be elongated in an E–W direction (e.g. Arzacq and Tarbes basins; BRGM 1974).


(a)
(b)
(c)
(d)
Rift domains
Proximal
Necking
Hyperthinned
Exhumed mantle (serpentinized)
Oceanic
Upper plate margin
Weakly thinned continental ribbons
Structures (map)
Transfer faults
Normal faults
Major thrusts
Thrusts
Western Approach
Armorican SZ
Armorican margin
North Iberian margin
West Iberia margin
Toulouse F.
Pamplona F.
Ventaniella F.
Landes High
Parentis
Basque-Cantabrian
Arzacq-Mauléon
Ebro block
Cameros
500 km

A progressive increase in accommodation space is recorded in the Bay of Biscay and Iberian–Range rift systems until late Aptian to Mid Albian time. However, while mantle exhumation is suggested to occur in the Bay of Biscay (Thinon *et al.* 2003; Tugend *et al.* submitted, see chapter 2), the Iberian Range rift system undergoes only minor thinning (Salas & Casas 1993).

4.2. Break-up and tentative localization of the plate boundary (Late Aptian-Early Albian to Late Cretaceous)

The initiation of sea-floor spreading processes in the Western Bay of Biscay at Aptian-Albian time (Montadert *et al.* 1979) is related to a major change in the subsidence history of the rift systems. In the Iberian–Range rift system, tectonic subsidence progressively decreases (Salas & Casas 1993; Capote *et al.* 2002), suggesting a cessation of rifting within this domain. In contrast, from late Aptian to Early Cenomanian, an increasing subsidence is recorded in the Pyrenean-Basque-Cantabrian rift system (e.g. Basque–Cantabrian basin: García-Mondéjar *et al.* 1996, 2005; Martin-Chivelet *et al.* 2002; Pyrenean basins: Debroas 1987, 1990). This change may be interpreted as a delocalization of extensional deformation during the initiation of hyperextensional processes in the Pyrenean-Basque-Cantabrian rift system starting from Late Aptian to Early Albian. The NE–SW segmentation of the Pyrenean-Basque-Cantabrian rift system underlines a change in the overall divergence orientation that may be related to the localization of sea-floor spreading processes in the Western Bay of Biscay (fig. 3b). At the scale of the European and Iberian plates, the transition from localized sea-floor spreading to a diffused extensional deformation propagating onshore may be interpreted as the tentative propagation of a proto plate boundary enhanced by the northward propagation of the Atlantic Ocean (fig. 3c).

Fig. 3: *Evolutionary model of the European-Iberian plate transition. The partitioning of the deformation is illustrated in inset cartoons. (a) Oblique rifting stage (Late Jurassic to Late Aptian) (b) Lithospheric break-up and initiation of seafloor spreading (Aptian–Albian) (c) Tentative localization of the plate boundary (prior to Santonian) (d) Subduction initiation (Late Cretaceous)*

4.3. Subduction initiation (Santonian-Campanian to Eocene)

A late Cretaceous onset of compressional deformation is recorded in the Bay of Biscay and Pyrenean–Basque–Cantabrian rift system (Garrido-Megías & Rios 1972; Thinon *et al.* 2001; McClay *et al.* 2004) but not in the Iberian Range (fig. 3d). As previously discussed this initial deformation is interpreted as to the progressive subduction/underthrusting of the exhumed mantle and hyperthinned crust. The fact at this deformation is not observed in the Iberian Range may be related to the relatively poor thinning of the continental crust (Salas & Casas 1993) and the absence of developped hyperextended domains (in particular exhumed mantle) where deformation may preferentially initiate (Tugend *et al.* submitted, see chapter 2).

4.4. Continental Collision (Eocene-Oligocene)

The generalization of collisional processes is initiated in Eocene. This stage is related to the collision of the former (weakly thinned) necking and proximal domains of the conjugate margins of the Pyrenean–Basque–Cantabrian rift system. The later initiation of inversion in the Iberian Range rift system may be related to proggressive migration of compressional deformation towards the Betics to the South.

5. Implications/Conclusions

The interpretation proposed in this work brings new insights on the strain distribution and partitioning of the deformation between the European and Iberian plates from rift initiation to progressive reactivation. The polyphase evolution described further illustrates the importance of pre-breakup deformation. The lateral displacements acquired prior to lithospheric breakup remain, however, difficult to determine and can only be quantified into a plate kinematic model if the pre-breakup deformation of the southern North Atlantic can be described. Plate restorations based on restoration of magnetic anomalies alone (e.g. Olivet 1996; Sibuet *et al.* 2007; Vissers *et al.* 2012) are likely to misinterpret the amount and timing of movements between plates.

The boundary between Europe and Iberia is more complex than previously assumed. It is characterized by several rift systems spatially disconnected by weakly thinned crustal blocks (Ebro and Landes high; see also Salas *et al.* 2001; Roca *et al.* 2011). In the western Bay of Biscay the initiation of oceanic spreading coincides with a diffuse NE–SW extensional deformation in rift basins from the Pyrenean–Basque–Cantabrian rift system. This event may be interpreted as the tentative initiation of a plate boundary propagating eastwards into continental lithosphere. However, to the east in the present-day Pyrenees, the deformation failed to localize.

Additionally, the European–Iberian plate boundary is at the junction between the proto-Atlantic and Tethysian rift systems and its complex polyphase evolution remains to be fully integrated in the understanding of the northwards propagation of the Atlantic Ocean but also of the evolution of the Tethysian–Alpine system.

Finally, the oblique rifting proposed to characterize the formation of the Bay of Biscay may partly represent an analogue to further constrain the mechanisms preceding break-up and related to the formation of transform and segmented rifted continental margins observed worldwide.

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SYNTHÈSE ET DISCUSSION

Ma thèse a eu un objectif double : dans un premier temps décrire l'évolution spatiale et temporelle des mécanismes d'extension lithosphérique qui forment les systèmes de rift " hyper-amincis " et les marges passives peu magmatiques. Un autre but a été de mieux caractériser le rôle de l'architecture héritée du rift lors de la formation des orogènes de collisions. De nombreuses études ont conceptualisé l'évolution polyphasée et l'amincissement extrême de la croûte observé dans les marges peu magmatiques actuelles (e.g. Lister *et al.* 1986, 1991 ; Reston 1988, 2007 ; Brun & Beslier 1995 ; Driscoll & Karner 1998 ; Whitemarsh *et al.* 2001 ; Mohn *et al.* 2012). Cependant le partitionnement de cette déformation dans le temps et dans l'espace reste peu contraint dans les exemples actuels et fossiles même si certains concepts ont déjà été proposés (e.g. Lavier & Manatschal 2006).

Ce travail utilise le Golfe de Gascogne et les Pyrénées (à la frontière entre les plaques Européenne et Ibérique) comme laboratoire naturel. Cette région préserve un ancien domaine de rift qui a abouti à la formation de domaines hyper-amincis qui sont à présent préservés à la fois en mer (e.g. bassin de Parentis, Pinet *et al.* 1987 ; Bois & Gariel 1994 ; Jammes *et al.* 2010a, 2010b, 2010c) et à terre (Bassins d'Arzacq–Mauléon : Jammes *et al.* 2009 ; Lagabrielle *et al.* 2010 ; Basque–Cantabre : Pedreira *et al.* 2007 ; Roca *et al.* 2011 ; Aulus : Lagabrielle & Bodinier 2008 ; Lagabrielle *et al.* 2010 ; Clerc *et al.* 2012). Ces domaines ont par suite été partialement réactivés et même intégrés à l'orogène Pyrénéenne. La structuration observée implique d'importantes variations latérales de l'architecture de ces systèmes de rift, suggérant un partitionnement complexe de la déformation extensive, mais également lors des processus de réactivation.

La nouvelle cartographie de ces systèmes de rift préservés à la transition entre Europe et Ibérie a représenté un enjeu majeur de ma thèse et constitue le fil conducteur de cette discussion (fig 1). Dans cette partie, les questions scientifiques à l'origine de ce mémoire et présentées dans l'introduction sont discutées à la lumière des résultats obtenus pour l'exemple du Golfe de Gascogne et des Pyrénées.

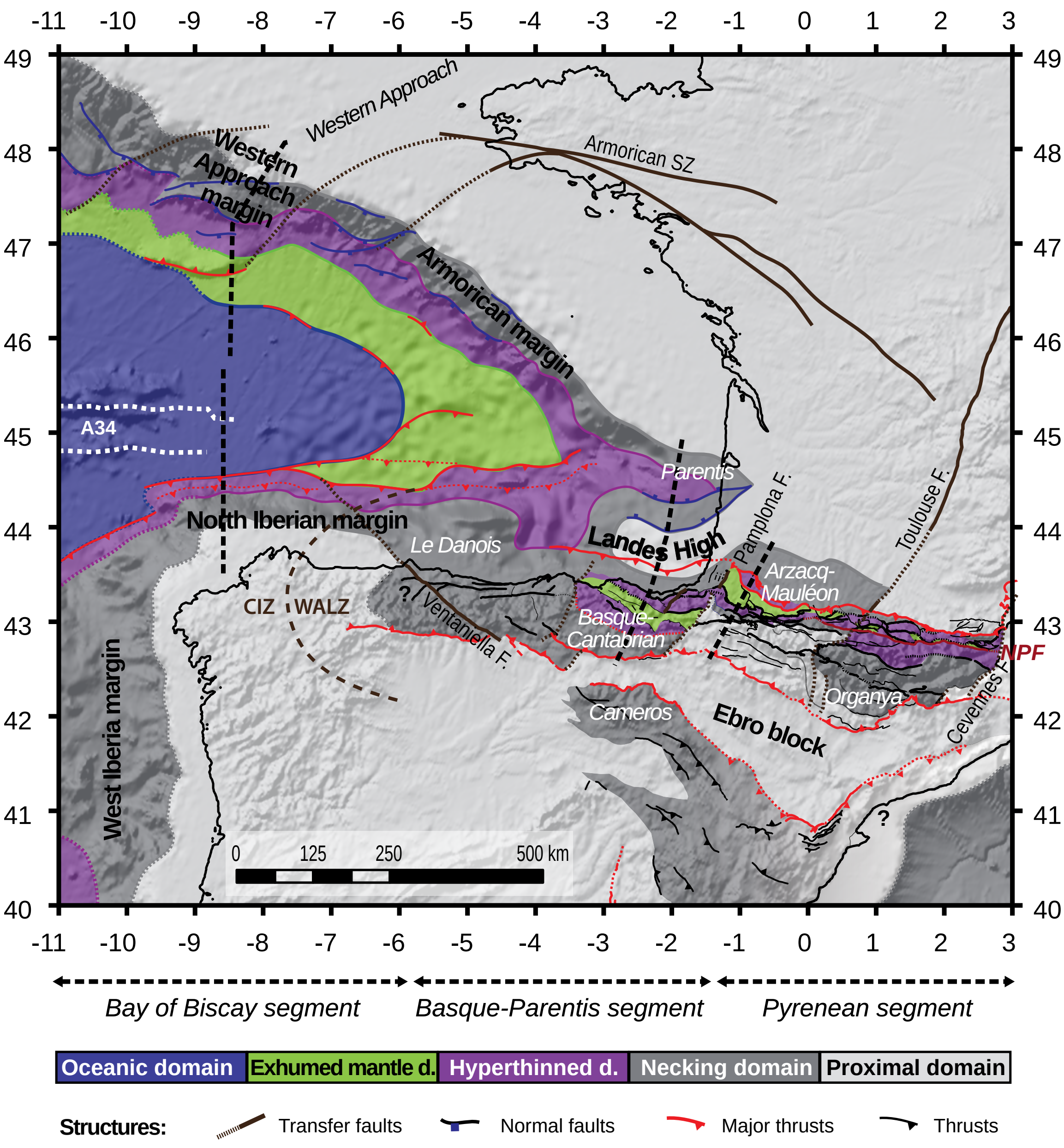


Fig. 1: *Carte des domaines de rift préservés à la transition entre les plaques Européenne et Ibérique.*

1. Comment peut-on caractériser et identifier des domaines de rift en mer et dans leurs analogues fossiles à terre ?

(Chapitre 1 : Approche, chapitre 2 : Cartographie des domaines de rift et Annexes)

Depuis les débuts de la tectonique des plaques, la compréhension des processus d'amincissement lithosphérique et crustal et l'évolution des concepts ont fortement bénéficié des analogies faites entre les données marines et les observations de terrain. Ces informations sont évidemment complémentaires mais bien souvent difficiles à comparer et à relier entre elles. En effet, la plupart du temps, elles proviennent de systèmes extensifs caractérisés par un héritage lithosphérique et crustal différent et une histoire thermique, structurale et sédimentaire distincte. De plus, la comparaison entre ces informations nécessite un aller-retour constant entre des résolutions et des échelles d'observations différentes. En effet, les techniques de sismique réflexion et réfraction dans les marges passives actuelles permettent d'imager l'architecture crustale et stratigraphique. Au contraire, les analogues fossiles sont réactivés et l'architecture de la marge n'est que localement préservée, cependant ils fournissent des informations essentielles sur la nature des roches et les structures qui forment les marges.

Le Golfe de Gascogne et les Pyrénées utilisés comme chantier dans le cadre de cette thèse représentent un laboratoire privilégié qui permet d'avoir accès à un système de rift imagé par sismique réflexion, foré mais également réactivé et exposé à terre. En particulier, le bassin de Parentis a été, à la fois, imagé par sismique réflexion et foré, permettant ainsi de faire le lien entre les données marines et les observations géologiques. Grâce à cette particularité, j'ai pu proposer une nouvelle approche pour identifier et caractériser des domaines de rift des marges actuelles et de leurs analogues fossiles préservés dans les orogènes.

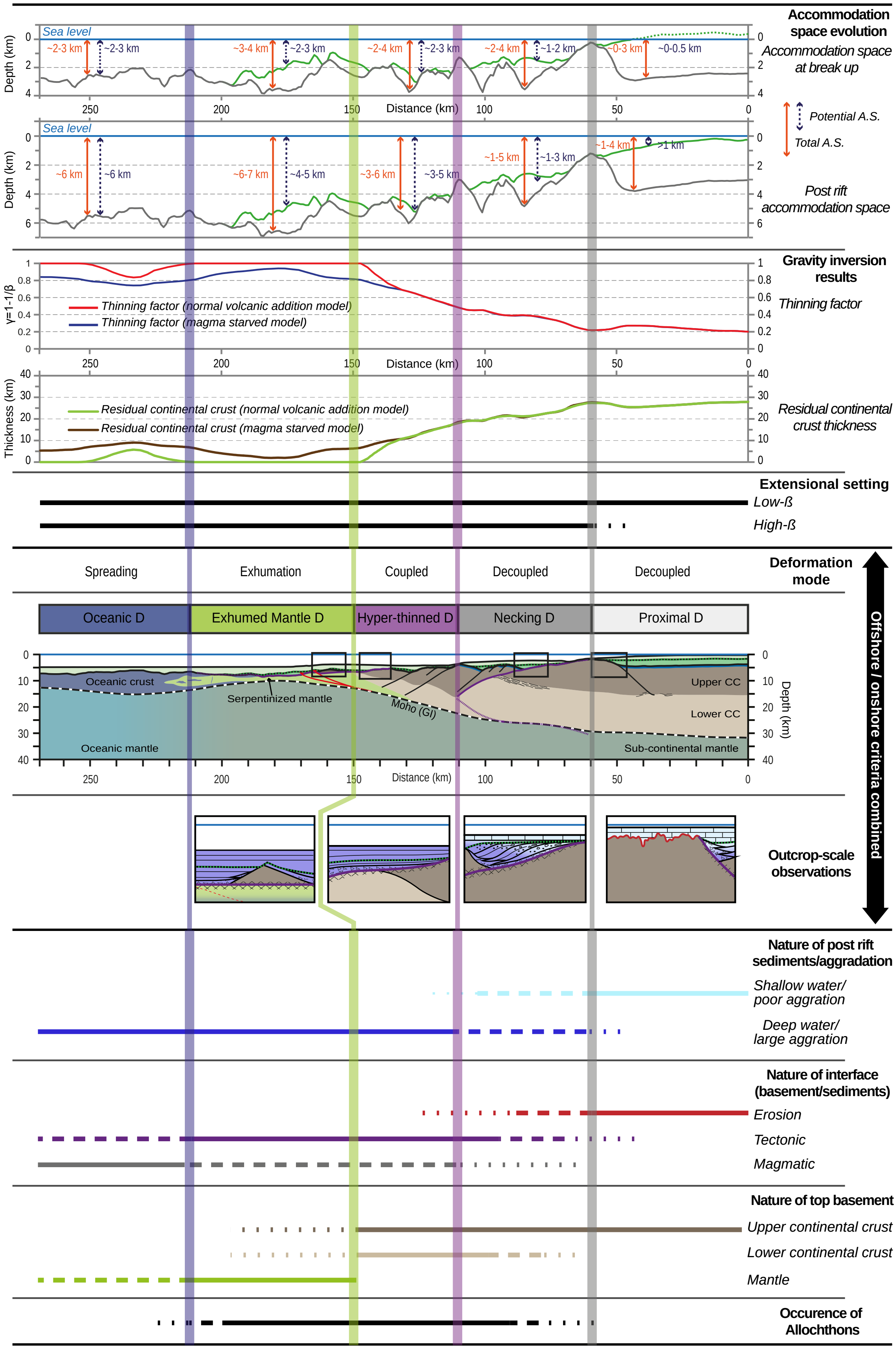
Ainsi, pour les exemples en mer, des interprétations de sismique réflexion ont été associées avec des techniques d'inversion gravimétrique et de " backstripping " flexural afin d'estimer l'espace d'accommodation, l'amincissement crustal et lithosphérique et d'identifier deux types de domaines extensifs ("low and high- β settings" également décris par Wilson *et al.* 2001). Les données de terrain provenant d'affleurement clefs dans le bassin de Mauléon en association avec les forages du bassin de Parentis, ont permis de déterminer la nature des sédiments et du socle ainsi que les structures qui caractérisent les restes fossiles des domaines de marge.

Cette analyse qualitative et quantitative des observations à terre et en mer fait ressortir des éléments de diagnostic essentiels pour caractériser et identifier 5 domaines différents dans les marges passives actuelles et fossiles : les domaines proximaux, de " necking ", hyper-amincis, de manteau exhumé et océaniques (fig 2).

Cette approche, associant géologie et géophysique, peut également servir d'interface entre les observations à terre et en mer et facilite les analogies entre les deux types d'observations. Ainsi, les interprétations sismiques peuvent bénéficier des observations de terrain sur la nature des sédiments, du socle et de leur interface. Inversement, l'approche quantitative proposée sur les exemples en mer permet de replacer les analogues fossiles dans un contexte de marge passive.

Cette approche terre-mer a permis dans un second temps de cartographier les domaines paléogéographiques issus du système de rift lié à l'ouverture du Golfe de Gascogne et partiellement intégré à l'orogène Pyrénéenne (Chapitre 2). Néanmoins, cette approche résumée sur la figure 2 peut également être utilisée dans un cadre plus général pour identifier, caractériser et cartographier des domaines de rift dans des contextes d'amincissement extrême, comme observés dans les marges passives actuelles (Reston 2009 ; Reston & Manatschal 2011 et références cités) et les analogues fossiles préservés dans les chaînes de montagne (e.g. Alpes: Manatschal 2004; Mohn *et al.* 2010, 2012; Masini *et al*. 2011, 2012 ; Calédonides: Andersen *et al.* 2012).

Fig. 2 : *Diagramme synthétique des éléments diagnostiques qui permettent l'identification des domaines de rift à terre et en mer. Dans la partie supérieure du diagramme les observations résultent de méthodes quantitatives, dans la partie inférieure, d'observations de géologie de terrain.*


Accommodation space evolution
Accommodation space at break up
Sea level
Depth (km)
Distance (km)
~2-3 km
~3-4 km
~2-4 km
~1-2 km
~0-3 km
~0-0.5 km
Potential A.S.
Total A.S.
Post rift accommodation space
~6 km
~6-7 km
~4-5 km
~3-6 km
~3-5 km
~1-5 km
~1-3 km
~1-4 km
>1 km
Gravity inversion results
Thinning factor
γ=1-1/β
Thinning factor (normal volcanic addition model)
Thinning factor (magma starved model)
Thickness (km)
Residual continental crust (normal volcanic addition model)
Residual continental crust (magma starved model)
Residual continental crust thickness
Extensional setting
Low-ß
High-ß
Spreading
Exhumation
Coupled
Decoupled
Decoupled
Deformation mode
Oceanic D
Exhumed Mantle D
Hyper-thinned D
Necking D
Proximal D
Oceanic crust
Serpentinized mantle
Moho (GI)
Upper CC
Lower CC
Oceanic mantle
Sub-continental mantle
Offshore / onshore criteria combined
Outcrop-scale observations
Nature of post rift sediments/aggradation
Shallow water/ poor aggration
Deep water/ large aggration
Nature of interface (basement/sediments)
Erosion
Tectonic
Magmatic
Nature of top basement
Upper continental crust
Lower continental crust
Mantle
Occurence of Allochthons

2. Quelle est l'évolution tectonique, spatiale et temporelle des systèmes de rift soumis à un amincissement extrême dans le contexte des marges passives peu magmatique ?

(Chapitre 2: Cartographie des domaines de rift et chapitre 3 : Nature de la limite Ibérie-Europe)

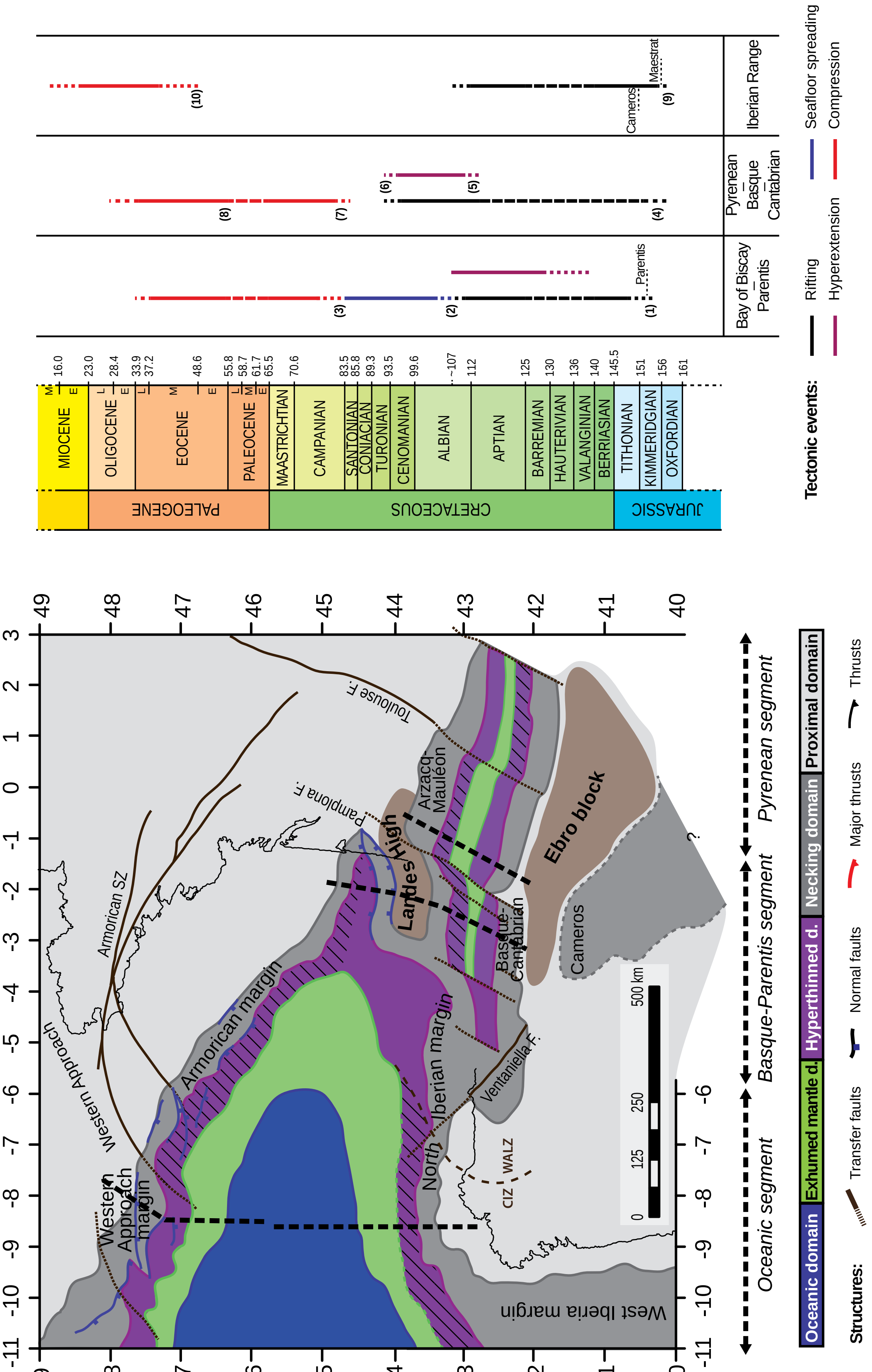
La cartographie des domaines de rift lié à l'ouverture du Golfe de Gascogne a permis de caractériser l'évolution spatiale et temporelle des systèmes de rift soumis aux processus d'hyper-extension (chapitre 2 et 3). La restauration du domaine souligne l'existence de deux systèmes extensifs : les systèmes "Golfe de Gascogne–Parentis" et "Pyrénéen–Basque–Cantabre". Ils sont séparés par le bloc des Landes qui représente un "ruban" de croûte continentale pas ou peu aminci. La structure crustale de ce bloc des Landes est comparable aux "continental ribbons" décrits par Péron-Pinvidic & Manatschal (2010) dans des exemples actuels de l'Atlantique Nord (e.g. Flemish Cap, Rockall Bank, Galicia Bank).

Ces deux systèmes sont préservés à des étapes d'évolution différentes. Dans le Golfe de Gascogne, différentes phases d'amincissement sont enregistrées spatialement depuis sa terminaison Est dans le bassin hyper-aminci de Parentis jusqu'au stade ultime de l'océanisation à l'Ouest, à la jonction avec l'océan Atlantique. Le système de rift "Pyrénéen–Basque–Cantabre" n'a jamais abouti à la formation d'un domaine océanique et correspond à un domaine hyper-aminci avec l'exhumation locale de manteau, comme suggéré pour les bassins Arzacq–Mauléon, Basque–Cantabre ou d'Aulus (e.g. Lagabrielle & Bodinier 2008 ; Jammes *et al.* 2009 ; Lagabrielle *et al.* 2010 ; Roca *et al.* 2011 ; Clerc *et al.* 2012).

