Evolution of salt structures during extension and inversion of the Offshore Parentis Basin (Eastern Bay of Biscay)

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Abstract

The Late Jurassic-Cretaceous Parentis Basin (Eastern Bay of Biscay) illustrates a complex geological interplay between crustal tectonics and salt tectonics. Salt structures are mainly near the edges of the basin, where Jurassic-Lower Cretaceous overburden is thinner than in the basin centre and allowed salt anticlines and diapirs to form.

Salt diapirs and walls began to rise reactively during the Late Jurassic as the North Atlantic Ocean and the Bay of Biscay opened. Some salt-cored drape folds formed above basement faults from the Upper Jurassic to Albian. During Albian-Late

Cretaceous times, passive salt diapirs rose in chains of massive salt walls. Many salt diapirs stopped growing in the mid-Cretaceous when their source layer depleted. During the Pyrenean orogeny (Late Cretaceous-Cenozoic) the basin was mildly shortened. Salt structures absorbed almost all the shortening and were rejuvenated to form squeezed diapirs, salt glaciers, and probably subvertical welds, some of which were later reactivated as reverse faults. No new diapirs formed during the Pyrenean compression, and salt tectonics ended with the close of the Pyrenean orogeny in the Middle Miocene.

Using reprocessed industrial seismic surveys, we document how salt tectonics affected the structural evolution of this offshore basin largely unknown to the international audience.

Introduction

During the last decade or two, the important role of salt in facilitating shortening in sedimentary basins has been highlighted with examples around the world (for references, see reviews by Jackson, 1995, Letouzey et al., 1995, Rowan et al., 2004 and Hudec & Jackson, 2007). Many studies have focused on the salt-related effects of strong shortening, but much less work has addressed the subtler effects of mild shortening of salt structures. A subset of studies has focused on three types of salt tectonics: inversion controlled by basement faults (e.g., Stefanescu *et al.*, 2000; Stewart, 2007); thin-skinned salt structures over a detachment (e.g., Brun & Fort, 2004; Rowan *et al.*, 2004; Sherkati *et al.*, 2006); and salt-cored drape folds associated with extensional basement faults (e.g., Withjack & Callaway, 2000). The Parentis Basin has all three types of salt tectonics and is an excellent example of salt structures mildly affected by regional compression.

The Parentis Basin is in the northern foreland of the Pyrenean chain. In this basin salt structures were grew during Jurassic-Cretaceous rifting as the Bay of Biscay and the North Atlantic Ocean began to open and were subsequently inverted when the Iberian and Eurasian plates collided and Pyrenean shortening propagated northwards into the Parentis Basin in the late Cretaceous and Paleogene.

This basin is a western continuation of the onshore Aquitaine region (southwestern France) and widens westward into the Bay of Biscay (Figs. 1A & 1B). The eastern Bay of Biscay has two bathymetric domains separated by a bathymetric step (e.g. Sibuet *et al.* 2004a): a shallow part represented by the Landes shelf having a maximum depth of 200 m and the Landes Plateau between 200 and 1500 m deep (Fig. 2).

The Parentis Basin has been explored for petroleum for more than 50 years. Onshore exploration started in 1953 and in 1954 discovered the Parentis field (210.000.000 BBL of oil) (Biteau *et al.*, 2006). Offshore exploration drilling started in 1966 (Bourrouilh *et al.*, 1995; Le Vot *et al.*, 1996). The Parentis Basin has become a major oil-producing province in France with accumulations in carbonate reservoirs of Upper Jurassic to Lower Cretaceous sandstones. Most oil fields discovered are onshore (Parentis, Cazaux, Les Arbousiers, Les Pins, Lugos, Mothes, Lucats, Courbey, Tamaris and Les Mimosas), and no significant oil or gas has been reported offshore (Mascle *et al.*, 1994; Biteau *et al.*, 2006). After lack of success, offshore exploration decreased and some 3D-seismic surveys have been released since the 1990s in the shallow offshore. In contrast to the onshore and in the shallow offshore, the deep offshore (>200 m depth) has been little explored. 2D-seismic surveys have been carried out, but no exploration wells have been drilled in the deep offshore (Fig. 2).