Les processus d'hyper-extension et d'amincissement de ces deux systèmes sont diachrones. Dans le système Golfe de "Gascogne–Parentis", l'étude du bassin de Parentis ainsi que l'âge estimé de la première croûte océanique (Montadert *et al.* 1979) montrent que l'amincissement lithosphérique et crustal, initié à la fin du Jurassique, se termine à l'Apto-Albien. Dans le système " Pyrénéen–Basque–Cantabre", la subsidence s'accélère à partir de l'Apto-Albien (e.g. basin Basque–Cantabre: García-Mondéjar *et al.* 1996, 2005; Martin-Chivelet *et al.* 2002; basin Pyrénéen: Debroas 1987, 1990). L'amincissement extrême de la lithosphère et l'exhumation de manteau se poursuit jusqu'au début du Cénomanien, comme indiqué par le remaniement de morceaux de manteau dans des brèches Albiennes à Cénomanienne (bassins de Mauléon et d'Aulus : Jammes *et al.* 2009 ; Debroas *et al.* 2010 ; Lagabrielle *et al.* 2010 ; Clerc *et al.* 2012, 2013)

Une importante segmentation est associée à ces deux systèmes et semble être contrôlée par l'héritage crustal et lithosphérique issu des épisodes varisque à permo-triassique qui précédent l'hyper-extension. La réactivation partielle de certaines de ces structures est associée à des changements majeurs de l'architecture des domaines de rift attribués à des changements de vergence du système d'exhumation, lors de la phase finale du rifting (comme défini par Sutra *et al.* 2013). De telles variations sont notamment observées entre la marge Armoricaine et celle des Entrées de la Manche (précédemment décrite par Thinon 1999) et contrôlées par la zone de cisaillement Armoricaine. Un changement similaire d'architecture est proposé entre les Pyrénées orientales et occidentales et est délimité par la faille de Toulouse. Cette faille est soulignée par une discontinuité lithosphérique majeure, observée grâce aux résultats obtenus par inversion gravimétrique (chapitre 2) et en tomographie sismique (Chevrot *et al.* in prep).

La segmentation du système "Pyrénéen–Basque–Cantabre" est caractérisée par des structures orientée NE–SW (e.g. les failles de Pamplona, de Toulouse, des Cévennes) qui contrôlent l'évolution spatiale des domaines de rift (fig. 3). La restauration du système à la fin de l'hyper-extension telle quelle est proposée dans ce travail, laisse peu de place pour la mise en place d'un épisode transformant syn- à post hyper-extension (albien ou cénomanien), le long de la faille Nord Pyrénéenne ou dans la zone Nord-Pyrénéenne, comme dans les modèles d'évolutions précédemment proposés (Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Choukroune & Mattauer 1978).



Oceanic domain
Exhumed mantle d.
Hyperthinned d.
Necking domain
Proximal domain
Structures:
Transfer faults
Normal faults
Major thrusts
Thrusts
Oceanic segment
Basque-Parentis segment
Pyrenean segment
West Iberia margin
North Iberian margin
Armorican margin
Western Approach margin
Western Approach
Armorican SZ
Toulouse F.
Pamplona F.
Ventaniella F.
Landes High
Arzacq-Mauléon
Basque-Cantabrian
Cameros
Ebro block
WALZ
CIZ
500 km
Tectonic events:
Rifting
Hyperextension
Seafloor spreading
Compression
Bay of Biscay – Parentis
Pyrenean – Basque – Cantabrian
Iberian Range
Parentis
Cameros
Maestrat
JURASSIC
CRETACEOUS
PALEOGENE
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
APTIAN
BARREMIAN
HAUTERIVIAN
VALANGINIAN
BERRIASIAN
TITHONIAN
KIMMERIDGIAN
OXFORDIAN

3. Quel est le rôle de l'hyper-extension lors des processus de réactivation et de formation des orogènes?

(Chapitre 2: Cartographie des domaines de rift et chapitre 3 : Nature de la limite Ibérie-Europe)

La réactivation hétérogène des systèmes de rift permet d'accéder à différentes étapes de la déformation compressive. A l'Ouest, les étapes initiales de la déformation sont préservées, tandis que, les bassins Basque–Cantabre, d'Arzacq–Mauléon ou encore d'Aulus sont intégrés à l'orogène Pyrénéenne et permettent de comprendre comment les structures et domaines de rift sont réactivés.

L'intégration d'observations à travers l'ensemble du système à terre et en mer est au cœur de travail de thèse et a permis de proposer un modèle conceptuel pour expliquer le rôle de l'héritage du rift dans les processus de réactivation et de formation des orogènes, telles que les Pyrénées. La déformation compressive s'initie dans les domaines de manteau exhumé. La serpentinisation du manteau peut, en effet, représenter une importante zone de faiblesse, comme précédemment décrit pour la marge Ibérique (Péron-Pinvidic *et al.* 2008) et discuté par Lundin & Doré (2011). Ainsi, l'ancienne faille de détachement, coiffant le manteau exhumé et serpentinisé, peut servir de surface de décollement et permettre la propagation de la déformation compressive au domaine de croûte hyper-amincie. Le comportement cassant et la forte altération de ce domaine (Pérez-Gussinyé & Reston 2001; Reston & Pérez-Gussinyé 2007) peut permettre la formation d'un prisme d'accrétion. Dans les anciens bassins hyper-amincis (ex : Arzacq–Mauléon, fig. 4), la taille réduite du système va permettre, dans un stade ultérieur, d'initier une proto-subduction de l'ancien domaine de croûte hyper-amincie, lors de l'arrivée de la marge conjuguée. L'implication des anciens domaines conjugués de " necking ", dans le processus de collision, est soulignée par un changement du mode de déformation. La subduction des anciens domaines de manteau exhumé et de croûte hyper-amincie est essentiellement contrôlée par des processus asymétriques, tandis que, la collision continentale est associée à une déformation relativement symétrique, impliquant des pro-charriages et retro-charriages, comme observé dans le cas du bassin de Mauléon (fig. 4). L'architecture finale de la chaine est contrôlée par les domaines de " necking " qui vont jouer le rôle de deux "buttoirs" lors de la réactivation des failles de détachement conjuguées, interprétées comme les structures majeures des domaines de " necking " dans les marges actuelles (Mohn *et al.* 2012; Sutra *et al.* 2013).

Fig. 3 : *Restauration des systèmes de rift qui forment la limite Ibérie-Europe avant l'initiation de la convergence (avant le Santonien) et évolution des déformations extensives et compressives dans chacun des systèmes de rift.*

Ainsi, les nouvelles interprétations présentées dans ce travail suggèrent que les domaines "subductés", imagés par réfraction et tomographie sismique (e.g. Daignières *et al.* 1994; Roure & Choukroune 1998; Pedreira *et al.* 2003, 2007), correspondent en partie aux anciens domaines de manteau exhumé et de croûte hyper-amincie. Ainsi, l'essentiel de la déformation compressive est accommodée par les domaines de manteau exhumé et de croûte hyper-amincie et les structures de rift associées. Les anciennes limites de domaines représentent donc les zones majeures de réactivation.

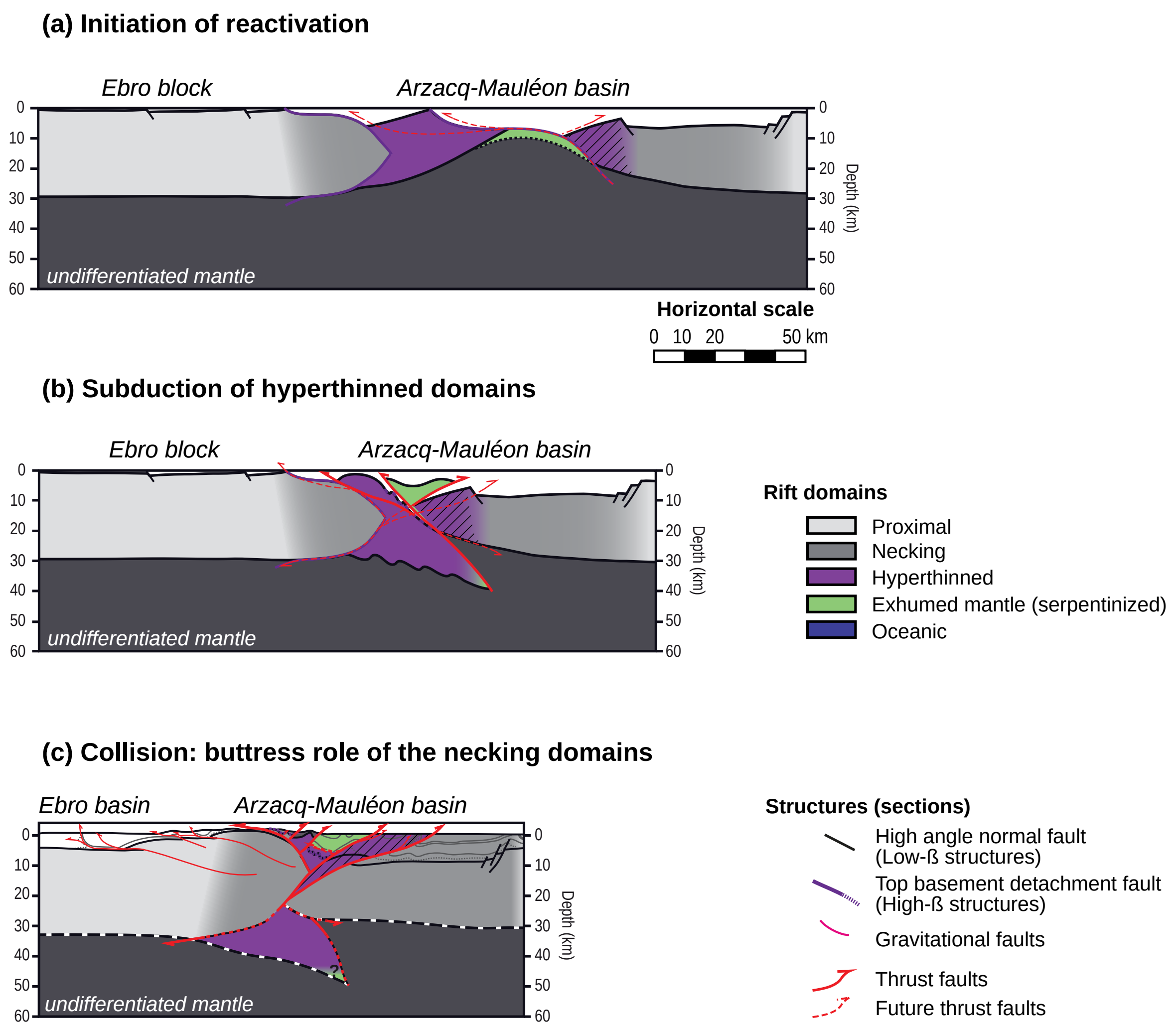


Fig. 4 : *Modèle conceptuel illustrant la réactivation progressive du bassin d'Arzacq-Mauléon. (a) Initiation de la réactivation dans le domaine de manteau exhumé et serpentinisé. (b) Subduction des domaines hyper-amincis. (c) Collision : les anciens domaines proximaux et de necking jouent le rôle de buttoir.*

4. Quelle est l’origine de l’évolution de l’architecture à travers le système ?

(Chapitre 2: Cartographie des domaines de rift & Chapitre 3 : Nature de la limite Ibérie-Europe)

Afin de comprendre l’origine des changements d’architecture qui sont observés d’Ouest en Est (e.g. Muñoz 2002), il a d’abord été nécessaire de déchiffrer les possibles évolutions spatiales de l’architecture des systèmes de rift. Les résultats issus de cette thèse suggèrent que la réactivation hétérogène observée est directement liée à une architecture pré-compressive complexe.

D’une part, la restauration du système avant l’initiation de la convergence souligne l’existence de deux systèmes de rift préservés, à des stades d’évolution différents et spatialement distincts. Cette architecture complexe va aboutir à une compétition entre les deux systèmes lors de la convergence. Cette situation est très bien illustrée par l’exemple des bassins de Parentis et Basque–Cantabre. En effet au Nord, le bassin de Parentis est relativement bien préservé, tandis qu’au Sud, le bassin Basque–Cantabre est complètement inversé et intégré à la terminaison Ouest des Pyrénées (fig. 5).

D’autre part, à l’échelle d’un même système de rift, des variations de l’architecture orogénique sont également observées, c’est notamment le cas entre l’Est et l’Ouest des Pyrénées (e.g. Muñoz 2002). La restauration du système de rift, précédemment discutée, permet de proposer certains éléments de réponse. En effet, un changement de vergence du système d’exhumation est proposé, de part et d’autre de la faille de Toulouse (précédemment suggéré par Jammes *et al.* 2009). En effet, dans les Pyrénées occidentales, les systèmes d’exhumation sont à pendage Nord (Jammes *et al.* 2009 ; Lagabrielle *et al.* 2010; Masini 2011) tandis qu’à l’Est, la situation est inversée, comme démontré Vauchez *et al.* (2013) dans le cas de l’exhumation du massif de l’Agly. De tels changements correspondent à des variations de l’architecture des domaines de rift, comme observé en mer, entre les marges Armoricaine et des Entrées de la Manche représentant respectivement une architecture de type “upper plate” et “lower plate” (voir définition par Sutra *et al.* 2013). Ainsi ces variations d’architecture du rift pourraient en partie expliquer la structuration différente observée entre l’Est et l’Ouest des Pyrénées.

Les résultats de ce travail montrent qu’il est nécessaire de prendre en compte l’architecture 3D des anciens domaines de rift hyper-amincis dans les modèles de formation des orogènes de collision comme les Pyrénées.

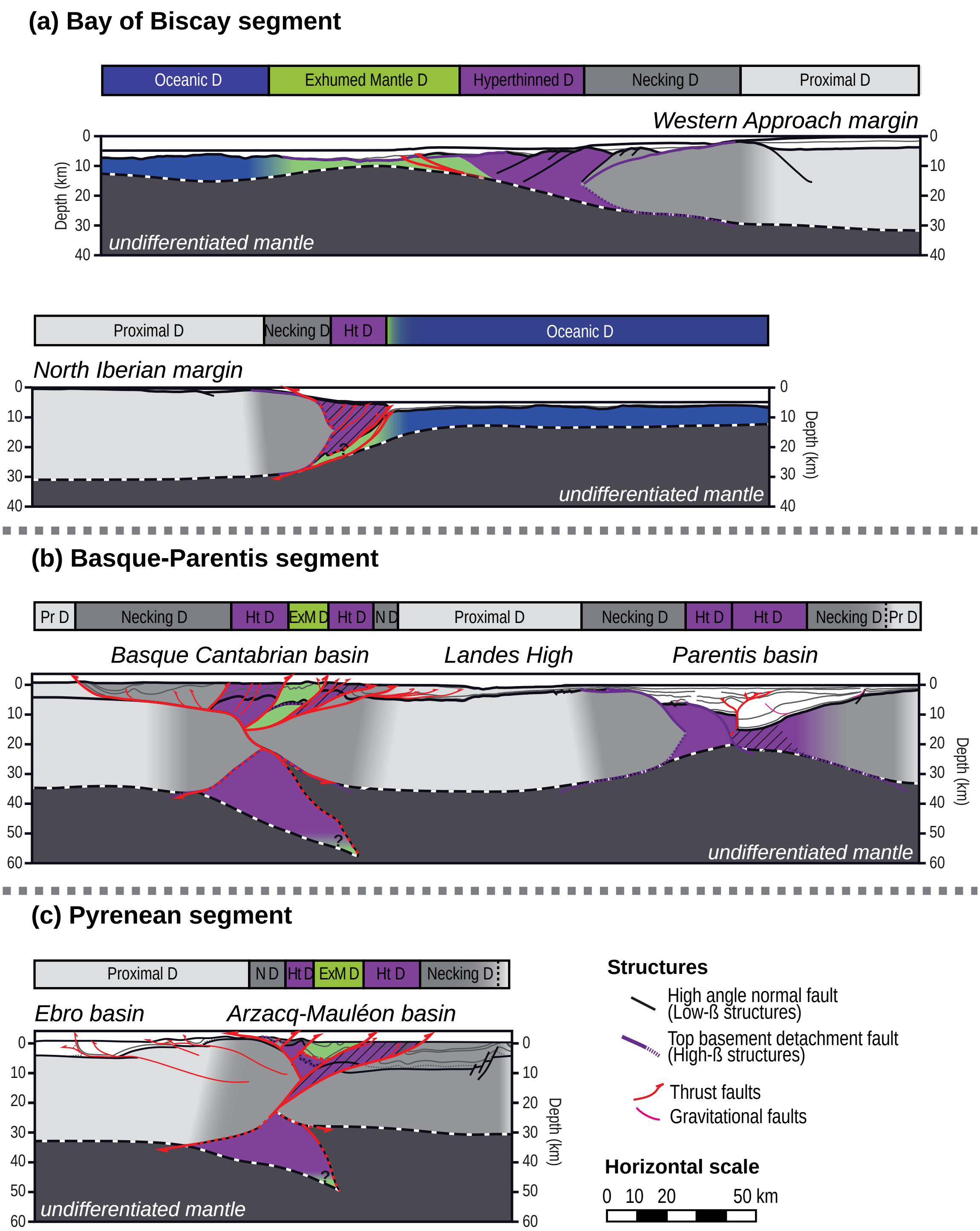


Fig. 5 : *Evolution de l'architecture du domaine d'Ouest en Est. L'architecture héritée du rift joue un rôle majeur. (a) Segment du Golfe de Gascogne. (b) Segment Basque-Parentis. (c) Segment Pyrénéen (l'exemple du bassin Arzacq-Mauléon).*

5. Quelle est la nature de la limite de plaque entre Ibérie et Europe ?

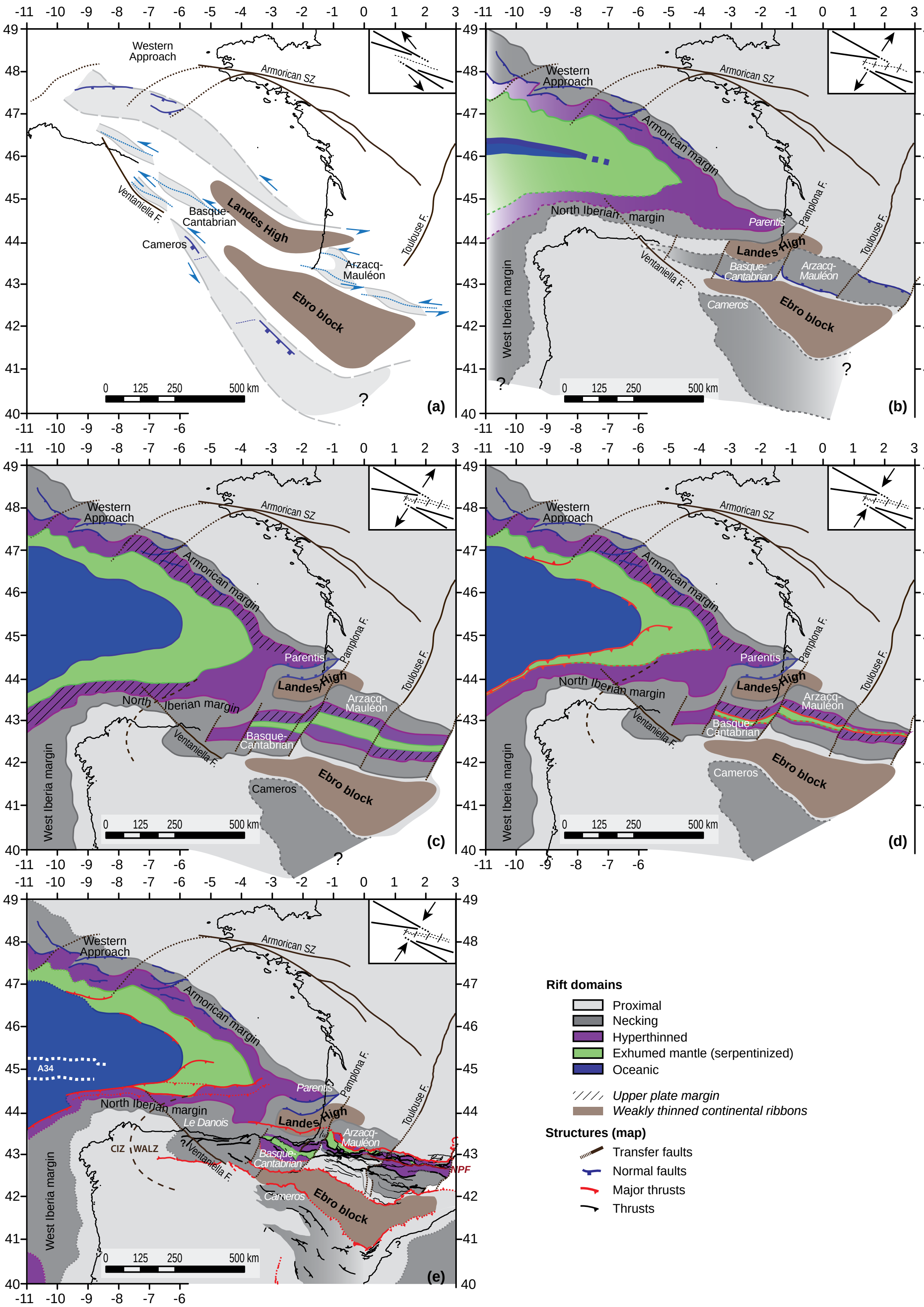
(Chapitre 3 : Nature de la limite Ibérie-Europe)

La cartographie des systèmes de rift préservés à la transition entre Europe et Ibérie met en avant une architecture 3D complexe et une évolution polyphasée, comme illustré par le diachronisme des processus d'amincissement extrême entre les systèmes "Golfe de Gascogne–Parentis" et "Pyrénéen–Basque–Cantabre". Ainsi dans une dernière partie, je me suis intéressée au contexte de formation de la limite de plaque entre l'Ibérie et l'Europe. Les reconstructions paléogéographiques régionales nécessitent un mouvement senestre de l'Ibérie dont le timing, le déplacement et les implications, pour le domaine Pyrénéen, sont au centre de nombreux débats. En effet, selon les restaurations proposées et basées sur des interprétations controversées des anomalies magnétiques (voir Bronner *et al.* 2011, 2012 ; Tucholke & Sibuet 2012), le domaine Pyrénéen est soit associé à une zone de subduction (e.g. Sibuet *et al.* 2004; Vissers & Meijers 2012) ou soit interprété comme un domaine transtensif (e.g. Le Pichon *et al.* 1971; Mattauer & Séguret 1971; Choukroune & Mattauer 1978).

La synthèse des précédentes observations associées aux résultats de ce travail a permis de proposer un modèle d'évolution de la déformation à l'échelle de la limite de plaque entre l'Ibérie et l'Europe. Ainsi, la mise en mouvement de la plaque Ibérique à partir du Jurassique supérieur est accommodée par une large zone de déformation, notamment partitionnée entre deux systèmes de rift d'orientation NW–SE qui se propagent dans des directions opposées : les systèmes "Golfe de Gascogne–Parentis" et "Intra-Ibérique". Ces deux systèmes présentent une segmentation relativement similaire et caractérisée par des structures de rift et des failles de transfert orientées NW–SE à NE–SW. Entre ces deux rifts, séparés par les blocs des Landes et d'Ebro, la déformation est diffuse entre plusieurs bassins individualisés d'orientation E–W (ex : Arzacq, Mirande). Le système de rift "Pyrénéen–Basque–Cantabre" est activé plus tardivement, lors de l'initiation des processus d'accrétion océanique dans la partie Ouest du Golfe de Gascogne. L'extension NE–SW de ce système est soulignée par sa segmentation. La déformation diffuse observée à terre est interprétée comme une tentative avortée de propagation et de localisation d'une limite de plaque divergente entre l'Ibérie et l'Europe. Cette interprétation a des conséquences majeures pour le domaine Pyrénéen qui n'est plus interprété comme un domaine de bassin en pull-apart à partir de l'Apto-Albien, comme précédemment proposé par Jammes *et al.* 2009 ou Lagabrielle *et al.* 2010 par exemple.

De plus, l'évolution polyphasée décrite souligne l'importance de la déformation pré-rupture continentale. Même si ces déplacements restent difficiles à quantifier, leur intégration dans les modèles cinématiques représente un enjeu majeur pour décrire l'évolution spatio-temporelle de l'ouverture de l'Atlantique Nord. Ainsi les restaurations basées uniquement sur la restauration des anomalies magnétiques (e.g. Sibuet *et al.* 2007 ; Vissers & Meijers 2012) peuvent conduire à des interprétations incompatibles avec les observations géologiques, comme suggéré par Norton *et al.* (2007) pour le Golfe de Gascogne.

Fig. 6 : *Evolution de la déformation à la transition entre les plaques Européenne et Ibérique. (a) Rifting oblique (Jurassique Supérieur à Aptien Supérieur) (b) Rupture lithosphérique et initiation du processus d'accrétion océanique dans le Golfe de Gascogne (Apto-Albien). (c) Tentative avortée de propagation et de localisation de la limite de plaque (avant le Santonien). (d) Initiation de la subduction des domaines hyper-amincis (Crétacé supérieur). (e) Collision continentale (Eocène-Oligocène) et carte des domaines de rift actuelle.*



(a)
(b)
(c)
(d)
(e)
Western Approach
Armorican SZ
Armorican margin
North Iberian margin
West Iberia margin
Landes High
Ebro block
Basque-Cantabrian
Arzacq-Mauléon
Cameros
Parentis
Ventaniella F.
Pamplona F.
Toulouse F.
Le Danois
CIZ
WALZ
A34
NPF
0 125 250 500 km
Rift domains
Proximal
Necking
Hyperthinned
Exhumed mantle (serpentinized)
Oceanic
Upper plate margin
Weakly thinned continental ribbons
Structures (map)
Transfer faults
Normal faults
Major thrusts
Thrusts

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CONCLUSIONS

L'objectif de la thèse était d'une part de caractériser l'évolution spatiale et temporelle des mécanismes de l'amincissement extrême de la croûte, et d'autre part de comprendre le rôle de ces domaines hyper-amincis lors des processus de réactivation et de formation des orogènes de collision. Pour répondre à ces thématiques, l'étude a été focalisée sur le Golfe de Gascogne et les Pyrénées qui permettent d'avoir accès à des données de géophysique marine et de géologie de terrain sur un même système de rift hyper-aminci.

Les principaux résultats de ce travail sont :

- Le développement et l'application d'une approche Terre-Mer qui facilite l'identification, la caractérisation et l'interprétation des domaines de rift dans les systèmes actuels et leurs analogues fossiles exposés dans les orogènes. Cette approche permet de faire le lien entre des données géophysiques et des observations de géologie de terrain d'échelles et critères complémentaires différents.
- La cartographie des domaines de rift à terre et en mer révèle une architecture complexe caractérisée par différents systèmes de rift préservés à des stades d'évolution différents et dont l'évolution est diachrone.
- Cette évolution spatiale complexe des systèmes de rift est caractérisée par une importante segmentation en partie contrôlée par la structuration pré-rift. Cette segmentation contrôle des variations d'architecture du rift mais pas celle du domaine océanique. De plus, elle remet en question l'interprétation classique du rôle de la faille Nord Pyrénéenne.
- L'architecture 3D complexe permet en partie d'expliquer la réactivation hétérogène observée à travers l'ensemble du domaine.
- Plusieurs étapes de la déformation compressive ont pu être distinguées et mises en relation avec l'architecture initiale du rift. La réactivation est initiée dans le domaine de manteau exhumé et serpentinisé. La fermeture progressive de ce domaine permet de propager la déformation au domaine de croûte hyper-amincie qui va pouvoir former un prisme d'accrétion. L'arrivée de la marge conjuguée permet d'amorcer une subduction des anciens domaines de croûte hyper-amincie. L'architecture finale de l'orogène résulte de la collision des anciens domaines de " necking " qui vont représenter deux buttoirs.
- La transition entre l'Ibérie et l'Europe est caractérisée par une architecture complexe et une évolution fortement polyphasée. Un épisode de rifting oblique essentiellement partitionné entre deux systèmes de rift permet d'accommoder le déplacement latéral de l'Ibérie. L'initiation des processus d'accrétion dans la partie Est du Golfe de Gascogne est associée à une déformation extensive diffuse, observée dans les bassins Pyrénéens et Basque–Cantabre. Cet épisode est interprété comme une tentative avortée de localisation de limite de plaque.

- Cette étude a montré l'étroite interaction entre l'héritage pré-rift et l'évolution spatiale des systèmes de rift et de marges passives actuelles. De plus, ce travail souligne l'importance de l'architecture issue du rift pour comprendre la réactivation et l'imbrication des marges passives dans les orogènes.

Les interprétations proposées dans cette thèse soulignent l'importance des mouvements pré-rupture lithosphérique pour les reconstructions cinématiques. Une estimation plus précise de ces mouvements serait essentielle pour comprendre et caractériser la cinématique de l'Atlantique Nord.

Ainsi la localisation de la limite de plaque entre l'Europe et l'Ibérie représente une importante jonction entre les systèmes de rift Atlantique et Téthysien. L'évolution polyphasée décrite dans ce manuscrit nécessiterait d'être intégrée à une plus grande échelle pour comprendre la propagation de l'Atlantique Nord, mais également l'évolution du système Alpin.

Enfin l'architecture complexe décrite, résultant d'un partitionnement complexe des déformations extensives et la tentative de localisation avortée de la limite de plaque entre l'Ibérie et l'Europe, nécessiterait d'être comparée à des analogues actuels comme les bassins de Woodlark et de «South China Sea».

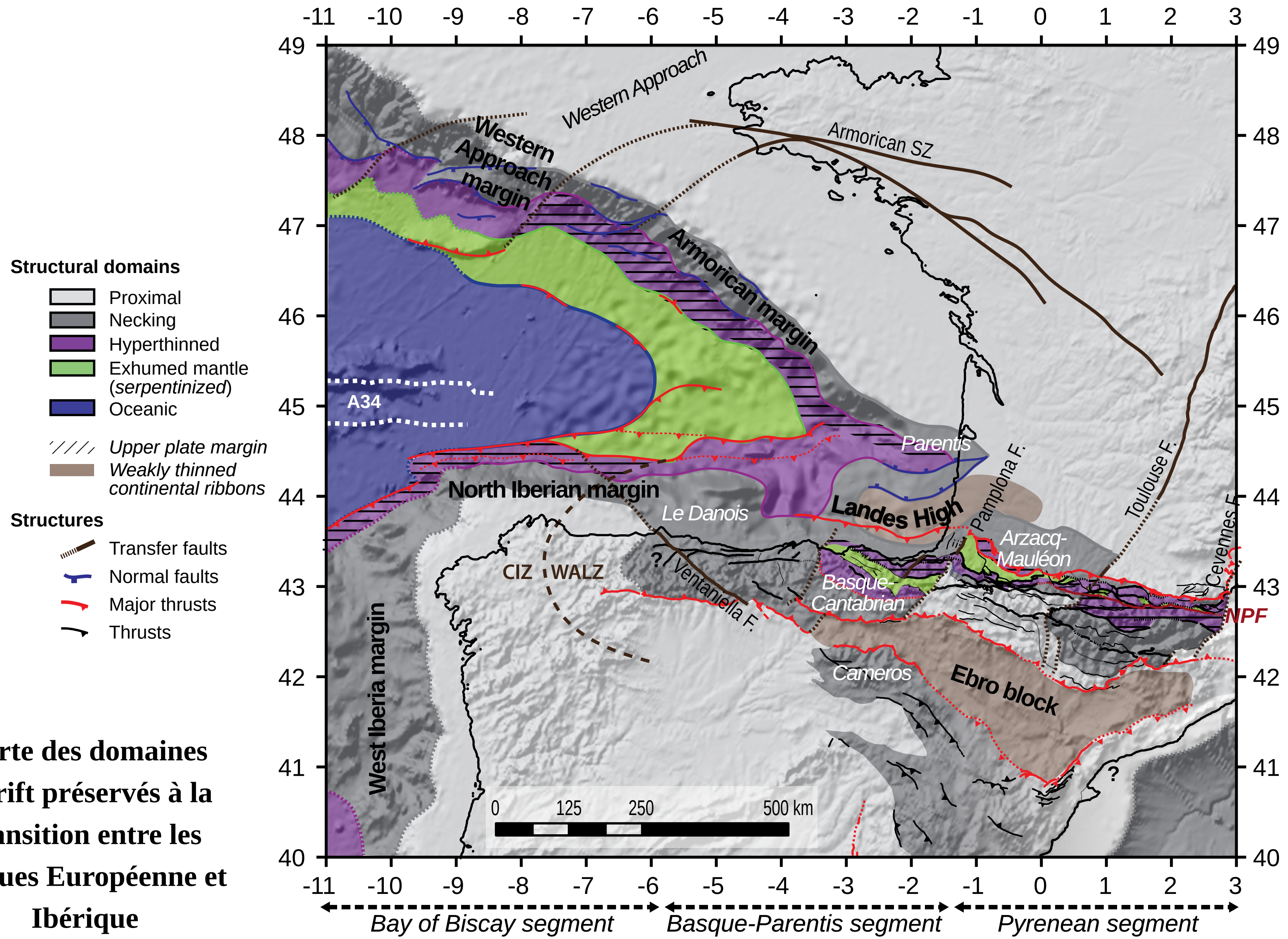
ANNEXES

ANNEXE 1: LES CARTES GÉOLOGIQUES

La carte des domaines de rift présentée en annexe est l'un des résultats majeurs de ce travail de thèse. L'approche développée dans le chapitre 1 et résumée dans le chapitre 2 a été appliquée à l'ensemble du système Golfe de Gascogne et Pyrénées.

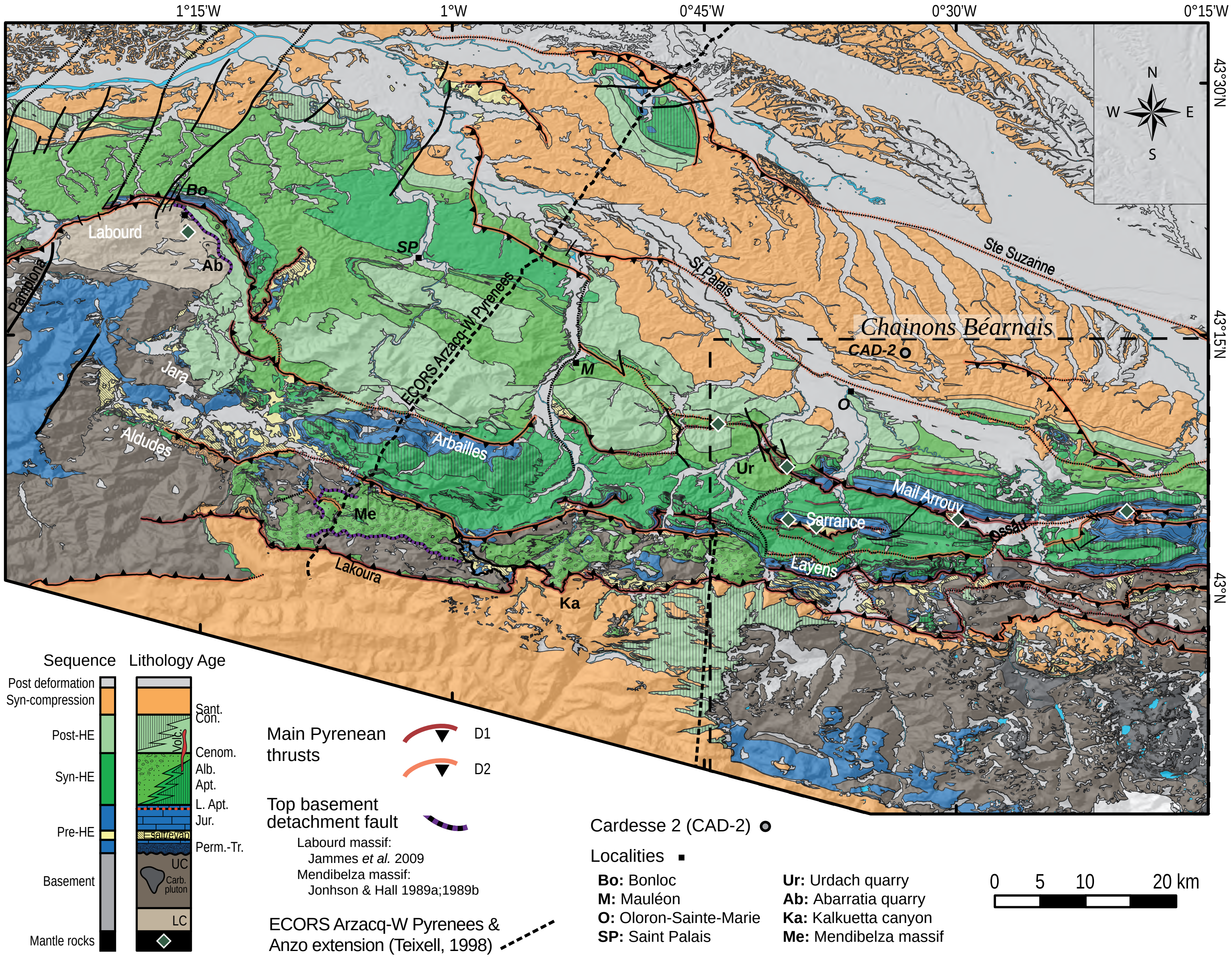
Les cartes géologiques du bassin de Mauléon et de la région des chainons Béarnais ont été réalisée à partir d'extraits de la carte géologique de France à 1/50 000 édités et vectorisées par le BRGM. Les cartes suivantes ont été utilisée pour ce travail : Arthez de Béarn, Argelès-Gazost, Gavarnie, Hasparren, Iholdy, Laruns-Somport, Larrau, Lourdes, Oloron-Sainte-Marie, Orthez, Mauléon-Licharre, Morlaàs, Pau, Saint-Jean-Pied-de-Port, Tardets

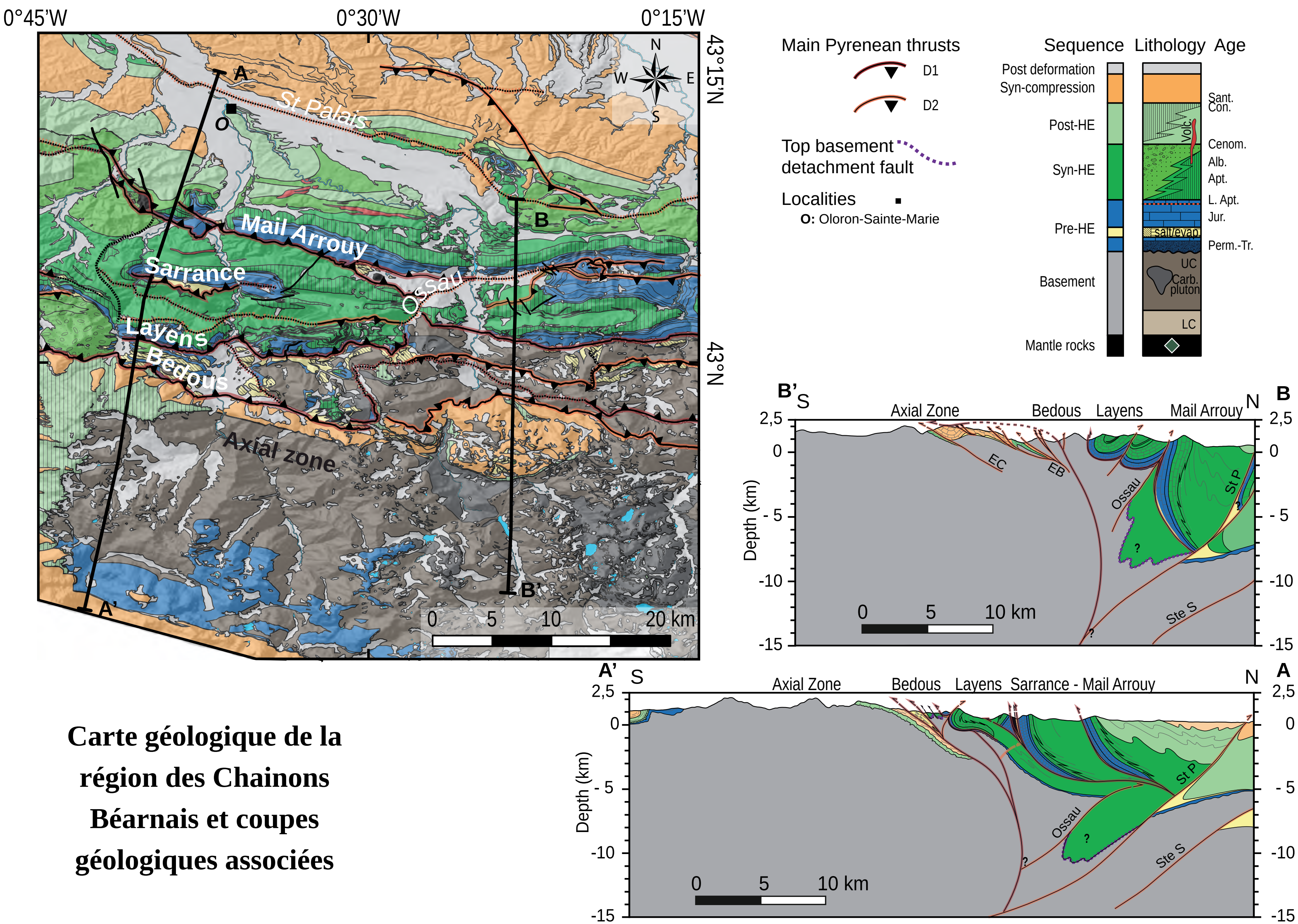
Le code couleur initial des formations géologiques de ces cartes a été unifié et simplifié. Les structures ont également été simplifiées et partiellement modifiées en fonction des nouvelles observations apportées dans le cadre de ce travail.



Carte des domaines de rift préservés à la transition entre les plaques Européenne et Ibérique

Carte géologique du bassin de Mauléon

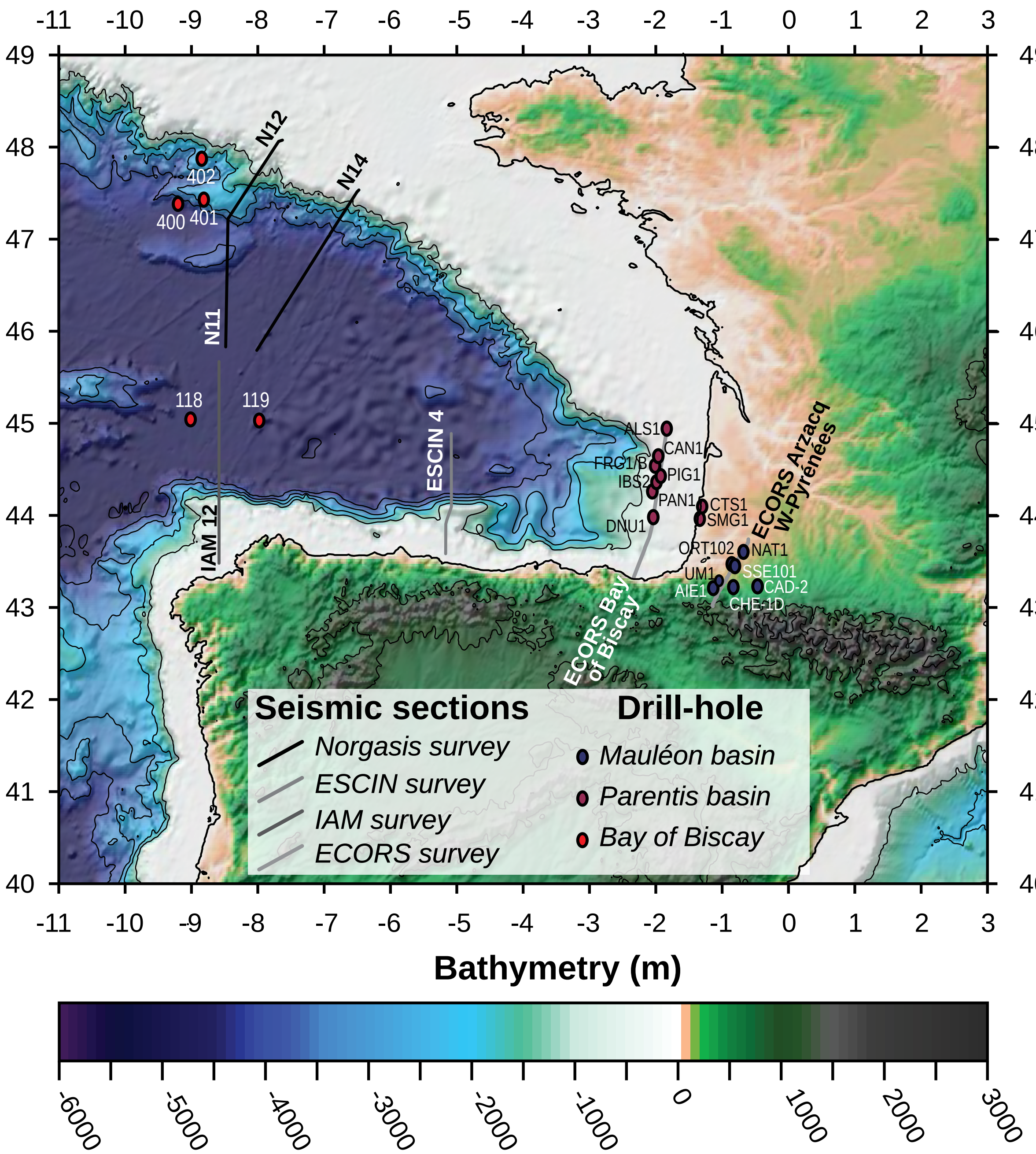




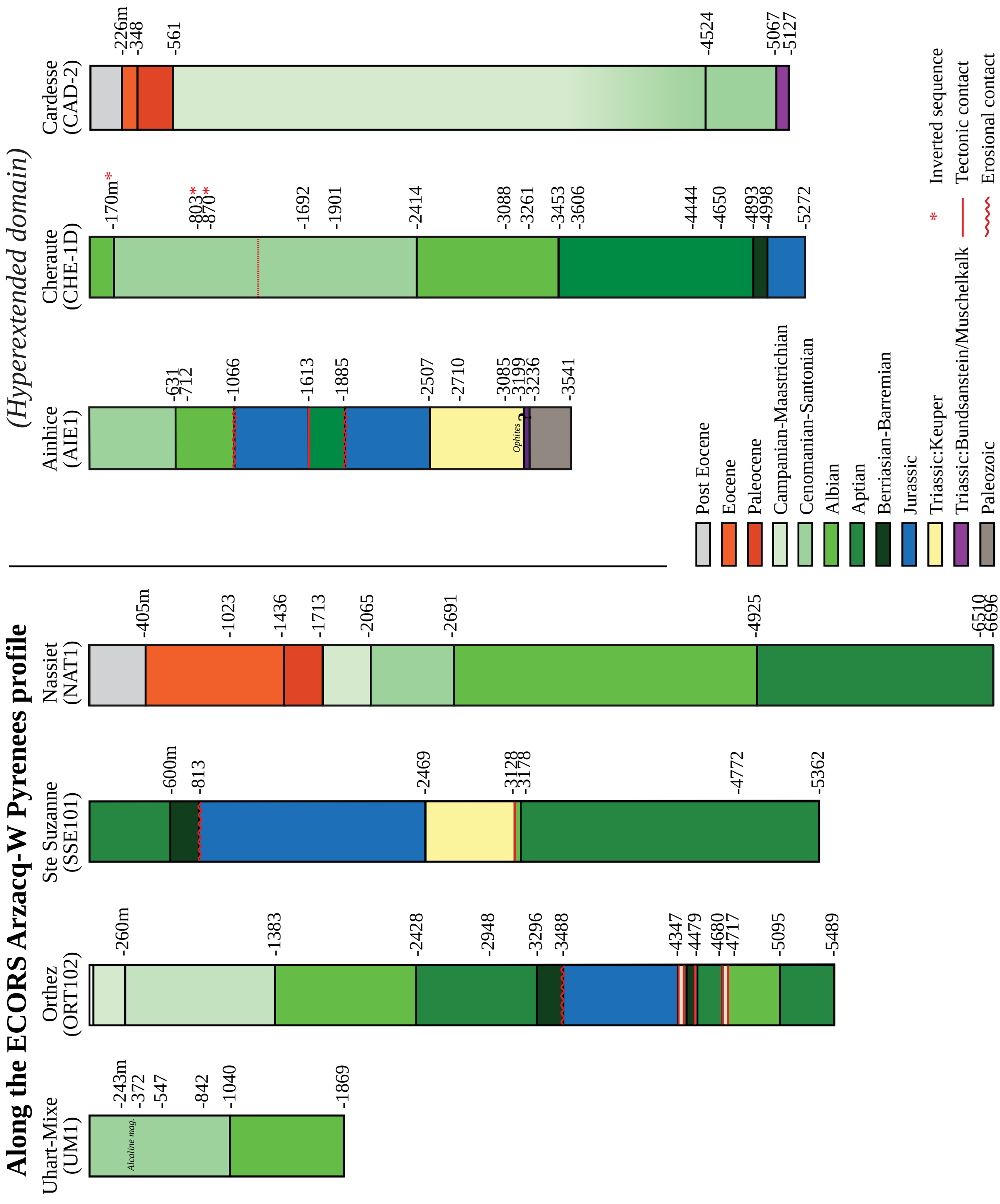
Carte géologique de la région des Chainons Béarnais et coupes géologiques associées

ANNEXE 2: LES DONNÉES DE FORAGES

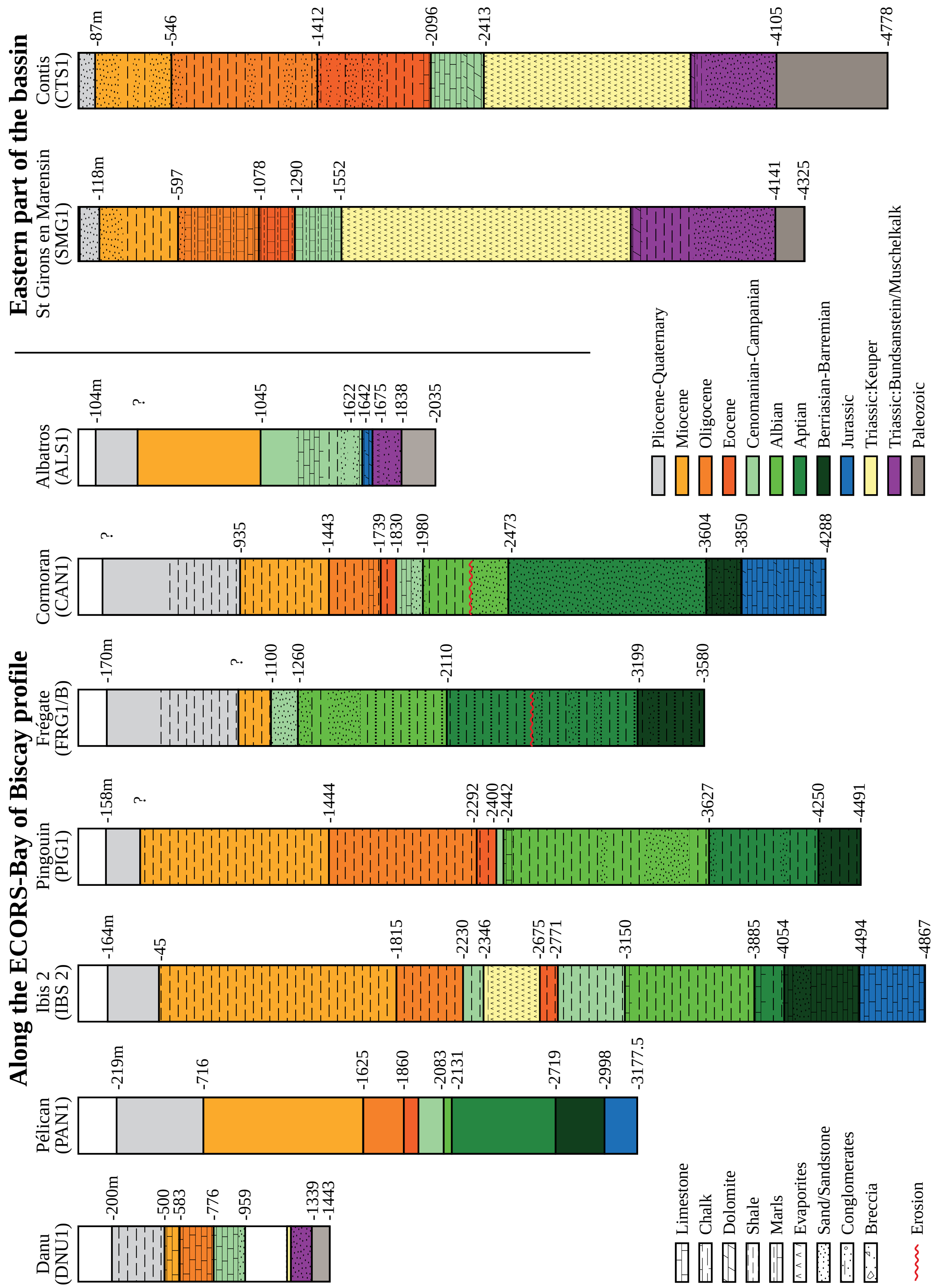
Les données de forages du Golfe de Gascogne proviennent des rapports DSDP (Deep Sea Drilling project) 48 (sites 400, 401, 402) et 12 (sites 118 et 119). Les données de forages du bassin de Parentis et du bassin de Mauléon sont issues des synthèses des données sur le bassin d'Aquitaine publiées par le BRGM (Bureau des Recherches Géologiques et Minières) : BRGM, 1974 ; Serrano et al., 2006. Des observations supplémentaires proviennent du Mémoire de la Société Géologique de France (171), Bois et al., 1997.