Although salt tectonics was important in the evolution of the basin, no publications have addressed this role in the entire offshore Parentis Basin. Several studies have covered the structural evolution of the basin (Mathieu, 1986; Mascle *et al.*, 1994; Biteau *et al.*, 2006); others have focused only on some salt structures in the eastern part of the basin (Curnelle & Marco, 1983; Mariaud, 1987; Mediavilla, 1987). In addition, these studies concentrated in the French part of the basin, neglecting the Spanish part against the North Iberian margin.

Using conventional recently reprocessed 2D seismic surveys and well data from the Parentis Basin, we present new interpretations of the timing of the main tectonic domains and the style and evolution of salt structures, focusing on extension of the continental margins and the subsequent Pyrenean inversion.

Geological and structural setting

The Bay of Biscay is an E–W oriented embayment of the Atlantic Ocean between the Iberian Peninsula and the western coast of France (Fig. 1A & 1B). The bay opened between late Barremian and Santonian times (e.g. Montadert *et al.*, 1979; Le Pichon & Barbier, 1987; García-Mondejar, 1996; Vergés & García-Senz, 2001; Sibuet *et al.*, 2004b). The Bay of Biscay has two distinct domains. The western domain is an abyssal plain 4–5 km deep underlain by transitional to oceanic crust (Gallastegui *et al.*, 2002; Thinon *et al.*, 2003; Sibuet *et al.*, 2004a, 2004b; Pedreira, 2004; Ruiz, 2007). The eastern domain is a shallow shelf and an intermediate plateau overlying continental crust 15–25 km thick (Pinet *et al.*, 1987; Ruiz, 2007). Both domains are bounded to the north-east by the Armorican Margin (Figs. 1B & 3), which is a Mesozoic passive margin having southwest-verging normal listric faults (Montadert *et al.*, 1979; Deregnaucourt & Boillot, 1982; Le Pichon & Barbier, 1987; Thinon *et al.*, 2003). The southern boundary of the Bay of Biscay is the North Iberian Margin (Figs. 1B & 3), a north-verging basement-involved thrust system of Late Cretaceous-Cenozoic age that overthrust Paleozoic to Cenozoic sediments of the Bay of Biscay abyssal plain (e.g. Sibuet *et al.*, 1971; Boillot, 1986; Álvarez-Marrón *et al.*, 1996; Gallastegui *et al.*, 2002; Ayarza *et al.*, 2004, Pedreira, 2004) (Fig. 1C). This thrust system is the northern front of the Pyrenean orogen, which separates the Iberian and the Eurasian plates (Fig. 1B & 1C). Adjoining this front is the North Pyrenean foreland basin which preserves the Mesozoic Parentis Basin in the eastern Bay of Biscay (Figs. 1C).

The Parentis Basin includes four bathymetric domains: the Landes shelf, the Armorican shelf, the Basque shelf and the deeper Landes Plateau enclosed by them (Fig. 2). The southern Parentis Basin is bounded by the major north-dipping Ibis and Landes faults, visible in the ECORS-Bay of Biscay and MARCONI-3 profiles (Figs. 3, 4A & 4B). In the Ibis area (Fig. 4A), the basin is only 40 km wide, but the extensional faults have up to 2 s two-way time (about 3400 m) throw (Bois et al., 1997). However, farther west (Fig. 4B), the basin widens to 70 km, and the extensional faults have less throw.

To the south of the main faults, the Landes High (Fig. 1B) is part of a plateau uplifted and eroded with an uppermost Cretaceous–Cenozoic thick sedimentary succession that unconformably overlies Hercynian basement or a thin and partially eroded Triassic–Jurassic cover (Gariel *et al.*, 1997) (Fig. 1C). To the north, the basin is limited by the Celt-Aquitaine Flexure, a hinge line (Fig. 3). South of this hinge the pre-Triassic basement is deepest and the Mesozoic and Cenozoic fill is thickest.