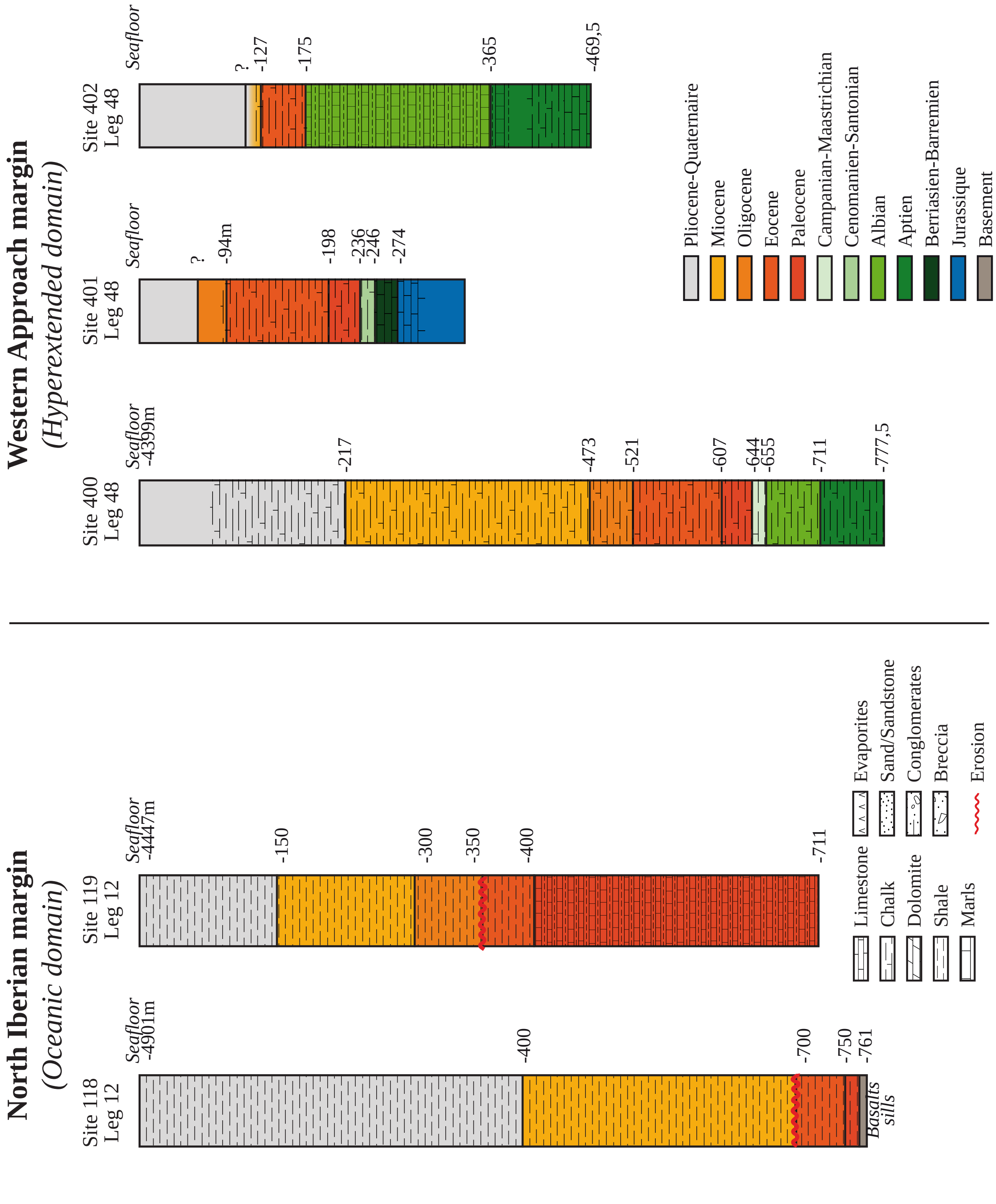
DRILL HOLE DATA IN THE ARZACQ-MAULÉON BASIN



Drill hole data in the Parentis basin



Drill hole data in the Bay of Biscay

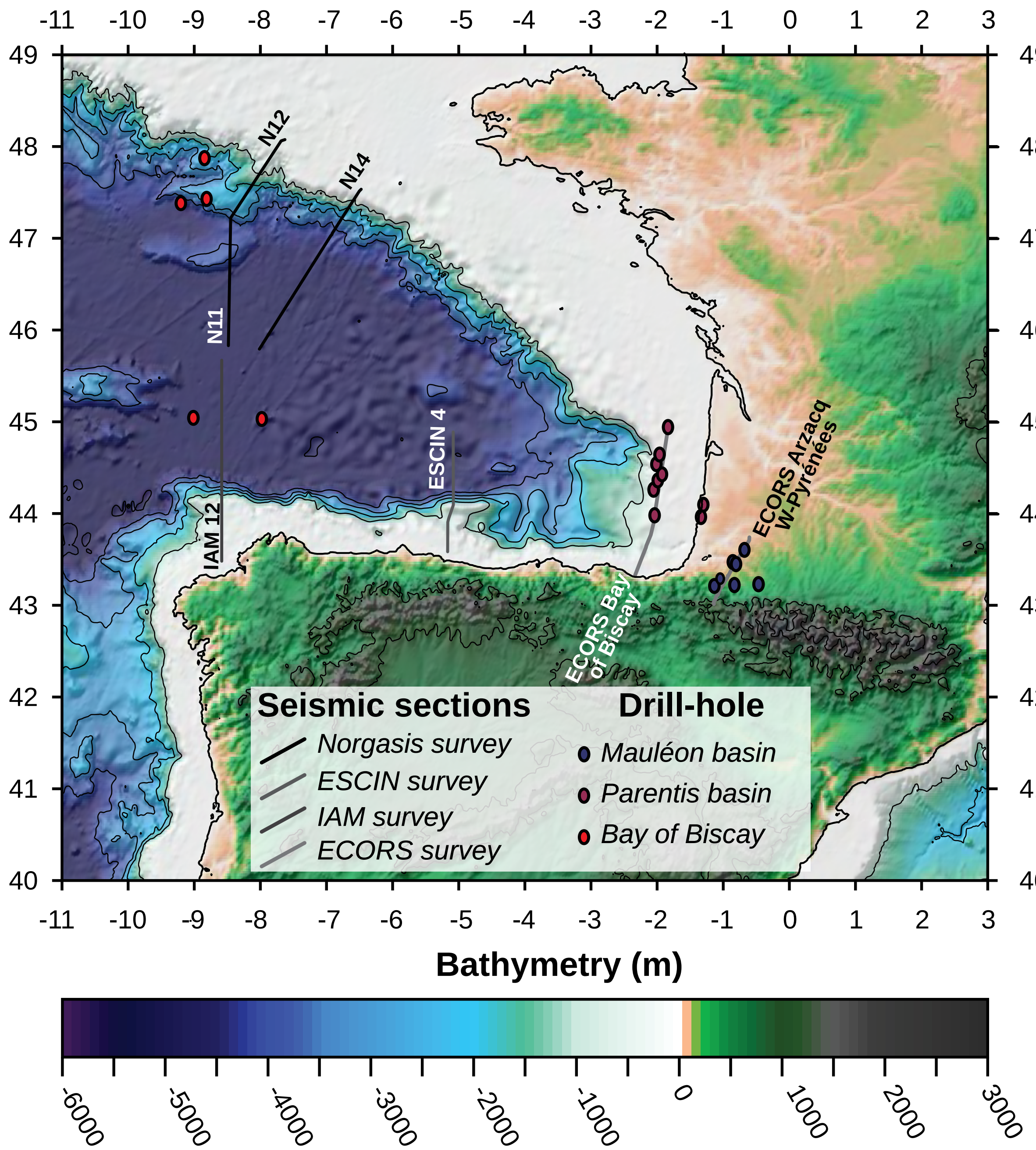


Annexe 3: Les données sismiques

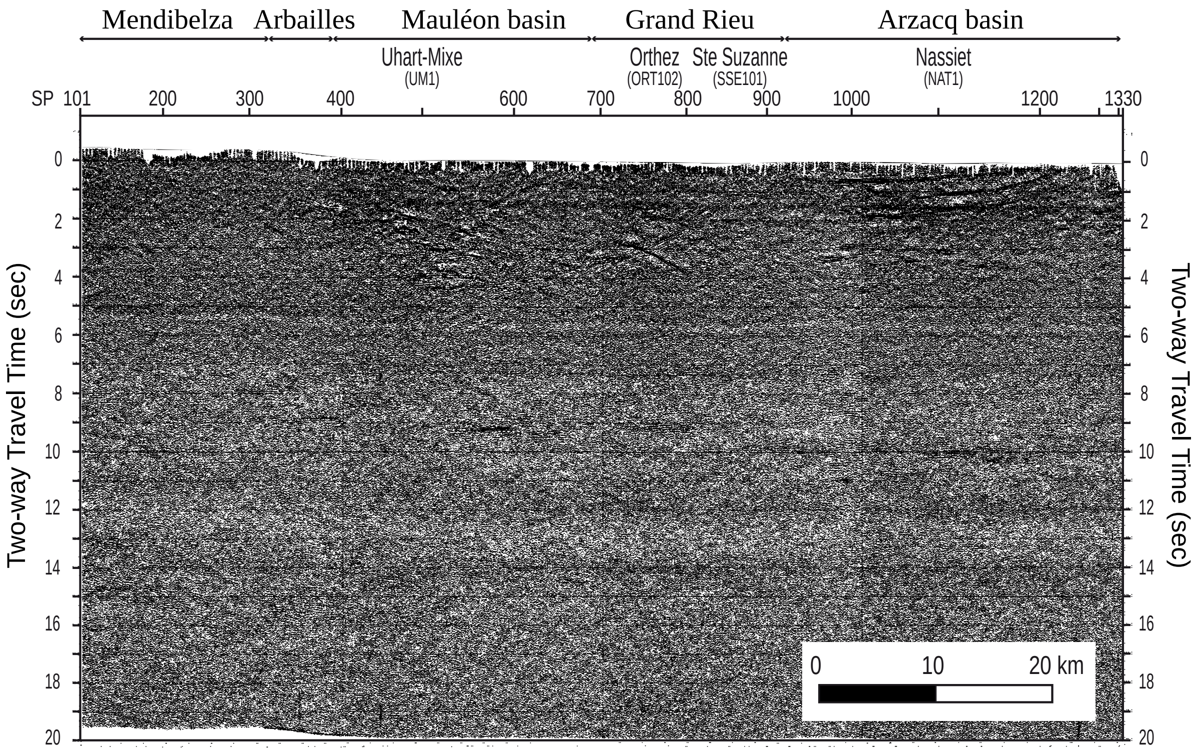
Les lignes sismiques de la campagne ECORS-Golfe de Gascogne et Arzacq-W Pyrénées ont été transmises par Marc Schaming (EOST, Strasbourg).

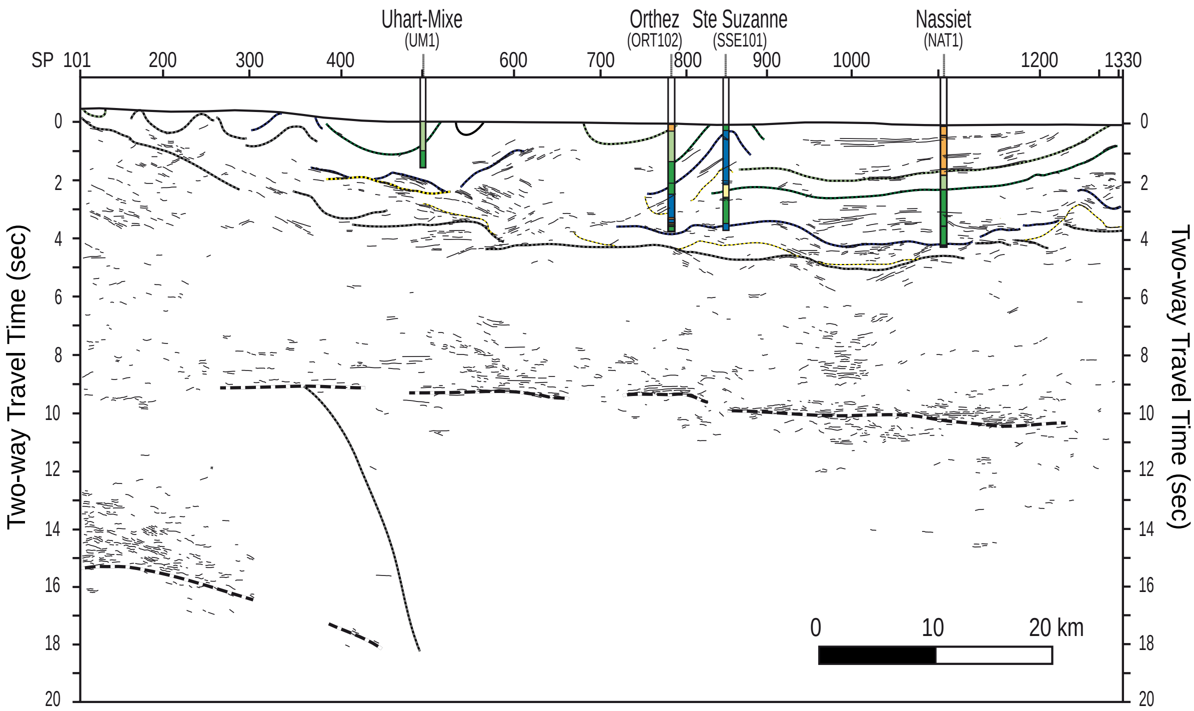
Les lignes sismiques de la champagne Norgasis (Avedik et al. 1993, 1996; Thinon 1999) ont été transmises par Isabelle Thinon.

Les lignes sismiques de la marge Nord Ibérie (ESCIN 4 and IAM 12) ont été transmises par l'Institut de Ciències de la Terra J. Almera. http://www.ictja.csic.es/edt/.

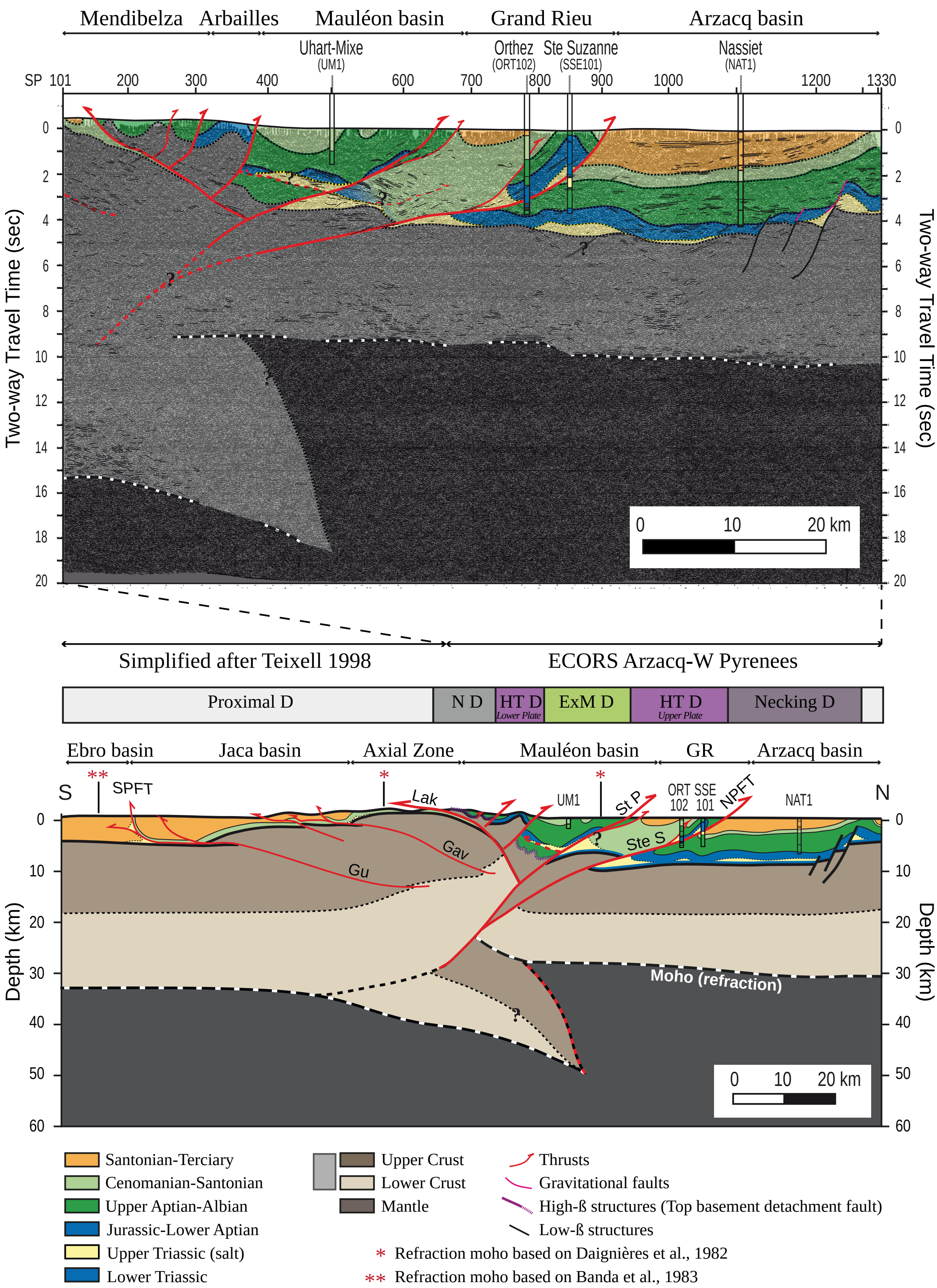


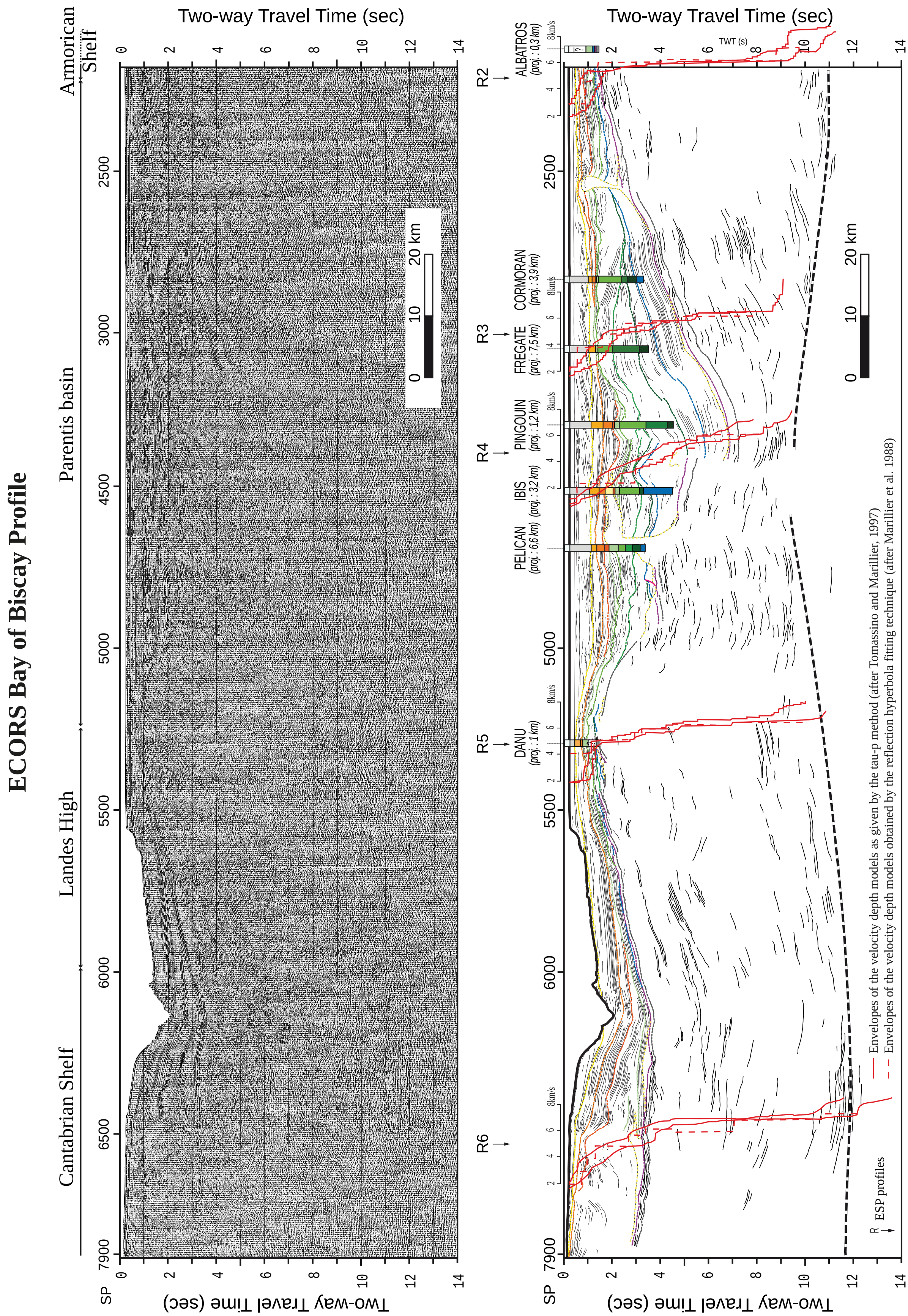
ECORS Arzacq-W Pyrenees Profile

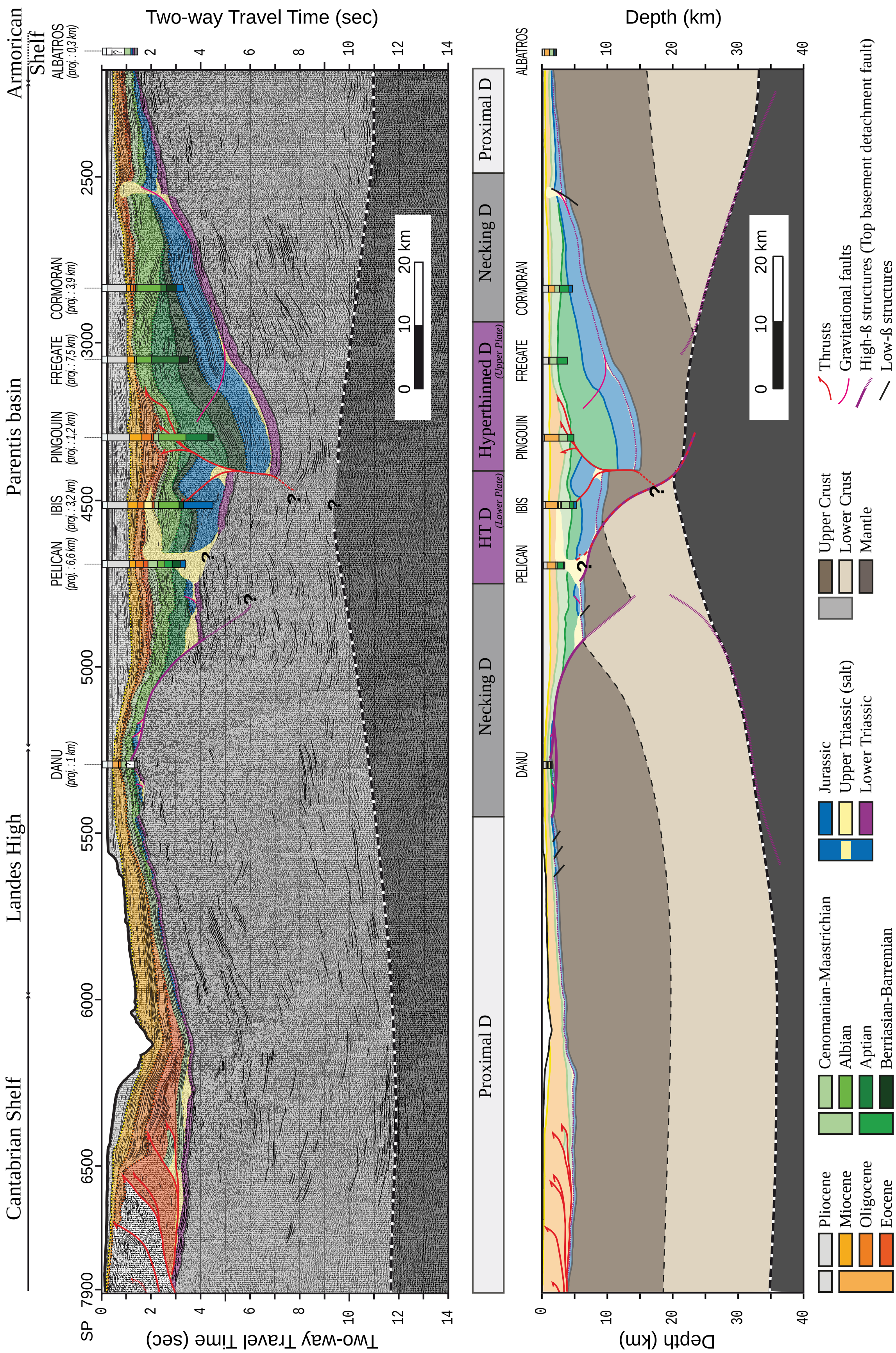


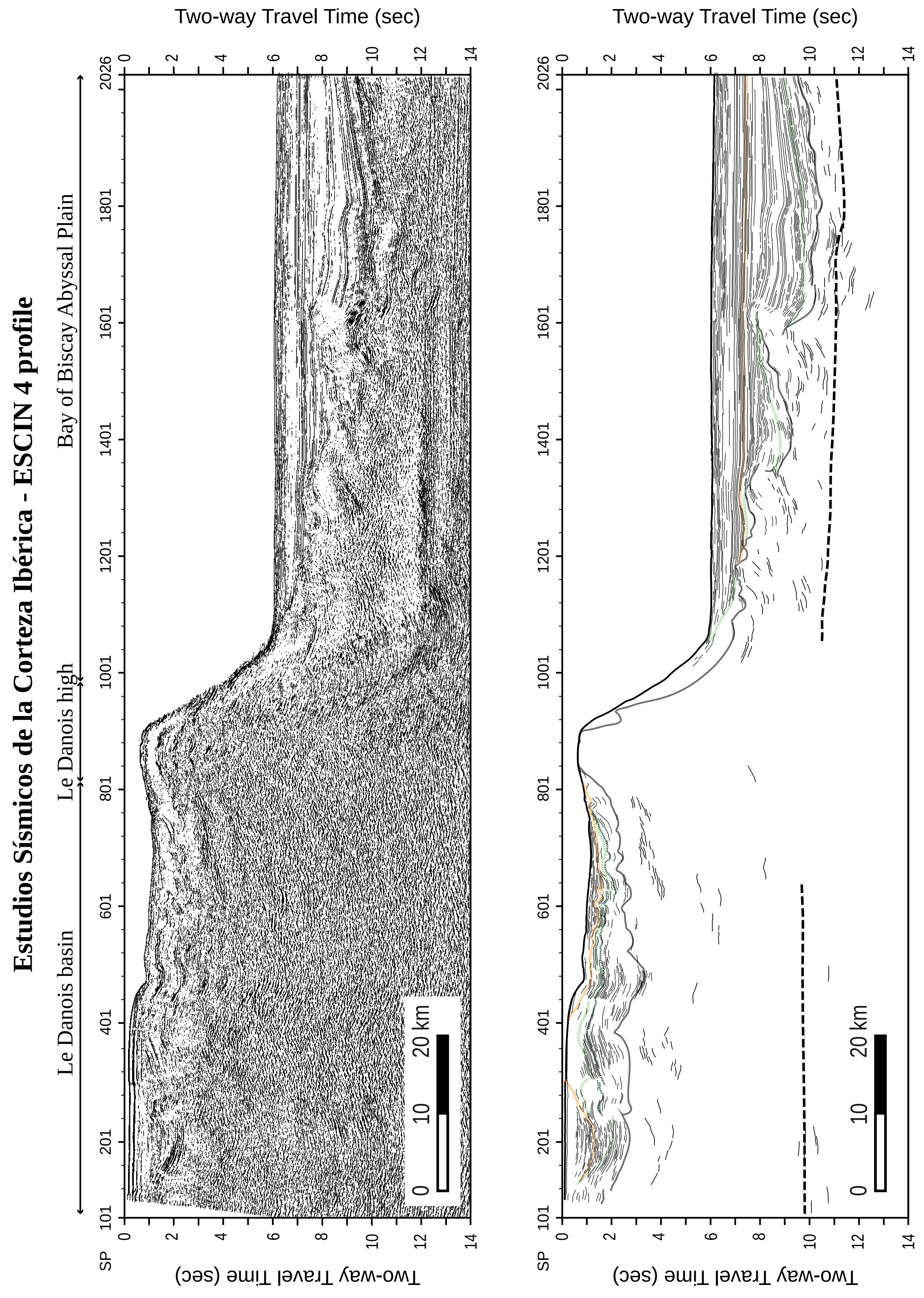


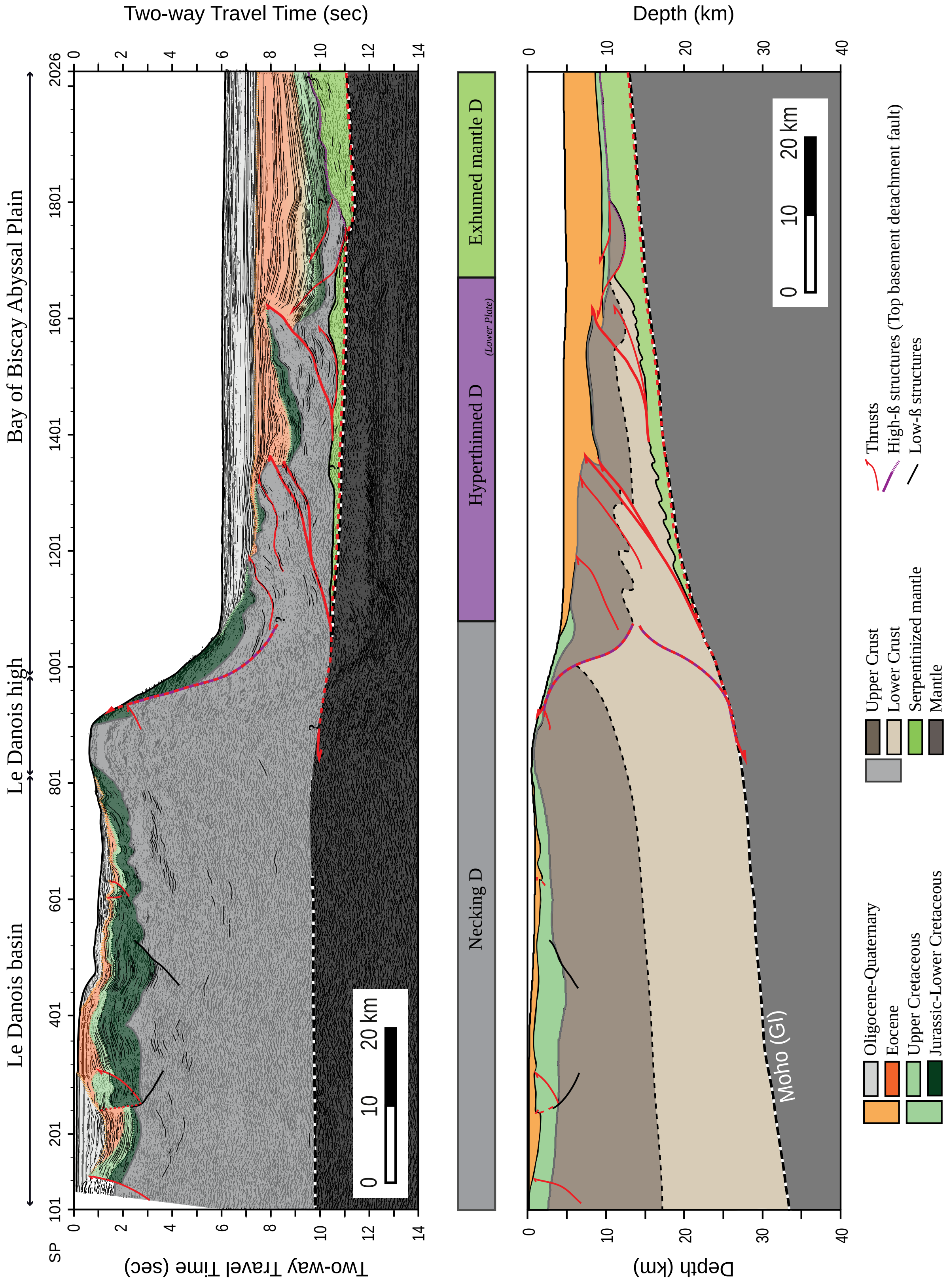
ECORS Arzacq-W Pyrenees Profile



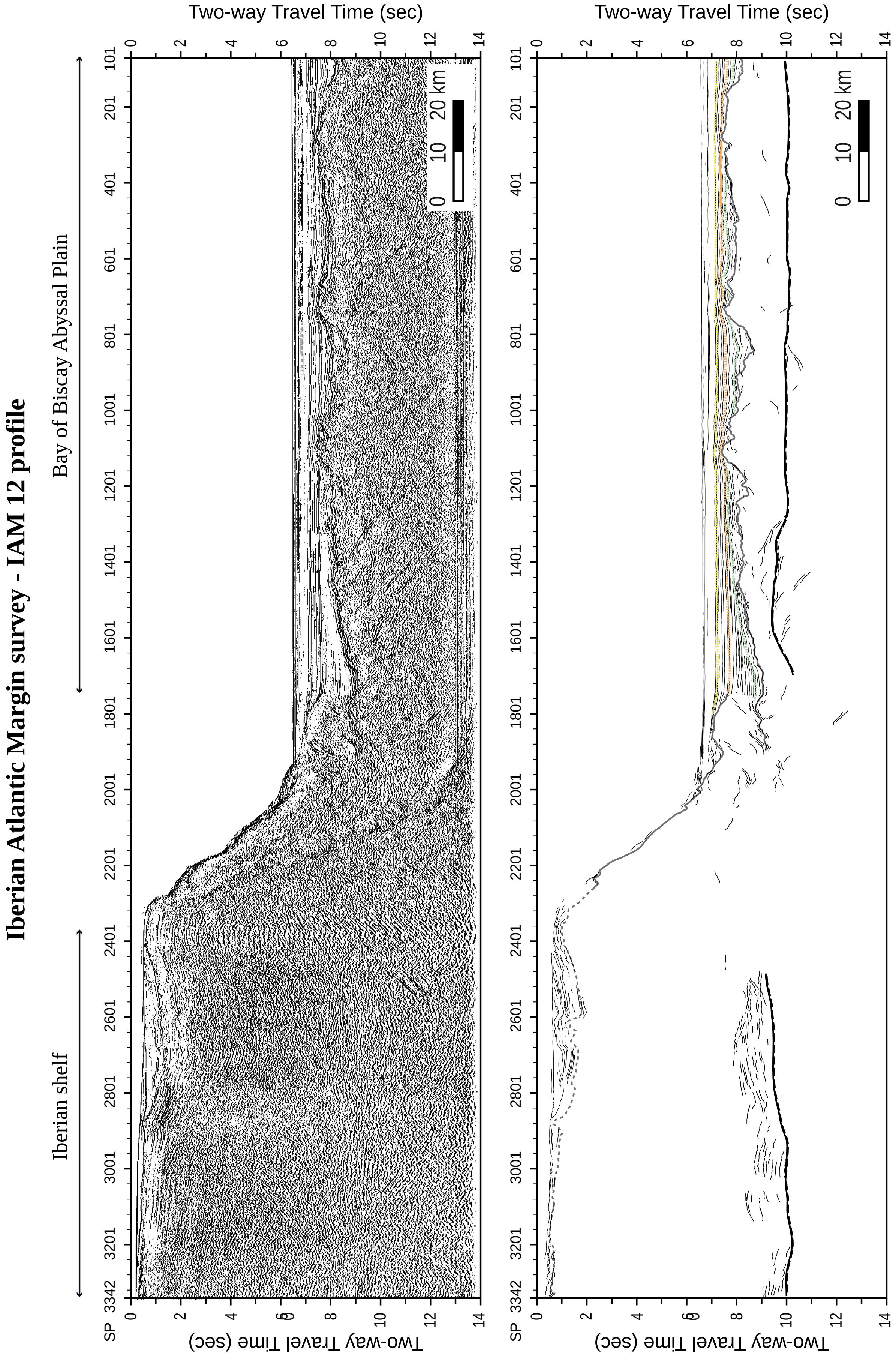

ECORS Bay of Biscay Profile
Armorican Shelf
Parentis basin
Landes High
Cantabrian Shelf
Two-way Travel Time (sec)
SP
0 10 20 km
R2
R3
R4
R5
R6
ALBATROS (proj. : 0.3 km)
CORMORAN (proj. : 3.9 km)
FREGATE (proj. : 7.5 km)
PINGOUIN (proj. : 1.2 km)
IBIS (proj. : 3.2 km)
PELICAN (proj. : 6.6 km)
DANU (proj. : 1 km)
TWT (s)
8km/s
Envelopes of the velocity depth models as given by the tau-p method (after Tomassino and Marillier, 1997)
Envelopes of the velocity depth models obtained by the reflection hyperbola fitting technique (after Marillier et al. 1988)
R ESP profiles

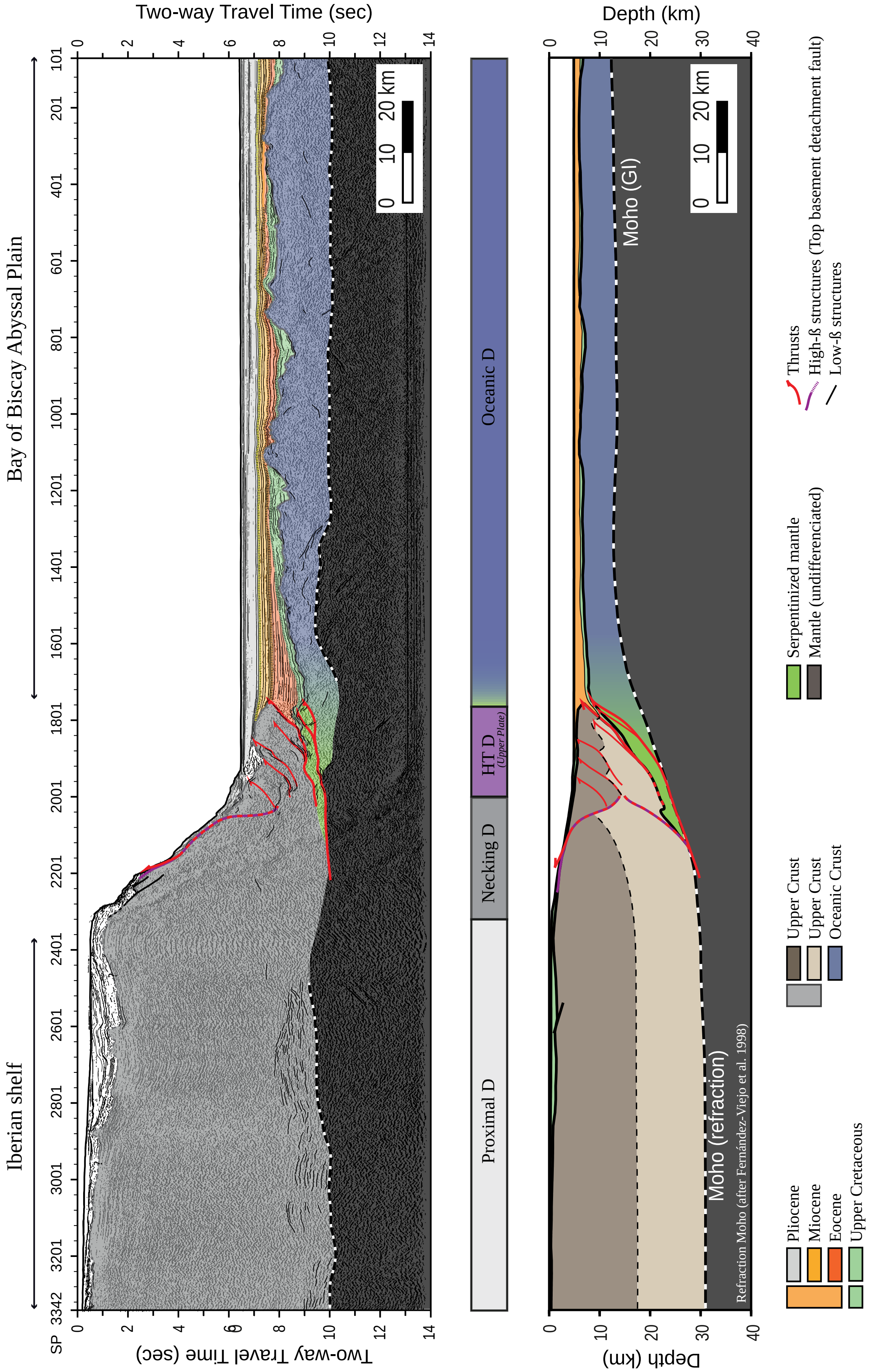

Armorican Shelf
Parentis basin
Landes High
Cantabrian Shelf
Two-way Travel Time (sec)
Depth (km)
SP
ALBATROS (proj.: 0.3 km)
CORMORAN (proj.: 3.9 km)
FREGATE (proj.: 7.5 km)
PINGOUIN (proj.: 1.2 km)
IBIS (proj.: 3.2 km)
PELICAN (proj.: 6.6 km)
DANU (proj.: 1 km)
Proximal D
Necking D
Hyperthinned D (Upper Plate)
HT D (Lower Plate)
Pliocene
Miocene
Oligocene
Eocene
Cenomanian-Maastrichian
Albian
Aptian
Berriasian-Barremian
Jurassic
Upper Triassic (salt)
Lower Triassic
Upper Crust
Lower Crust
Mantle
Thrusts
Gravitational faults
High-ß structures (Top basement detachment fault)
Low-ß structures

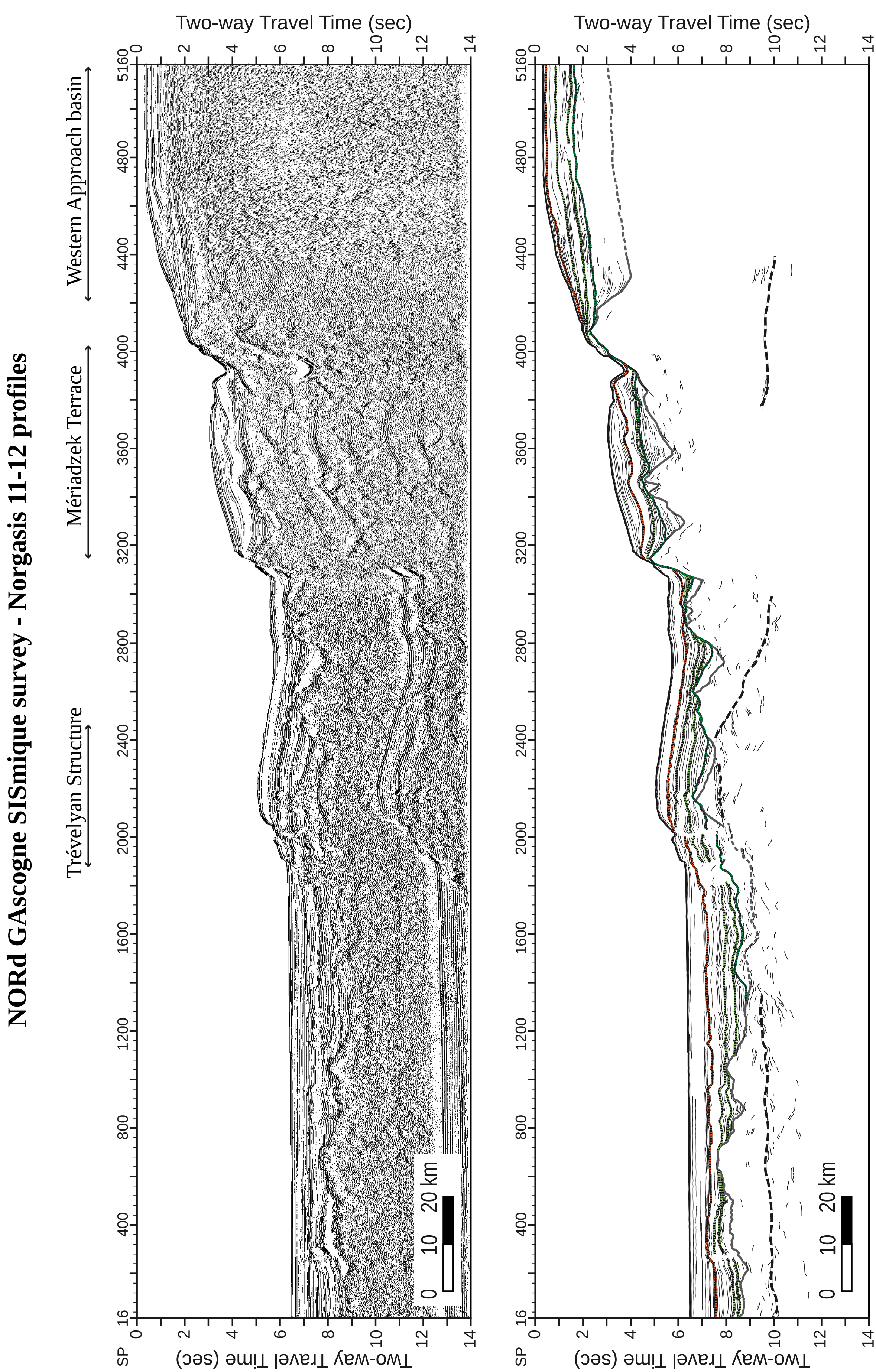

Estudios Sísmicos de la Corteza Ibérica - ESCIN 4 profile
Le Danois basin
Le Danois high
Bay of Biscay Abyssal Plain
Two-way Travel Time (sec)
SP
101
201
401
601
801
1001
1201
1401
1601
1801
2026
0
10
20 km

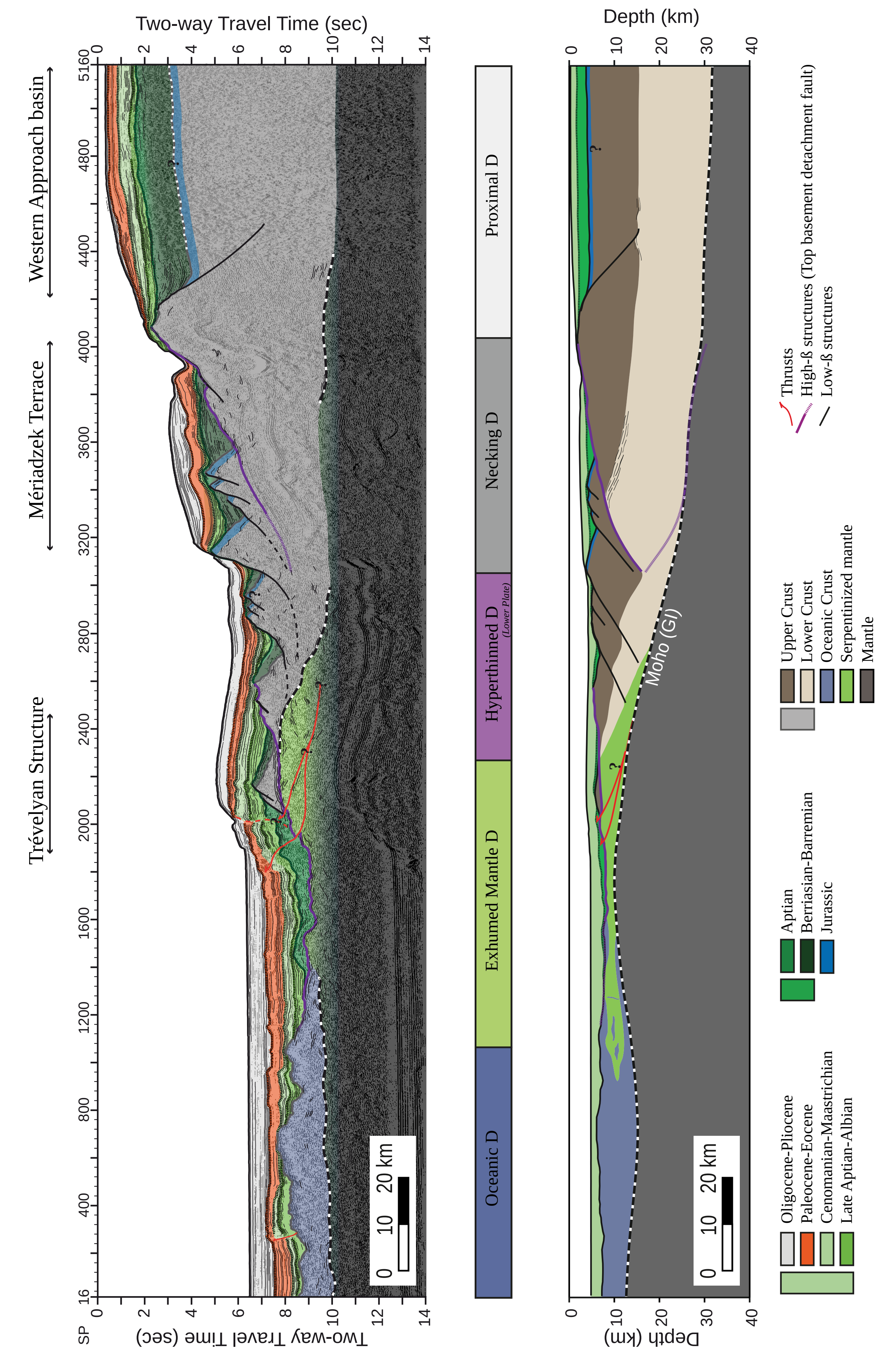


Le Danois basin
Le Danois high
Bay of Biscay Abyssal Plain
Two-way Travel Time (sec)
SP
Necking D
Hyperthinned D
(Lower Plate)
Exhumed mantle D
Depth (km)
Moho (Gl)
20 km
Oligocene-Quaternary
Eocene
Upper Cretaceous
Jurassic-Lower Cretaceous
Upper Crust
Lower Crust
Serpentinized mantle
Mantle
Thrusts
High-ß structures (Top basement detachment fault)
Low-ß structures