The Parentis Basin contains Mesozoic and Cenozoic strata nearly 15 km thick (Dardel & Rosset, 1971; Mathieu, 1986; Bourrouilh *et al.*, 1995; Bois & Gariel, 1994; Bois *et al.*, 1997) (Fig. 5). East-striking normal faults offset the Mesozoic succession

(Masse, 1997), and diapirs of upper Triassic evaporites pierce the basin fill (Curnelle & Marco, 1983; Mathieu, 1986; Mediavilla, 1987; Ferrer *et al.*, 2008a), including the Upper Cretaceous to Cenozoic synorogenic deposits (Curnelle & Marco, 1983; Bois *et al.*, 1997; Masse, 1997; Ferrer *et al.* 2008a) (Fig. 4A).

The Moho shallows northwards from about 30-35 km beneath the Basque shelf to 18-22 km below the Cap Ferret Canyon along the E-W boundary between the Landes Plateau and the Armorican shelf (Fig. 1C). North of this boundary, the Moho abruptly deepens to 30-36 km beneath the Armorican shelf (Roberts & Montadert, 1980; Tomassino & Marillier, 1997; Thinon *et al.*, 2003). The crustal thickness decreases westwards from 7 km in the ECORS Bay of Biscay (Pinet *et al.*, 1987; Tomassino & Marillier, 1997) (Fig. 4A) to 6-5 km in the MARCONI-3 profile (Gallart et al., 2004; Ruiz, 2007) (Figs. 1C & 4B).

The formation and evolution of the Parentis Basin was controlled by the relative motion of the Iberian and Eurasian plates as the North Atlantic Ocean opened. Two rifting episodes (Permian-Triassic and latest Jurassic-Lower Cretaceous) related to the break-up of Pangea formed a transtensional to extensional plate boundary between Iberia and Eurasia (Srivastava *et al.*, 1990). The main Mesozoic sub-basins north of the Pyrenees formed, among them the Parentis Basin (Curnelle *et al.*, 1982; Bourrouilh *et al.*, 1995; Biteau et al., 2006). The main depocentre of the Parentis Basin formed from the Barremian to middle Albian (Fig. 4A). From late Santonian, faster opening of the South Atlantic Ocean and increased northward translation of the African plate caused the Iberian and Eurasian plates to converge then collide (Ziegler, 1988; Rosenbaum *et al.*, 2002) and partially closed the Bay of Biscay. Although Pyrenean collision inverted the most important Mesozoic Pyrenean extensional basins along the Iberian-European plates boundary (e.g. Basque-Cantabrian, Lacq-Mauleon or Organyà basins)

(Choukroune & ECORS Team, 1989; Roure *et al.*, 1989; Muñoz, 1992, Álvarez-Marrón *et al.*, 1996; Bourrouilh *et al.*, 1995), the Parentis Basin was only slightly inverted (Mathieu, 1986; Pinet *et al.*, 1987; Bois & ECORS Scientific Party, 1990; Verges & García-Senz, 2001), probably because the Landes High blocked northward propagation of Pyrenean shortening (Fig. 1C) (Ferrer *et al.*, 2008a). This buttressing may have been induced by stronger or thicker crust below the Landes High.

Dataset and methodology

The seismic dataset comprised 12 reprocessed 2D conventional seismic lines shot between 1974 and 1990 and two reprocessed deep seismic surveys (ECORS and MARCONI), covering almost all the Eastern Bay of Biscay, except part of the northern Parentis Basin, the Armorican extensional margin and the deep offshore (Fig. 2). Recent reprocessing of conventional seismic data has significantly improved imaging, especially of allochthonous salt. However, the top and the base of the autochthonous salt are still poorly imaged in many seismic profiles and in places interpretation is merely speculative.

Twenty three wells constrain seismic interpretation in the eastern Parentis Basin (Fig. 2). Because of water depths of >1000 m, no wells have been drilled in the western part of the basin, so surveys there have been correlated from wells in the east.