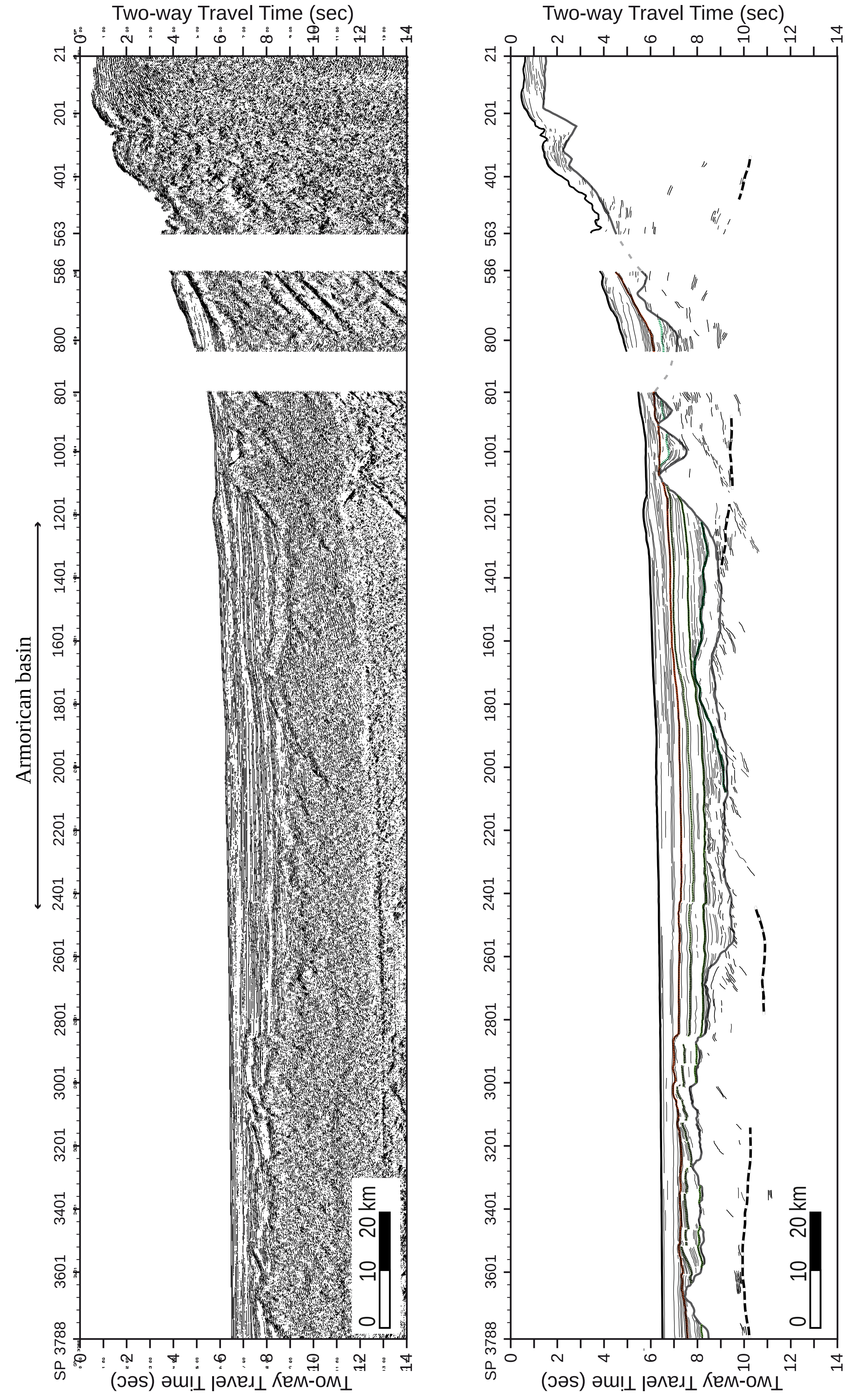

Iberian Atlantic Margin survey - IAM 12 profile
Bay of Biscay Abyssal Plain
Iberian shelf
Two-way Travel Time (sec)
0 10 20 km
SP
101
201
401
601
801
1001
1201
1401
1601
1801
2001
2201
2401
2601
2801
3001
3201
3342
0
2
4
6
8
10
12
14

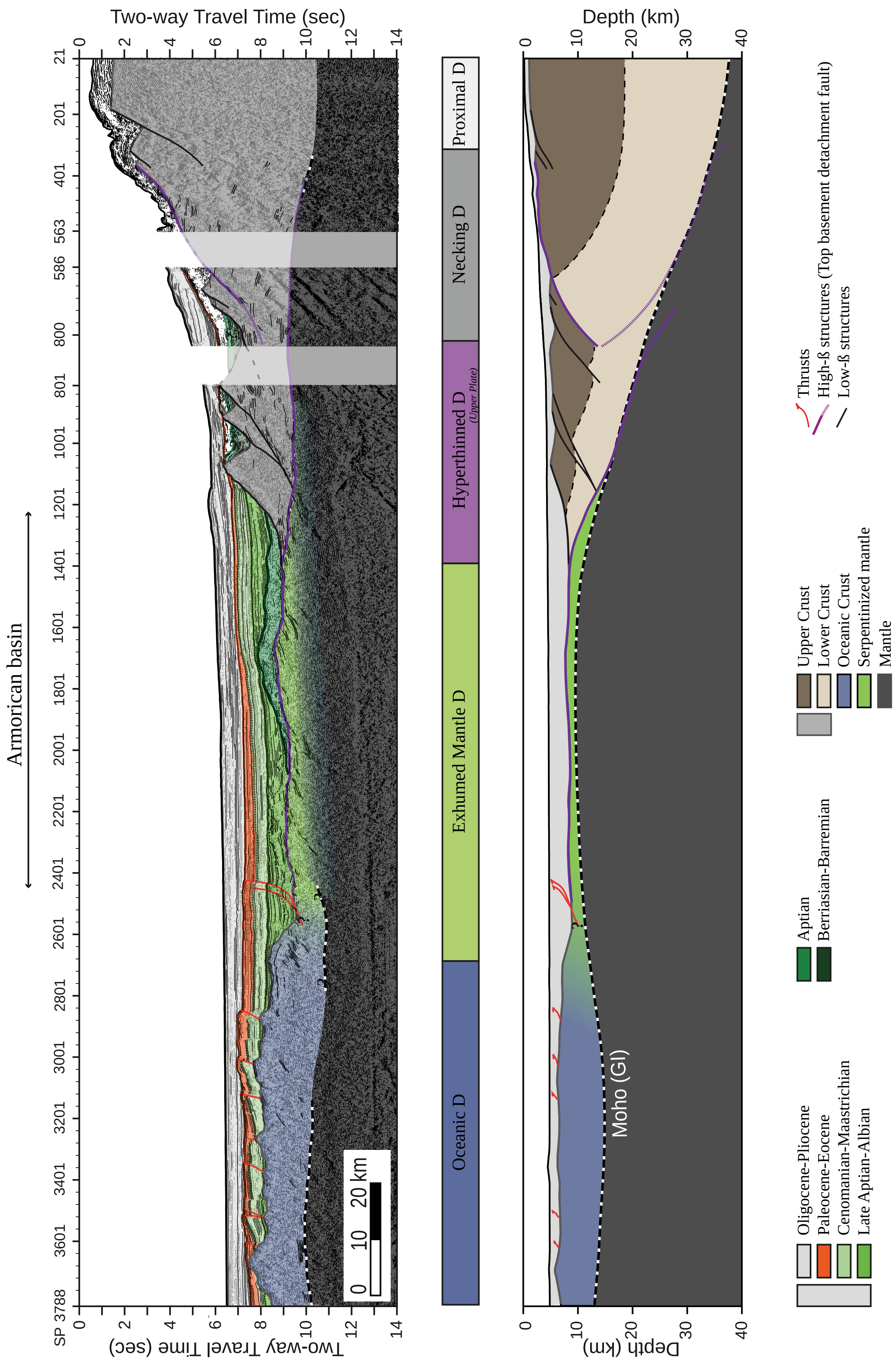

Two-way Travel Time (sec)
Bay of Biscay Abyssal Plain
Iberian shelf
SP
0 10 20 km
Oceanic D
HT D
(Upper Plate)
Necking D
Proximal D
Depth (km)
Moho (GI)
Moho (refraction)
Refraction Moho (after Fernández-Viejo et al. 1998)
Pliocene
Miocene
Eocene
Upper Cretaceous
Upper Crust
Upper Crust
Oceanic Crust
Serpentinized mantle
Mantle (undifferenciated)
Thrusts
High-ß structures (Top basement detachment fault)
Low-ß structures


NORd GAscogne SISmique survey - Norgasis 11-12 profiles
Western Approach basin
Mériadzek Terrace
Trévelyan Structure
Two-way Travel Time (sec)
SP
0 10 20 km


Western Approach basin
Mériadzek Terrace
Trévelyan Structure
Two-way Travel Time (sec)
SP
Proximal D
Necking D
Hyperthinned D (Lower Plate)
Exhumed Mantle D
Oceanic D
0 10 20 km
Depth (km)
Moho (GI)
Oligocene-Pliocene
Paleocene-Eocene
Cenomanian-Maastrichtian
Late Aptian-Albian
Aptian
Berriasian-Barremian
Jurassic
Upper Crust
Lower Crust
Oceanic Crust
Serpentinized mantle
Mantle
Thrusts
High-ß structures (Top basement detachment fault)
Low-ß structures

NORd GAscogne SISmique survey - Norgasis 14 profile




Two-way Travel Time (sec)
Depth (km)
Armorican basin
Proximal D
Necking D
Hyperthinned D
(Upper Plate)
Exhumed Mantle D
Oceanic D
Moho (GI)
SP 3788
0 10 20 km
Oligocene-Pliocene
Paleocene-Eocene
Cenomanian-Maastrichian
Late Aptian-Albian
Aptian
Berriasian-Barremian
Upper Crust
Lower Crust
Oceanic Crust
Serpentinized mantle
Mantle
Thrusts
High-ß structures (Top basement detachment fault)
Low-ß structures

Annexe 4: Les analyses quantitatives

Gravity Inversion

Residual Depth Anomaly (RDA)

Subsidence Analysis

Synthesis of results

Gravity Inversion

Datasets and Methods

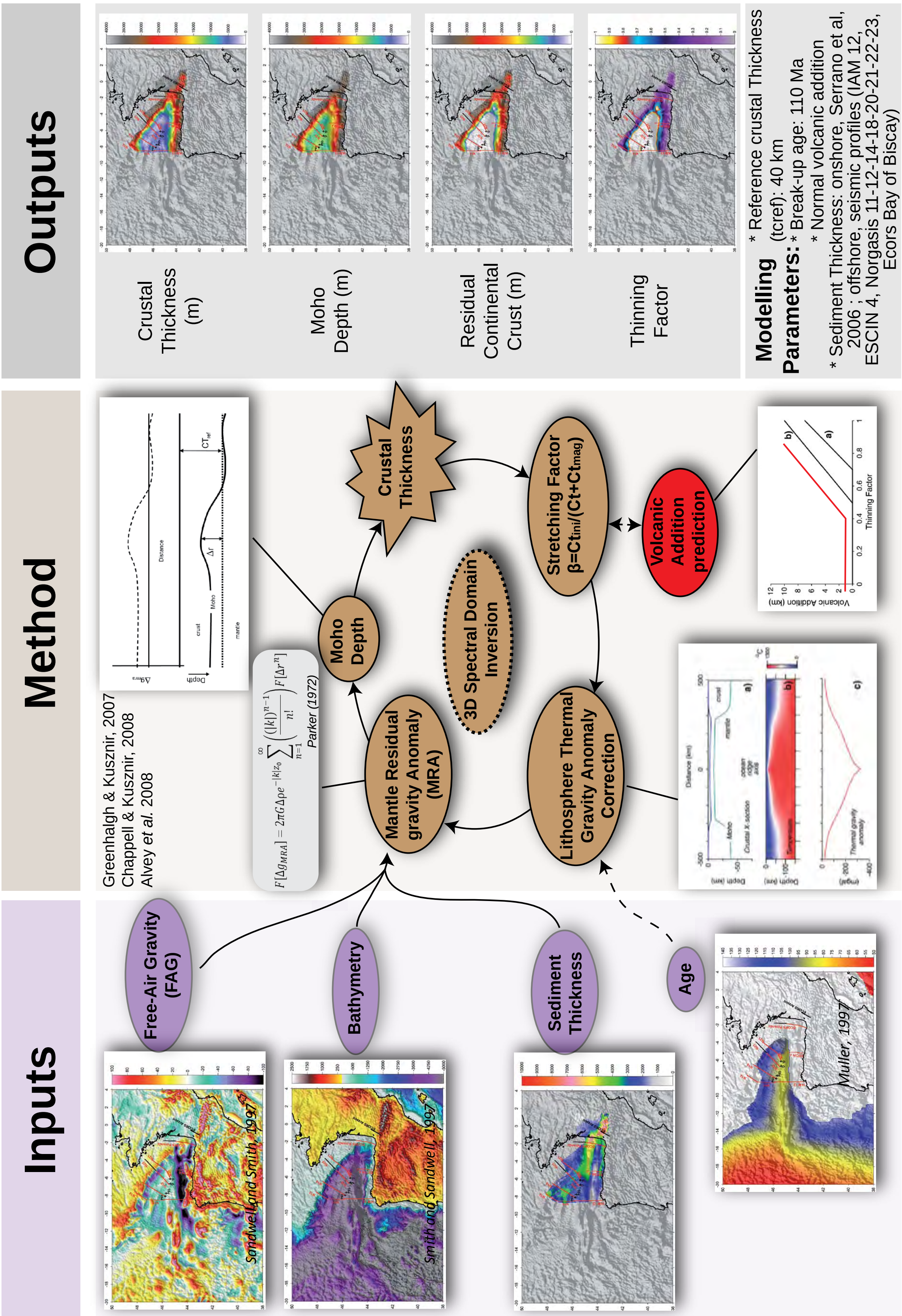
Parameter Calibration

Reference Crustal Thickness

Sediment thickness

Volcanic Addition

Gravity Inversion - Datasets and Methods



Moho depth, crustal thickness, continental lithosphere thinning factor and residual continental crust maps were determined from gravity inversion. This technique, summarized in the following, is an iterative approach. Gravity inversion is based on public domain data: free air gravity (Sandwell & Smith, 2009), bathymetry (Smith & Sandwell, 1997) and oceanic isochrones (Müller *et al.,* 1997). Information on sedimentary thickness comes from a compilation of offshore seismic interpretations combined with the depth to basement map of the Aquitaine basin from Serrano *et al.* (2006). Seismic reflection data are derived from different surveys: the Norgasis survey (Avedik *et al.,* 1993, 1996; Thinon 1999), the ECORS Bay of Biscay section (Pinet *et al.,* 1987; Bois & Gariel, 1994), the IAM 12 and the ESCIN 4 seismic lines (e.g. Banda *et al.,* 1995; Gallastegui *et al.,* 2002; Gallart *et al.,* 2004).

In order to predict the Moho depth from gravity inversion at rifted continental margins and in oceanic lithosphere, it is necessary to determine the gravity anomaly due to Moho topography variations. The observed free-air gravity anomaly (Δg_{fag}) mainly depends on the gravity anomaly signal from bathymetry (Δg_b) and sediment thickness (Δg_s), the Moho relief anomaly (Δg_{mra}) and the lithosphere thermal gravity anomaly (Δg_t).

$$\Delta g_{fag} = \Delta g_b + \Delta g_s + \Delta g_{mra} + \Delta g_t$$

Therefore, Moho relief anomaly (Δg_{mra}) is given by:

$$\Delta g_{mra} = \Delta g_{fag} - \Delta g_b - \Delta g_s - \Delta g_t$$

The gravity inversion of Moho relief anomaly (Δg_{mra}) is carried out in the 3D spectral domain based on the scheme of Parker (1972) to determine Moho relief (Δr). In order to remove short-wavelength errors in Δg_b and Δg_s and gravity anomalies due to crustal basement density variations, a Butterworth low-pass filter is used with a cut-off wavelength of 100 km prior to gravity inversion. Moho depth d and basement crustal thickness (C_t) may be calculated from the Moho relief (Δr), a reference Moho depth (C_{tref}) and bathymetry.

$$d = Ct_{ref} + \Delta r$$

$$C_t = d - b$$

The reference Moho depth (C_{tref}) is strongly influenced by mantle dynamic topography and represents the thickness of a piece of crust that has zero bathymetry, zero sediment thickness and zero free-air gravity anomaly. It may be calibrated using seismic refraction data (see model calibration).

A crustal basement density of 2850kgm^{-3} is assumed for both oceanic and continental crust (Chappel & Kusznir 2008) in order to avoid proposing a priori information on the ocean-continent boundary.

A gravity anomaly correction is made for the sediments (Δg_s) assuming a compaction controlled density-depth relationship for sedimentary sequences.

The initiation of thinning processes at rifted continental margins that may finally result in seafloor spreading in oceanic domains gives an elevated lithosphere geotherm. The lithosphere thermal gravity anomaly ($\Delta g_{t)}$ is caused by lateral variations in density and temperature within this complex structuration of the lithosphere. Detailed descriptions of the lithosphere thermal gravity anomaly correction (Δg_t) are proposed by Greenhald & Kusznir, 2007, Alvey *et al.* 2008 and Chappell & Kusznir 2008. The lithosphere thermal gravity anomaly correction (Δg_t) is based on the lithosphere stretching and thinning model of McKenzie 1978. The lithosphere stretching factor, β, is used at each model location to define the lithosphere thermal perturbation and the lithosphere thermal re-equilibration time. For oceanic lithosphere $\beta=\infty$ and for continental margin lithosphere, the lithosphere β factor is determined from gravity inversion results using:

$$\beta = \frac{Ct_{ini}}{Ct - Ct_{mag}}$$

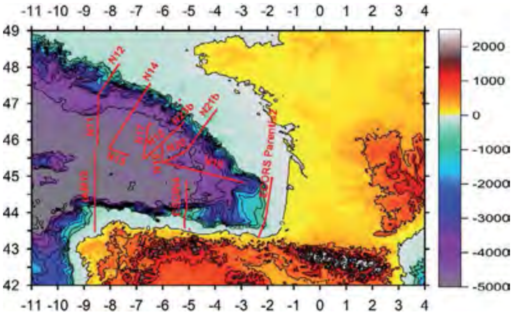
Where Ct_{ini} is the initial crustal thickness, Ct is the present-day thickness derived from the gravity inversion process and Ct_{mag} corresponds to the volcanic addition thickness from decompression melting (White and McKenzie 1989). The lithosphere is assumed to deform by pure shear, i.e. lithosphere stretching and thinning is equal to crustal stretching and thinning (*Greenhald & Kusznir,* 2007, Alvey *et al.* 2008 and *Chappell & Kusznir* 2008). The thickness of volcanic addition may be estimated from the decompression melting model of *White and McKenzie* 1989, using the lithosphere thinning factor, γ, determined from gravity inversion, where

$$\gamma = 1 - \frac{1}{\beta}$$

A critical thinning factor for the initiation of decompression melting must be defined as well as a maximum volcanic addition predicting a maximum oceanic crustal thickness. The thermal equilibration time of continental margin lithosphere is defined by the break-up age.

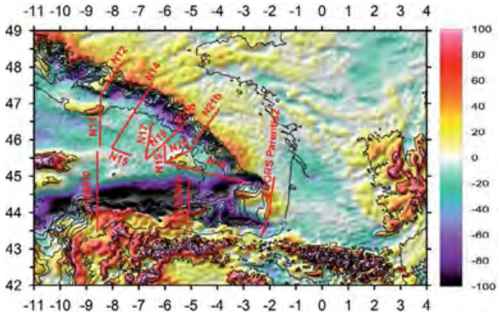
Gravity Inversion – Input data

Bathymetry



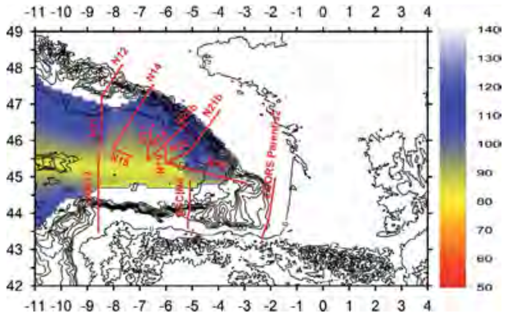
Smith and Sandwell, 1997

Free Air Gravity



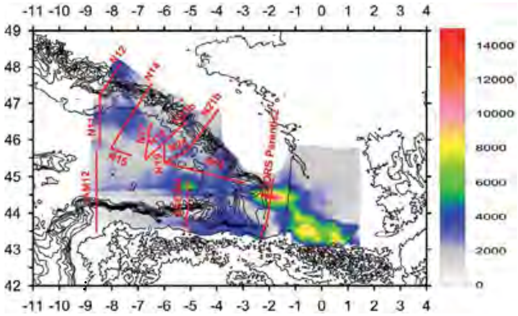
Sandwell and Smith, 1997

Oceanic Isochrons (Age)



Muller, 1997

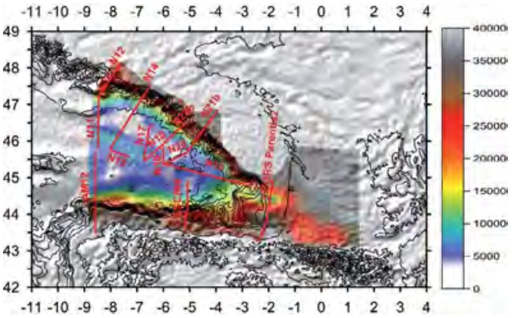
Sediment Thickness



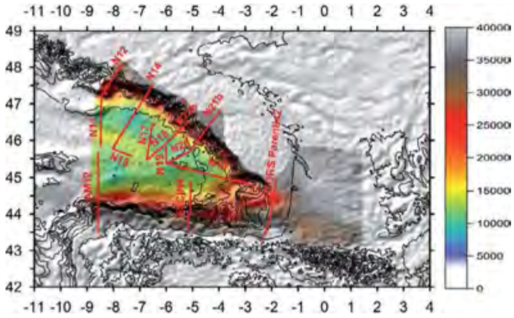
Onshore Serrano et al, 2006
Offshore: Seismic profiles (IAM12, ESCIN4, Norgasis 10-11-12 14-15-17-18-19-20-21-22-23, ECORS Bay of Biscay)

Gravity Inversion – Output data

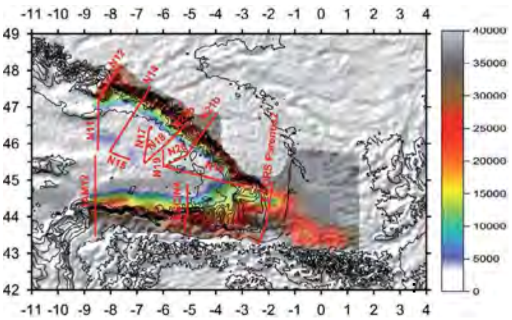
Crustal Thickness (m)



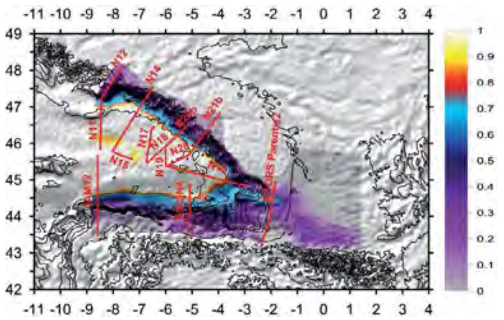
Moho Depth (m)



Residual Continental Crust (m)



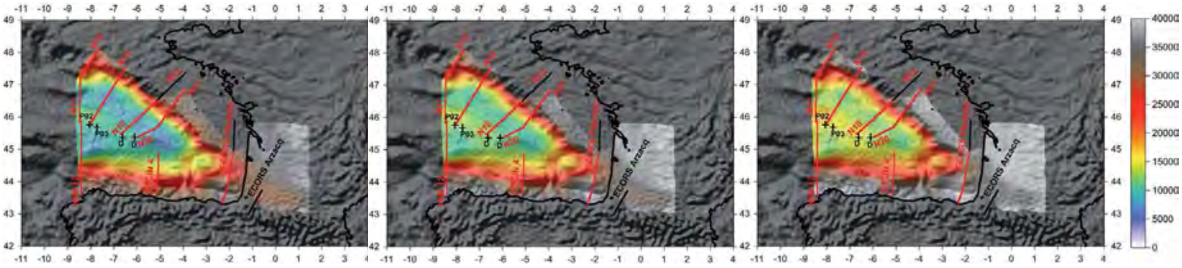
Thinning Factor



Gravity Inversion – Reference Crustal Thickness Calibration & Sensitivity to Ctref

Moho depth is deepening with increasing reference crustal thickness (Ctref)

Moho Depth, Ctref=37,5 km **Moho Depth, Ctref=40 km** **Moho Depth, Ctref=42,5 km**



Refraction Data:

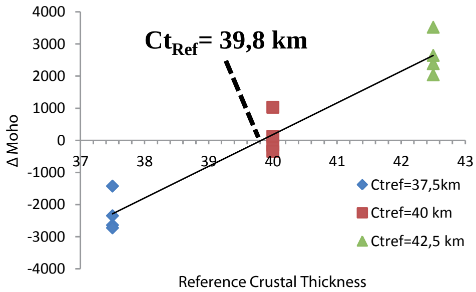
- P02, Luis Mattias, in Thinon 1999
- P03, Thinon et al 2003
- D profile, Avedik and Howard, 1979

Gravity inversion parameters:

- Reference Crustal Thickness: 37,5 km, 40 km, 42,5 km
- Sediment Thickness (offshore: Seismic profile onshore: BRGM, 2006)
- Break-up age: 110.0 Ma
- Normal volcanic addition

B1_ctref40_inct35_NorVol_g0.7_oc7_ra110_Soft_oldisoch110_Julie_Seds
B2_ctref37.5_inct35_NorVol_g0.7_oc7_ra110_Soft_oldisoch110_Julie_Seds
B3_ctref42.5_inct35_NorVol_g0.7_oc7_ra110_Soft_oldisoch110_Julie_Seds

Reference Crustal Thickness Calibration

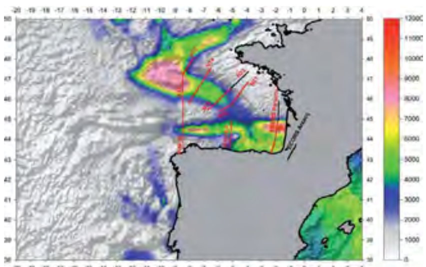


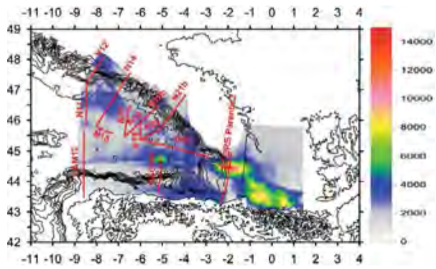
ΔZ moho is the difference between the **Refraction Moho Depth** and the **Gravity inversion MohoDepth**

Gravity Inversion – Sensitivity to Sediment Thickness

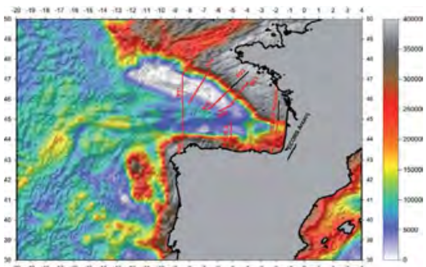
Sediment thickness determines the precision of gravity inversion results. Underestimations of sediment thickness leads to overestimations of crustal thickness derived from gravity inversion.

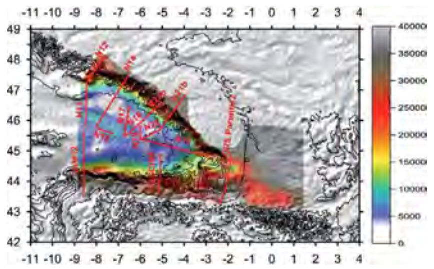
INPUT DATA: SEDIMENT THICKNESS





CRUSTAL THICKNESS MAPS FROM GRAVITY INVERSION



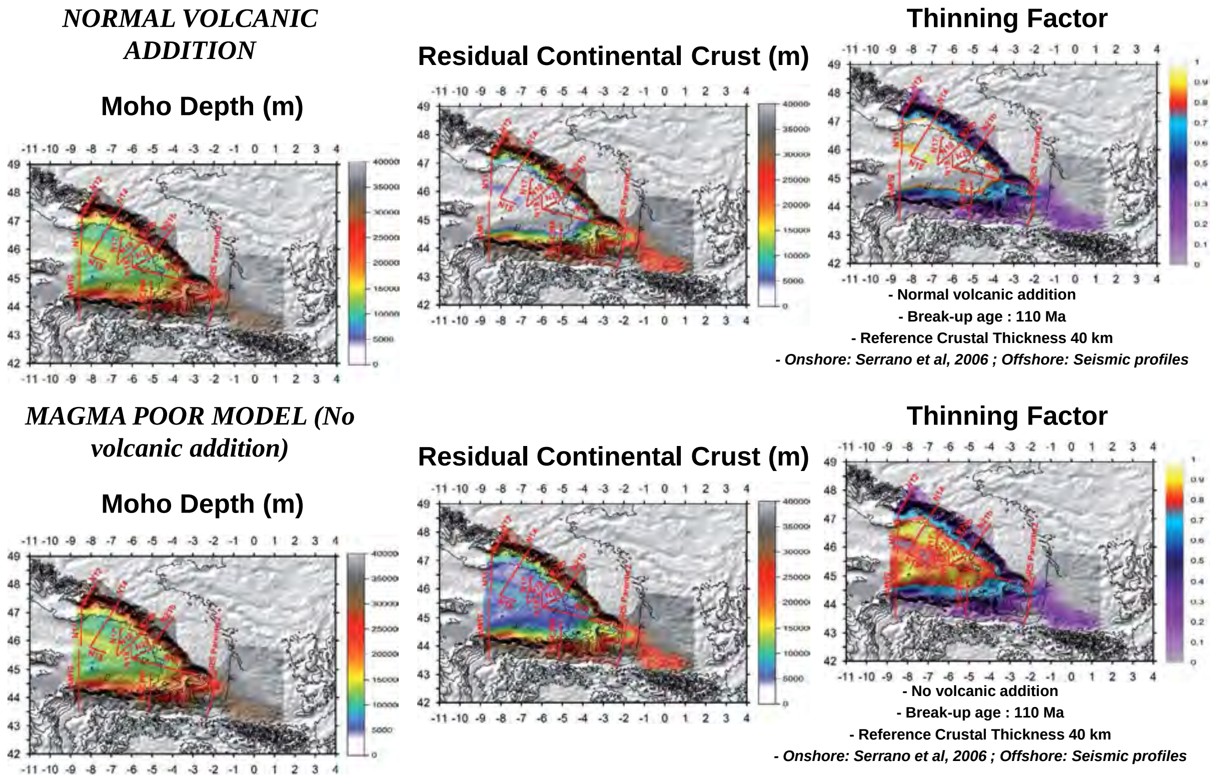


- Reference crustal Thickness: 40 km
- Normal Volcanic addition: 7 km
-Break-up age: 110 Ma
- Sediment thickness: NOAA database

- Reference crustal Thickness: 40 km
- Normal Volcanic addition: 7 km
-Break-up age: 110 Ma
***- Sediment thickness:** Onshore: Serrano et al, 2006 Offshore: IAM 12, ESCIN 4, Norgasis 10-11-12-14-15-17-18-19-20-21-22-23, ECORS Bay of Biscay)*

Gravity Inversion – Sensitivity to Volcanic Addition

(Residual continental crust & Thinning factor variation)



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Residual Depth Anomaly

Sediment Corrected RDA

Synthetic RDA

Delta RDA

Residual Depth Anomaly - Method

Sediment Corrected RDA

$$RDA_{sed} = B_{ob} - B_{pr}$$

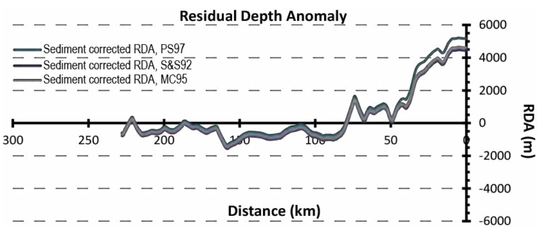
B_{obs} is corrected from sediment load by flexural backstriping

B_{pr} is calculated from the thermal plate model of:

- Parsons and Sclater (1997) =>PS97
- Stein and Stein (1992) => S&S92
- Mc Kenzie et al. (2005)=>MC05

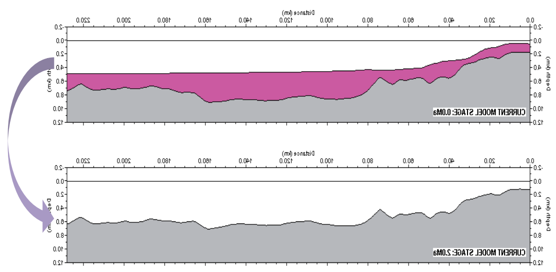
Sediment Corrected RDA

Armorican margin: Norgasis 14



Sediment corrected Bathymetry from Flexural Backstriping

Norgasis 14



Positive values indicate that bathymetry is shallower than what is expected from the thermal models

No indication on the origin of the subsidence anomaly

Residual Depth Anomaly - Method

Synthetic RDA

$$RDA_{synt} = \frac{(tc_g - toc_{ref})(\rho_m - \rho_c)}{\rho_m - \rho_{sw}}$$

tc_g : crustal thickness from gravity inversion

toc_{ref} : global averal oceanic crustal thickness (7,1km, White et al, 1992)

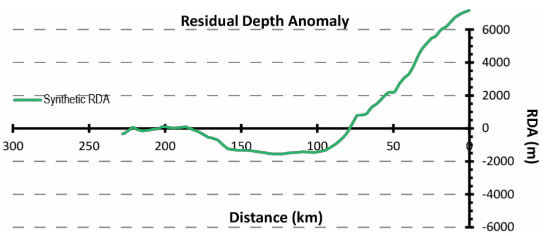
ρ_m: average mantle density (3300 kgm^{-3})

ρ_c: average crustal density (2850 kgm^{-3})

ρ_{sw}: average density of sea water (1500 kgm^{-3})

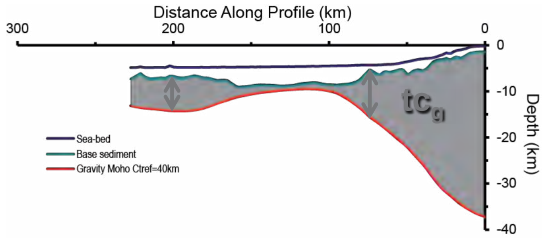
Synthetic RDA

Armorican margin: Norgasis 14



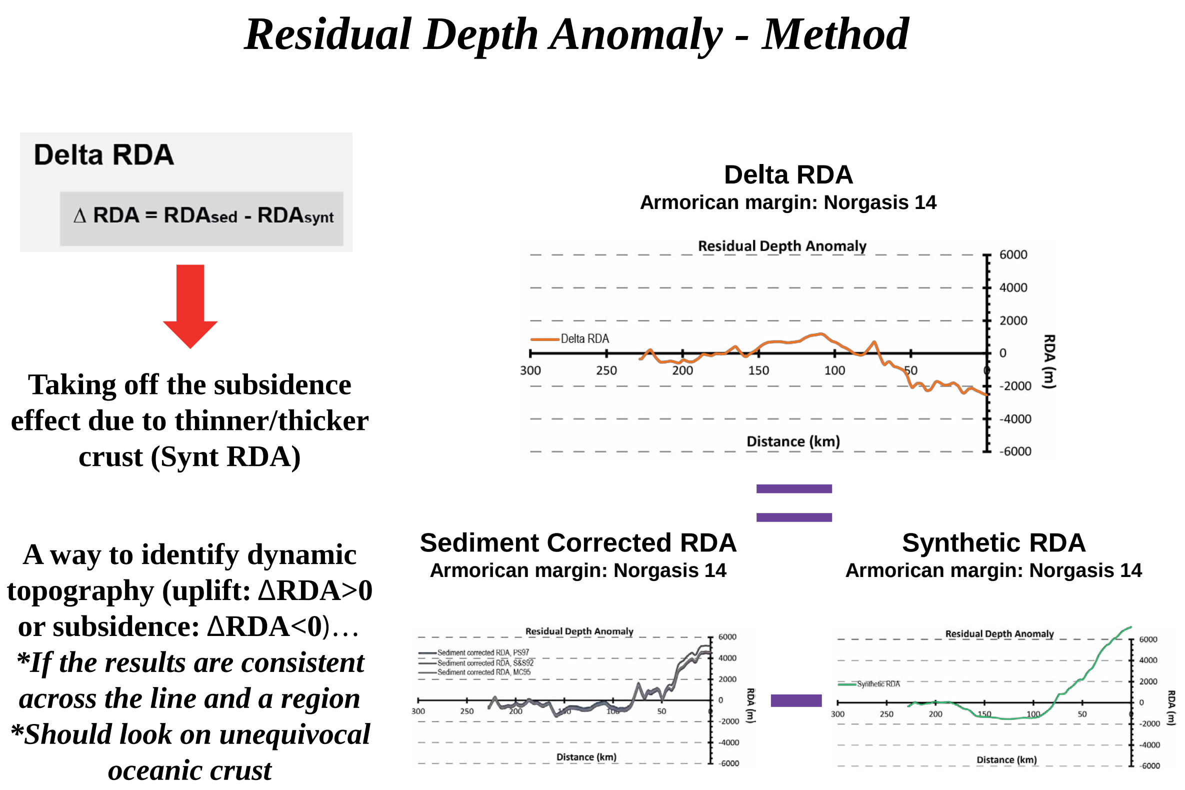
Crustal thickness from gravity inversion (tc_g)

Norgasis 14



Local isostatic response to variation in gravity-inverted crustal thickness from the 7,1 km global oceanic average [White et al., 1992]

Subsidence anomaly due to thinner/thicker crust (compared to 7km Oceanic Crust Standard)



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SUBSIDENCE ANALYSIS

METHODOLOGY

INFLUENCE OF PARAMETERS:

- Volcanic Addition
- Initial Crustal Thickness
- Crustal density
- Break Up time (Post rift time)

CONTROL OF PARAMETERS USED FOR THE MODEL

SUBSIDENCE ANALYSIS - Work flow

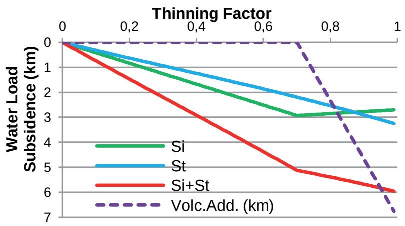
Subsidence Widget Software

SUBSIDENCE WIDGET PARAMETERS:

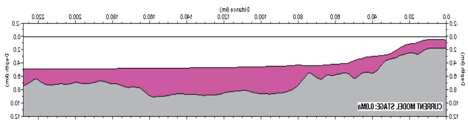
- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = 0.7
- Volcanic Addition thickness = 7km
- Break Up Age (Post rift time) = 110 Myr

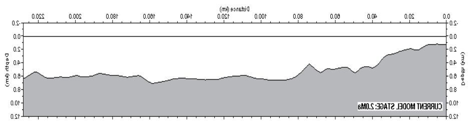
RESULTS (BASED ON MCKENZIE SUBIDENCE MODEL):

- Waterloaded subsidence (S) evolution as a fonction of the thinning factor
- **Si: initial subsidence**
- **St: thermal subsidence**
- ***Si+St: total subsidence***



Waterloaded subsidence from Flexural Backstriping

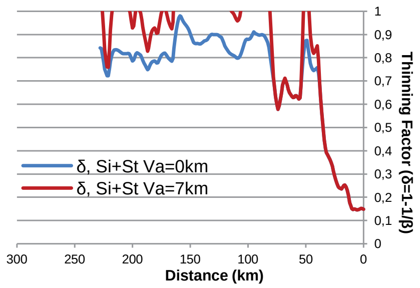


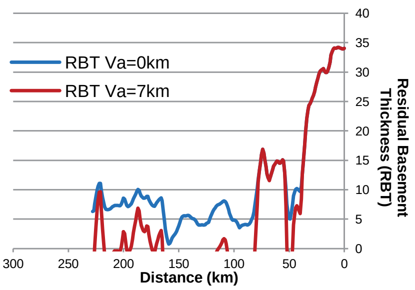


Norgasis 14

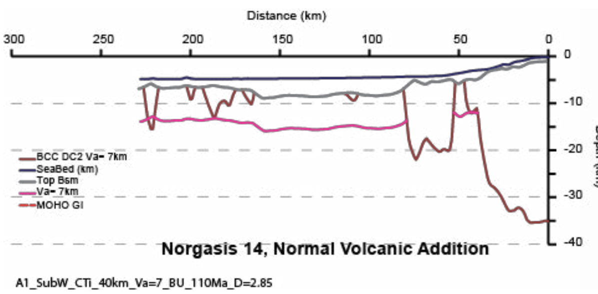
Thinning Factor & Residual Basement Thickness (RBT)

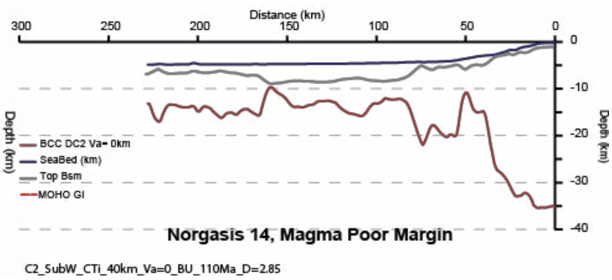
For a Normal Volcanic (Va=7km) & Magma Poor Solution (Va=0km)

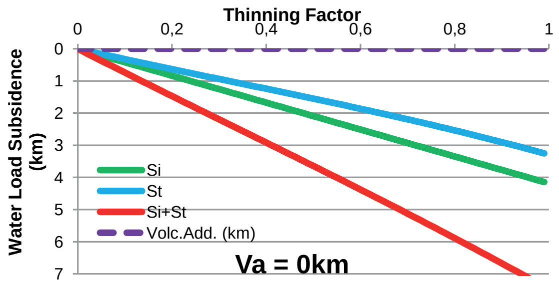


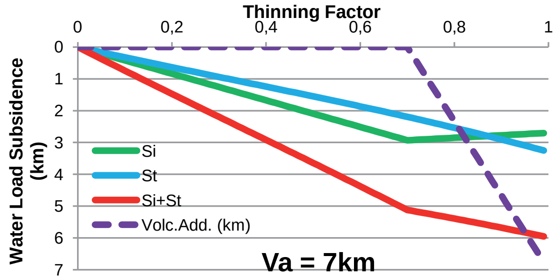


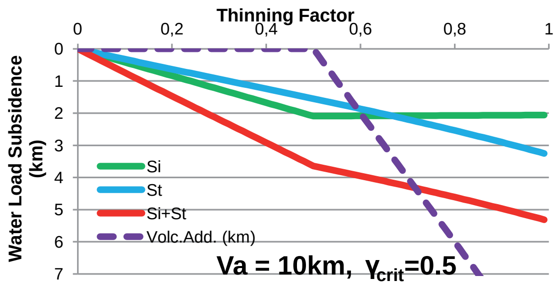
Normal Volcanic & Magma Poor Profile from Subsidence analysis











INFLUENCE OF PARAMETERS:
- VOLCANIC ADDITION

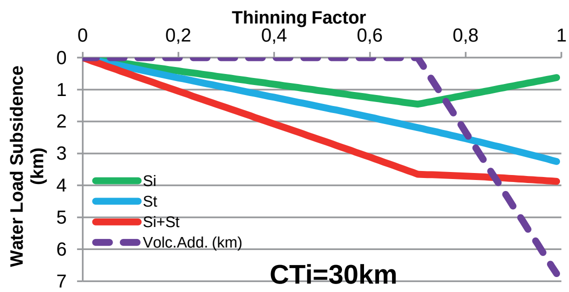
VOLCANIC ADDITION:

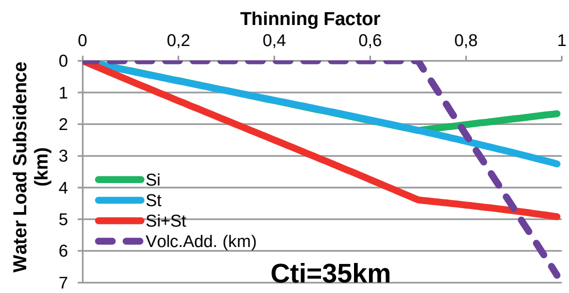
• **No volcanic addition (Va=0km):** there is an almost linear relashionship between the subsidence and the thinning factor evolution

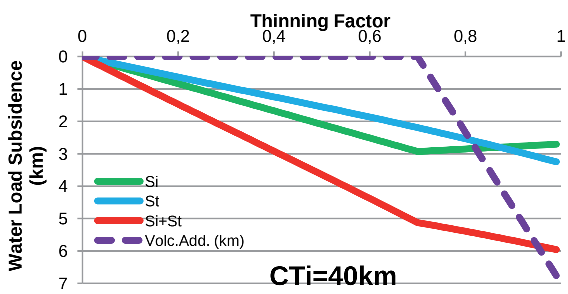
• **Volcanic addition:**
-No influence on the thermal subsidence evolution (St)
-Total subsidence (Si+St) decreases with increasing volcanic addition thickness

SUBSIDENCE WIDGET PARAMETERS:

- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = *0.7 & 0.5*
- Volcanic Addition thickness = **0km ; 7km ; 10km**
- Break Up Age (Post rift time) = 110 Myr







INFLUENCE OF PARAMETERS:
- INITIAL CRUSTAL THICKNESS

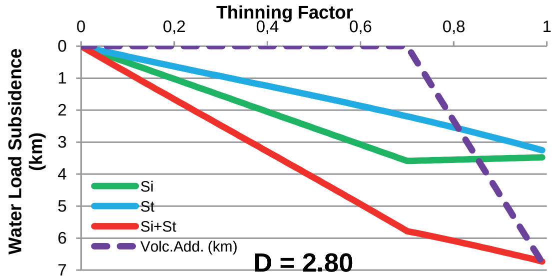
INITIAL CRUSTAL THICKNESS:

• **No influence on the thermal subsidence evolution (St)**

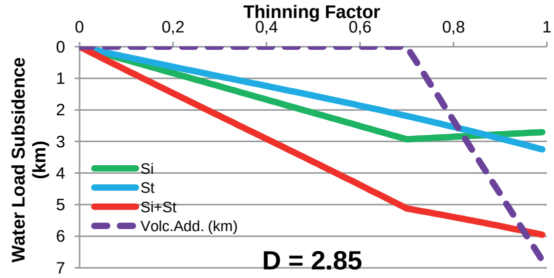
• **Syn-rift (Si) and total subsidence (Si+St) increase if the initial crust is thicker**

SUBSIDENCE WIDGET PARAMETERS:

- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = **30, 35, 40 km**
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = 0.7
- Volcanic Addition thickness = 7 km
- Break Up Age (Post rift time) = 110 Myr

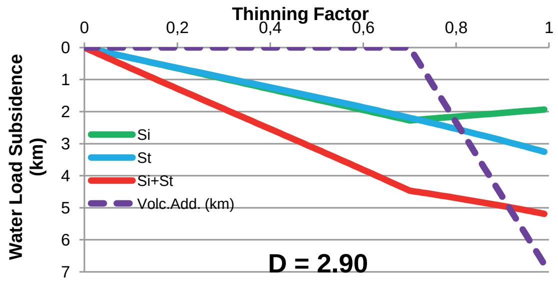


INFLUENCE OF PARAMETERS:
- CRUSTAL DENSITY



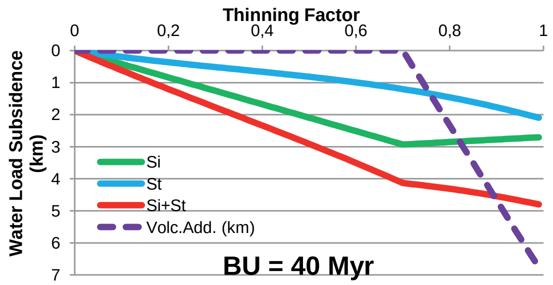
CRUSTAL DENSITY:

- **No influence on the thermal subsidence evolution (St)**
- **Syn-rift (Si) and total subsidence (Si+St) decrease for a denser crust**



SUBSIDENCE WIDGET PARAMETERS:

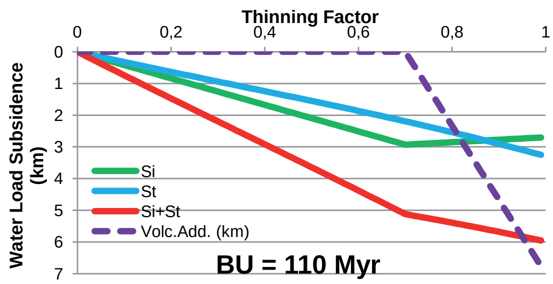
- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = 40 km
- Crustal Density = **2.80 ; 2.85 ; 2.90**
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = 0.7
- Volcanic Addition thickness = 7 km
- Break Up Age (Post rift time) = 110 Myr

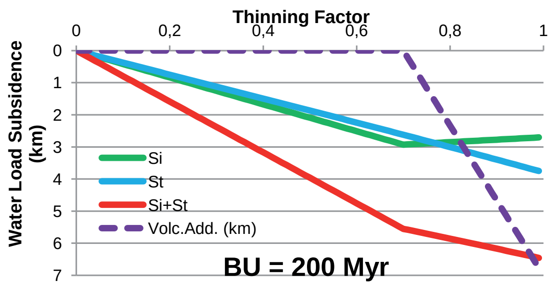


INFLUENCE OF PARAMETERS:
- BREAK UP AGE (POST RIFT TIME)

RIFT AGE:

- **No influence on the initial syn-rift subsidence evolution (Si)**
- **Syn-rift (Si) and total subsidence (Si+St) are more important in older rift basins**

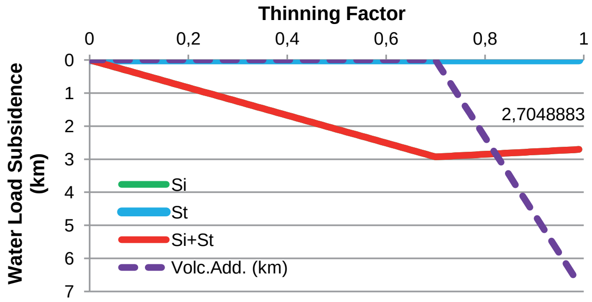




SUBSIDENCE WIDGET PARAMETERS:

- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = *0.7*
- Volcanic Addition thickness = 7km
- Break Up Age (Post rift time) = **40 ; 110 ; 200 Myr**

CONTROL OF PARAMETERS USED FOR THE MODEL



Subsidence Widget parameters:

- Lithosphere Thickness = 125 km
- Crustal Thickness (Initial) = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333 C
- Basin Infill Density = 1.0400000
- Volcanic Addition (Critical Gamma) = 0.7
- Volcanic Addition thickness = 7km
- Break Up Age (Post rift time) = **0 Myr**

Parameter control may be done by looking at the associated subsidence for a mid oceanic ridge (BU=0). The total subsidence (Si+St) should be around 2.6 km for γ=1, if there is no dynamic topography (uplift or subsidence).

Reference

MCKENZIE, D. 1978. Some remarks on the development of sedimentary basins. Earth and Planetary Science Letters, **40**, 25–32.