Salt-related structures

The offshore Parentis Basin exhibits a wide array of salt-related structures. Triassic Keuper salt or its welded equivalent underlies the entire Parentis Basin within the study area. The ECORS-Bay of Biscay profile illustrates the structural style of the eastern Parentis Basin domain (Fig. 4A). Here the basin is asymmetrical and its deepest part overlies the thinnest crust. This thinning was controlled by the Ibis Fault whose slip tilted the hanging wall to the south. Salt walls or salt anticlines cluster along the edges of the Parentis Basin (Fig. 3), where the overburden is thinnest. The viscous salt layer accommodated an extensional drape (forced) fold above the Ibis Fault.

Conversely, the basin geometry is different in the west, as shown by the MARCONI-3 seismic profile (Fig. 4B) 52 km farther west. Instead of merely pinching out, the basin ends abruptly with a half-graben controlled by a major north-dipping fault (Landes Fault). The half-graben is filled by thick Jurassic-Upper Cretaceous carbonates deformed by ENE-trending salt-cored anticlines and squeezed salt walls near the master fault (Ferrer *et al.*, 2008a). In contrast to those farther east, salt-cored anticlines have longer wavelength (up to 10 km), amplitude and lateral continuity (between 15 and 20 km).

Eastern domain

In the eastern domain the deepest part of the basin is the Parentis Trough (nearly 12 km deep) (dash blue line in Fig. 3), a syncline trending east for more than 30 km. The trough results from extension on the Ibis Fault between the Neocomian and the early Albian and from deformation of the Keuper evaporites and its overburden. This trough continues eastwards in the onshore Parentis Basin (Fig. 3). The thickness of Lower Cretaceous sediments reaches more than 3400 m in the hanging wall of the Ibis Fault, where the fault has a throw of about 2500 m.

Salt structures of contrasting geometry formed near the northern and southern basin margins. In the north, gentle salt-cored anticlines trending WNW deform all the Mesozoic sequences, which were mildly flexed across the northern basin-margin hinge (Céphée-Aldeberan Ridge) (Figs. 3 & 6). Conversely in the southern margin, salt structures are more complex and include ENE-trending salt anticlines (Eridan-Antares Ridge), diapiric salt walls, isolated teardrop diapirs, stocks and allochthonous salt sheets (Figs. 3, 6, 7 & 8). These salt structures affect Mesozoic and Cenozoic strata.

Along the southern margin of the Parentis Trough, the Ibis/Eridan-Antares ridges comprise salt-cored anticlines along the Ibis Fault and deform only Mesozoic strata. These ridges stretch 55 km and trend WNW near the French coast to ENE farther west (Fig. 3). Folded strata include a thick Jurassic unit, a Lower Cretaceous unit that thickens in the Parentis Trough, and an Albian-Upper Cretaceous unit that thickens markedly north of the anticline (Fig. 6). These last syn-folding deposits are cut by north-verging break thrusts in the northern limbs of the salt-cored anticlines. The structural complexity of the Eridan-Antarés anticline increases towards the east, where its north flank is highly faulted. In the southern limb of the anticline the Jurassic-Cretaceous transition is marked by southward thinning and the onlapping of Lower Cretaceous strata. Farther south, a strongly erosive regional unconformity marks this transition.

South of the Ibis/Eridan-Antares ridges, the apparent narrowness and inverted tear drop shape of ENE-trending salt ridges, such as Alcyon and Puffin, suggest that the diapirs were squeezed or even welded (Fig. 6). The bulbs overlie what we interpret as subvertical secondary salt welds, formed by pinching off of the diapir stems during lateral compression.

Lateral cutoffs of adjoining strata suggest that these salt walls grew as passive diapirs from Jurassic to early Miocene times. These diapirs are associated with minor north-dipping normal faults displacing the Jurassic and Lower Cretaceous series and apparently detached in the Upper Triassic evaporites. No diapirs reach the present sea floor.