Synthesis of results

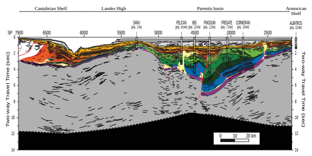
Gravity Inversion Parameters :

- Reference crustal Thickness (tcref): 40 km
- Initial Crustal Thickness = 35 km
- Crustal Density = 2.85
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr
- Sediment Thickness: *Onshore*, Serrano & al, 2006 ; *Offshore*, seismic profiles (IAM 12, ESCIN4, Norgasis 11-12-14--15-17-18-19-20-21-23, ECORS Bay of Biscay

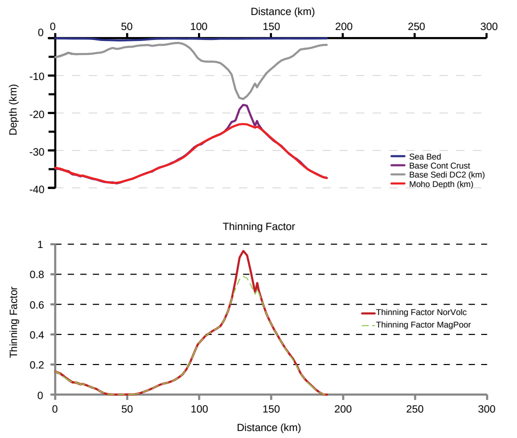
Subsidence Widget Parameters :

- Lithosphere Thickness = 125 km
- Initial Crustal Thickness = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333°C
- Basin Infill Density = 1.04
- Volcanic addition (Critical Gamma) = 0.7
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr

Normal Volcanic Addition



Gravity Inversion



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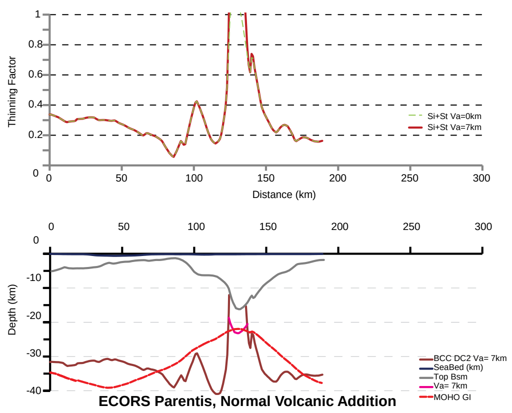
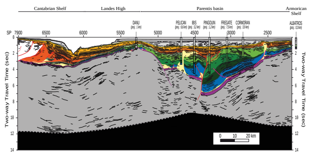
Subsidence Analysis Results

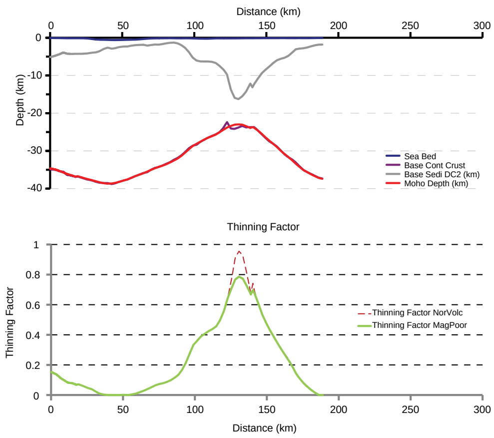


A1_SubW_CTi_40km_Va=7_BU_110Ma_D=2.85

Magma Poor



Gravity Inversion



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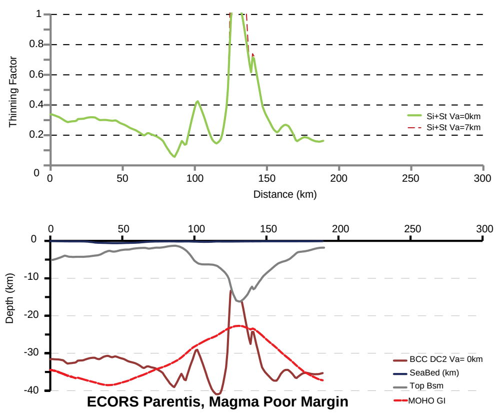
Subsidence Analysis Results

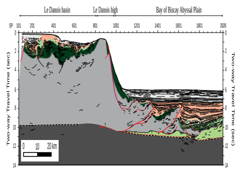


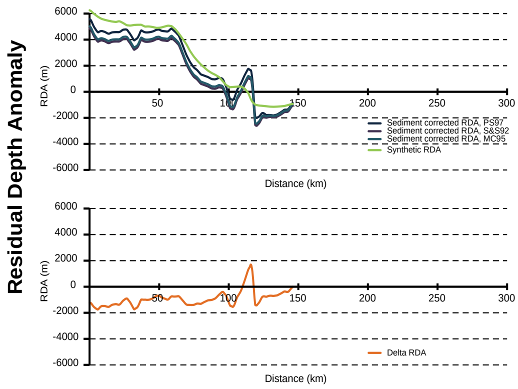
C2_SubW_CTi_40km_Va=0_BU_110Ma_D=2.85

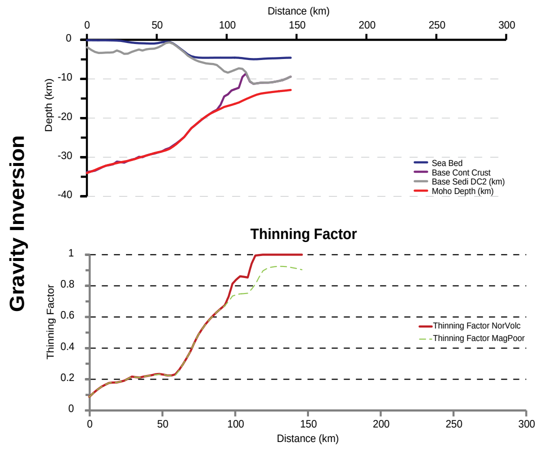
ECORS Parentis 2

SYNTHESIS OF RESULTS

Normal Volcanic Addition







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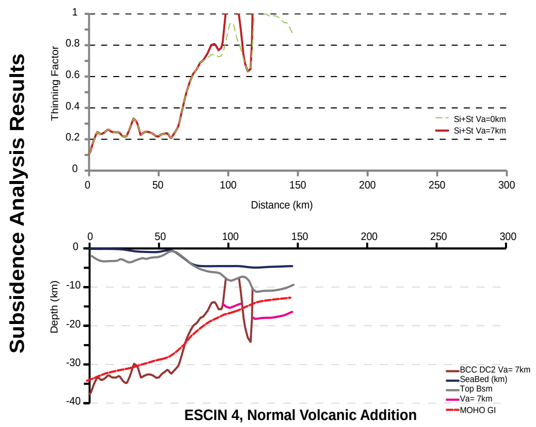


A1_SubW_CTi_40km_Va=7_BU_110Ma_D=2.85

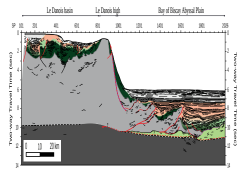
Gravity Inversion Parameters :

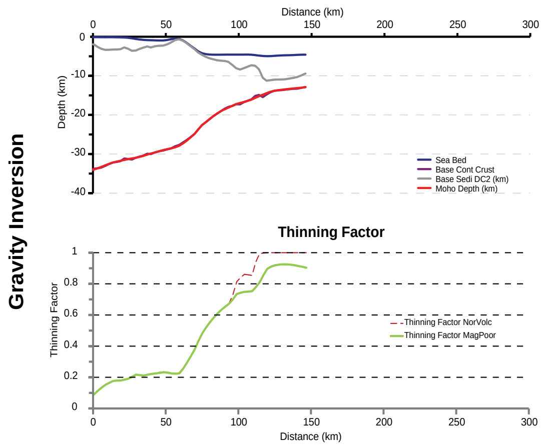
- Reference crustal Thickness (tcref): 40 km
- Initial Crustal Thickness = 35 km
- Crustal Density = 2.85
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr
- Sediment Thickness: *Onshore*, Serrano & al, 2006 ; *Offshore*, seismic profiles (IAM 12, ESCIN4, Norgasis 11-12-14--15-17-18-19-20-21-23, ECORS Bay of Biscay

Subsidence Widget Parameters :

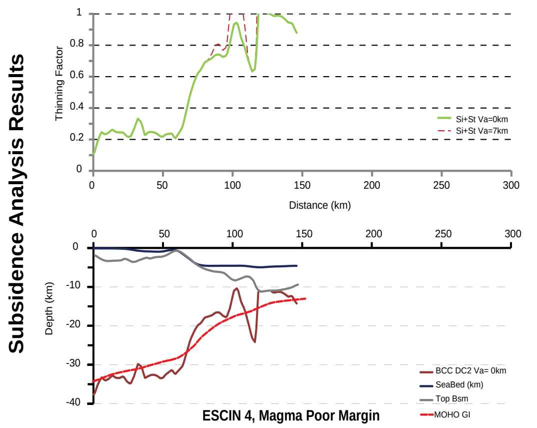
- Lithosphere Thickness = 125 km
- Initial Crustal Thickness = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333°C
- Basin Infill Density = 1.04
- Volcanic addition (Critical Gamma) = 0.7
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr

Magma Poor





F2_ctref40_inct35_MagPoor_g0.7_oc0_ra110_Soft_oldisoch110_Julie_Seds_DC2_Nav4

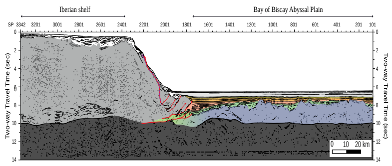


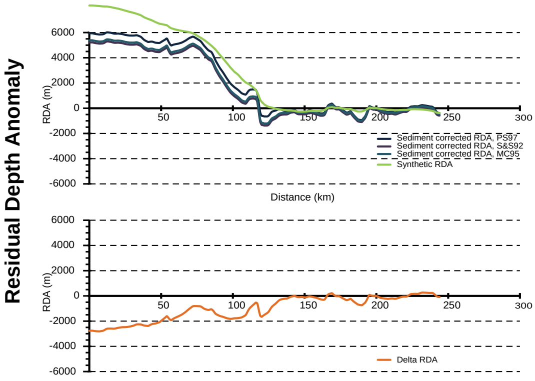
C2_SubW_CTi_40km_Va=0_BU_110Ma_D=2.85

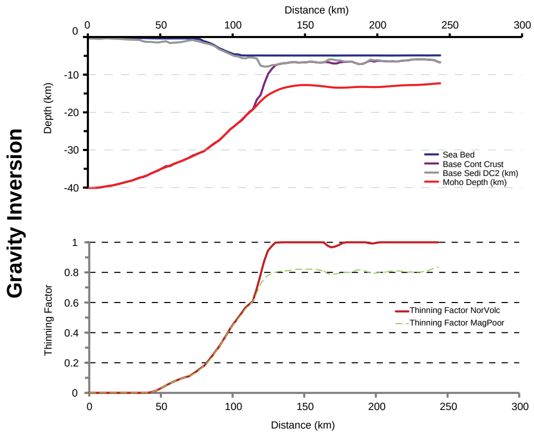
ESCIN 4

Synthesis of results

Normal Volcanic Addition







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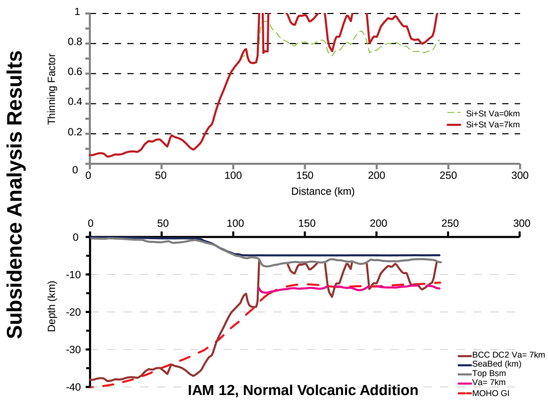


A1_SubW_CTi_40km_Va=7_BU_110Ma_D=2.85

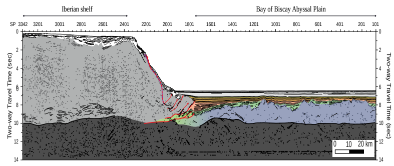
Gravity Inversion Parameters :

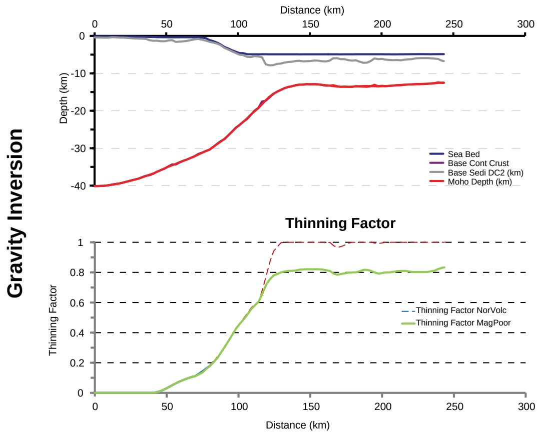
- Reference crustal Thickness (tcref): 40 km
- Initial Crustal Thickness = 35 km
- Crustal Density = 2.85
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr
- Sediment Thickness: *Onshore*, Serrano & al, 2006 ; *Offshore*, seismic profiles (IAM 12, ESCIN4, Norgasis 11-12-14--15-17-18-19-20-21-23, ECORS Bay of Biscay

Subsidence Widget Parameters :

- Lithosphere Thickness = 125 km
- Initial Crustal Thickness = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333°C
- Basin Infill Density = 1.04
- Volcanic addition (Critical Gamma) = 0.7
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr

Magma Poor





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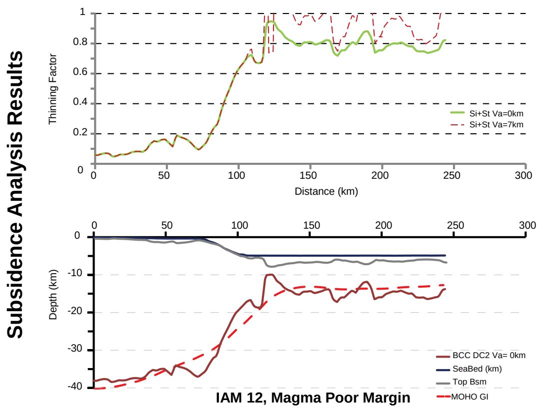
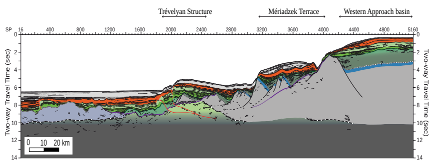


C2_SubW_CTi_40km_Va=0_BU_110Ma_D=2.85

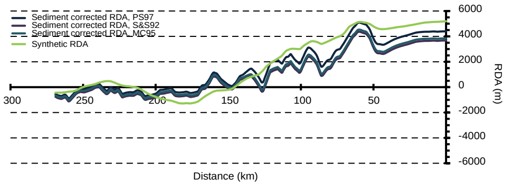
IAM 12

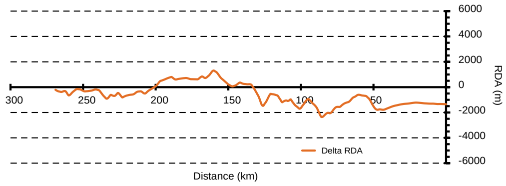
Synthesis of results

Normal Volcanic Addition

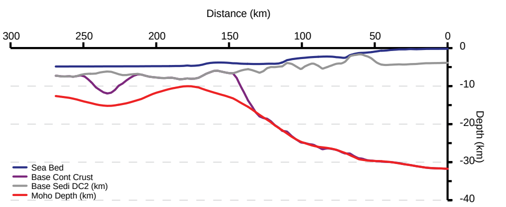


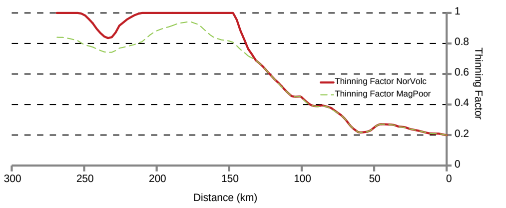
Residual Depth Anomaly





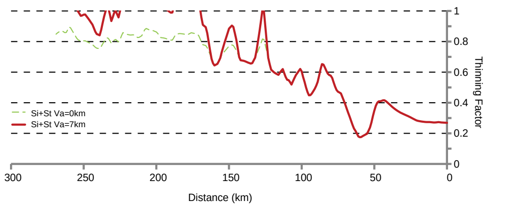
Gravity Inversion

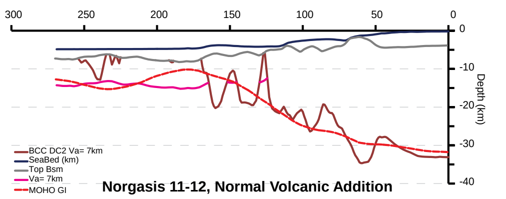




F1_ctref40_inct35_NorVol_g0.7_oc7_ra110_Soft_oldisoch110_Julie_Seds_DC2_Nav4

Subsidence Analysis Results





A1_SubW_CTi_40km_Va=7_BU_110Ma_D=2.85

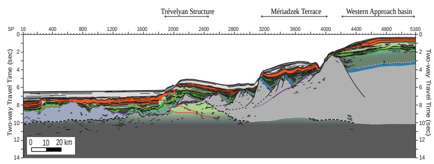
Gravity Inversion Parameters :

- Reference crustal Thickness (tcref): 40 km
- Initial Crustal Thickness = 35 km
- Crustal Density = 2.85
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr
- Sediment Thickness: *Onshore*, Serrano & al, 2006 ; *Offshore*, seismic profiles (IAM 12, ESCIN4, Norgasis 11-12-14--15-17-18-19-20-21-23, ECORS Bay of Biscay

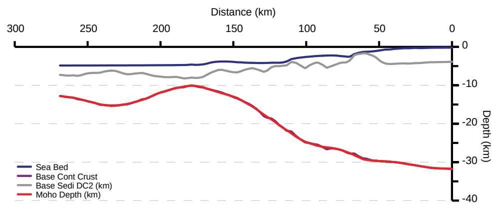
Subsidence Widget Parameters :

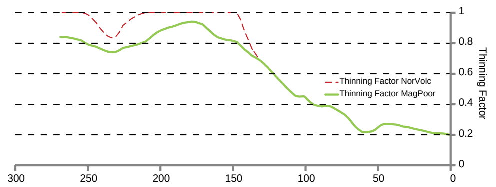
- Lithosphere Thickness = 125 km
- Initial Crustal Thickness = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333°C
- Basin Infill Density = 1.04
- Volcanic addition (Critical Gamma) = 0.7
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr

Magma Poor



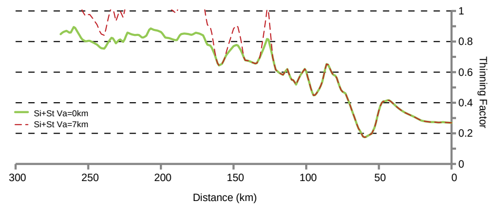
Gravity Inversion

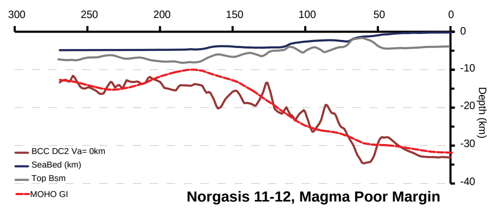




F2_ctref40_inct35_MagPoor_g0.7_oc0_ra110_Soft_oldisoch110_Julie_Seds_DC2_Nav4

Subsidence Analysis Results



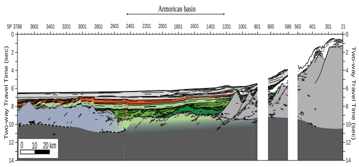


C2_SubW_CTi_40km_Va=0_BU_110Ma_D=2.85

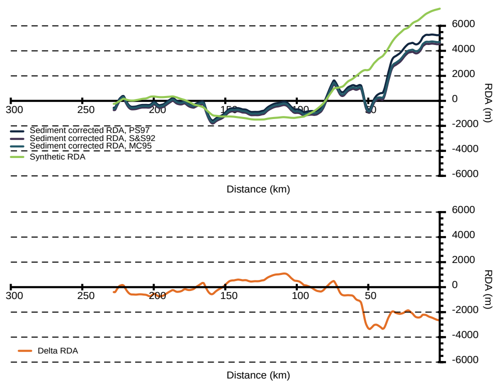
Norgasis 11-12

Synthesis of results

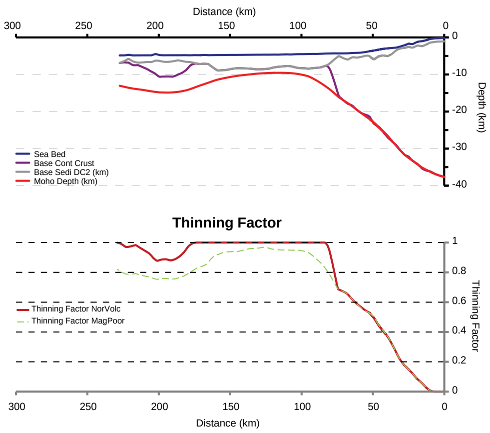
Normal Volcanic Addition



Residual Depth Anomaly

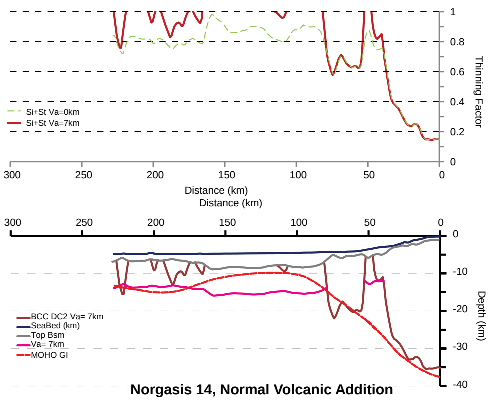


Gravity Inversion



F1_ctref40_inct35_NorVol_g0.7_oc7_ra110_Soft_oldisoch110_Julie_Seds_DC2_Nav4

Subsidence Analysis Results



A1_SubW_CTi_40km_Va=7_BU_110Ma_D=2.85

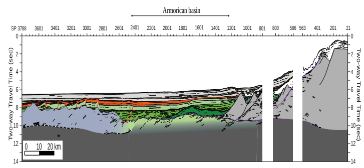
Gravity Inversion Parameters :

- Reference crustal Thickness (tcref): 40 km
- Initial Crustal Thickness = 35 km
- Crustal Density = 2.85
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr
- Sediment Thickness: *Onshore*, Serrano & al, 2006 ; *Offshore*, seismic profiles (IAM 12, ESCIN4, Norgasis 11-12-14--15-17-18-19-20-21-23, ECORS Bay of Biscay

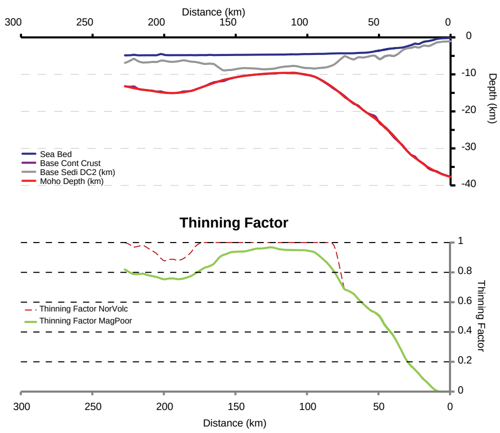
Subsidence Widget Parameters :

- Lithosphere Thickness = 125 km
- Initial Crustal Thickness = 40 km
- Crustal Density = 2.85
- Asthenosphere Temperature = 1333°C
- Basin Infill Density = 1.04
- Volcanic addition (Critical Gamma) = 0.7
- Volcanic Adddtion Thickness = 0 (MagPoor) & 7 km (NorVolc)
- Break Up Age =110 Myr

Magma Poor

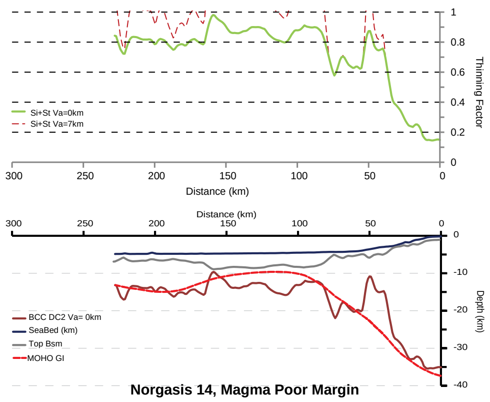


Gravity Inversion



F2_ctref40_inct35_MagPoor_g0.7_oc0_ra110_Soft_oldisoch110_Julie_Seds_DC2_Nav4

Subsidence Analysis Results



C2_SubW_CTi_40km_Va=0_BU_110Ma_D=2.85

Norgasis 14

Julie TUGEND

Rôle de l'hyper-extension pour la formation et la réactivation des systèmes de rift

Résumé

Les études couplant des observations provenant des marges passives actuelles et d'analogues fossiles ont permis de mieux appréhender les mécanismes d'extension de la lithosphère. Néanmoins, l'évolution spatiale et temporelle des processus de rupture continentale et de formation de croûte océanique reste mal contrainte. Le Golfe de Gascogne et les Pyrénées sont utilisés dans ce travail comme laboratoire naturel pour étudier la formation et la réactivation des systèmes de rift.

Le développement et l'application d'une approche terre-mer a permis d'identifier, caractériser et cartographier les domaines de rift formés lors de l'ouverture du Golfe de Gascogne et partiellement intégrés à l'orogène Pyrénéenne. Cette cartographie révèle l'architecture complexe de la limite de plaque Ibérie-Europe résultant d'une évolution fortement polyphasée. Plusieurs systèmes de rift spatialement distincts sont préservés à des stades d'évolution différents. Une segmentation importante partiellement héritée de la structuration pré-rift contrôle la formation des systèmes de rift ce qui a des implications pour la cinématique régionale.

Plusieurs étapes de la déformation compressive ont pu être distinguées et mises en relation avec l'architecture héritée du rift. La réactivation est initiée dans le domaine de manteau exhumé. Après la subduction de croûte hyper-amincie, la collision continentale est contrôlée par les domaines proximaux et de necking qui jouent le rôle de buttoirs.

Ces résultats soulignent l'interaction étroite entre l'héritage pré-rift et l'évolution spatiale des systèmes de rift ainsi que l'importance de l'architecture du rift pour comprendre la formation des orogènes.

Mots clefs: Amincissement crustal, rifting, réactivation, formation des orogènes, héritage, Pyrénées, Golfe de Gascogne

Abstract

Knowledge on lithosphere extensional mechanisms has greatly benefited from studies made both at present-day rifted margins and onshore fossil analogues. Nevertheless, the spatial and temporal evolution of the processes leading to continental break-up and oceanic crust formation remains poorly constrained. The Bay of Biscay and Pyrenees is used in this study as a natural laboratory to investigate the formation and reactivation of rift systems.

A new offshore-onshore approach is developed and applied to identify, characterize and map the rift domains inherited from the Bay of Biscay opening and partly integrated into the Pyrenean orogen. This mapping reveals the complex architecture of European-Iberian plate boundary resulting from a strongly polyphased evolution. Several rift systems spatially distinct are preserved at different evolutionary stages. An important segmentation partially inherited from the pre-rift structuration controls the formation of the rift systems, an observation that has important implications for regional kinematic restorations.

Several steps in compressional deformation can be distinguished and related to the rift inherited architecture. Reactivation is initiated in the exhumed mantle domain. Following the subduction of hyperthinned crust, continental collision processes are controlled by the proximal and necking domains acting as buttresses.

These results emphasize the role of pre-rift inheritance for the spatial evolution of rift systems and the importance of the rift-related architecture to unravel the formation of collisional orogen.

Keywords: Crustal thinning, rifting, reactivation, orogen formation, inheritance, Pyrenees, Bay of Biscay