One of these salt walls extruded to form the large Pelican Salt Sheet (Figs. 3 & 7) in the central part of the basin. Stratal cutoffs against the base of the sheet suggest ~20 km of advance of salt, probably by extruding as a glacier over the sea floor (Fletcher *et al.*, 1995; Hudec & Jackson, 2006). Cutoffs show that the Pelican Salt Sheet began to spread at the end of Albian time and extruded farthest in Paleocene-Eocene times during the Pyrenean orogeny. Regional shortening typically squeezes and expels salt from diapirs to form a vigorous salt extrusion at the surface (Jackson & Cramez, 1989; Letouzey & Sherkati, 2004; Hudec & Jackson, 2006; Callot *et al.*, 2007). The Ibis-2 well drilled nearly 250 m of Triassic salt in this salt-sheet between Eocene strata below and Senonian strata above. These Senonian rocks have been interpreted as a dismembered roof fragment carried during salt extrusion (Curnelle & Marco, 1983). Sporadic normal faults above the salt-sheet suggest accommodation of the overburden as salt was redistributed within the buried allochthonous sheet. The thin sedimentary roof of this structure was stretched by flow of underlying salt. Gentle folds in the roof of the Pelican structure indicate continued flow within the salt sheet caused by differential loading of the overburden or by Pyrenean shortening (Fig. 7). Well data indicate that this salt-sheet stopped advancing and was buried during the Oligocene, probably because its feeder stem finally pinched shut during the Pyrenean shortening.

Western domain

Unlike the eastern domain, the western domain of the Parentis Basin (Fig. 3 & 4B), widens and is filled by thicker (5000–8500 m) Jurassic-Lower Cretaceous sequences overlain by an uppermost Cretaceous-Cenozoic package (Ferrer *et al.*, 2008a). The southern boundary of the basin is the major Landes Fault (Fig. 4B), which lies southwest of the Ibis Fault (Ferrer *et al.*, 2008a) (Fig. 3). The footwall of the Landes

Fault tilts south, whereas the hanging wall is either broadly anticlinal (Fig. 4B) or near horizontal (Fig. 9). The dominant structures in this domain, which trend ENE, are the Izurde, Marratxo and Txipiroi anticlines defined by Jurassic-Lower Cretaceous overburden and cored by Upper Triassic evaporites (Fig. 9). These folds have a greater wavelength (up to 10 km), amplitude and lateral continuity (between 15 and 20 km) than do the eastern basin ridges. The Txipiroi salt-anticline overlies a gentle basement antiform (Fig. 4B). In contrast the Marratxo and Izurde salt anticlines formed where the basement was virtually flat (Fig. 9). Despite the different basement structure, these structures share the following features.

- The Jurassic-Aptian successions were extended by north-dipping normal faults detached in the Upper Triassic evaporites. These faults are located mainly in the northern flank of the anticlines, especially near the synclinal hinges (Fig. 9).
- In the southern flanks of these structures a strong erosional unconformity across the Lower and Upper Cretaceous successions is overlain by uppermost Cretaceous-Paleocene strata.
- The uppermost Cretaceous-Cenozoic units thin towards the crest of salt anticlines on both sides, suggesting that the fold limbs rotated during Pyrenean shortening.
- Salt anticlines are separated by autochthonous salt thinned to the point of welding (Fig. 9). Some of these salt anticlines may have produced salt diapirs, as in the crest of the Izurde Ridge (Fig. 9).

In the western domain diapirs are close to the Landes Fault (Figs. 9 & 10). Their bulbs have an inverted-teardrop shape suggesting a secondary near-vertical weld formed by closing of the diapir stem (Fig. 10). Although mechanically feasible, the presence of welded feeders is speculative because of poor data quality. These diapirs cut most of the folded Albian and Lower Cretaceous reflectors but only some uppermost Albian and Upper Cretaceous reflectors. Although seismic resolution is poor, some diapiric pedestals are interpreted at the base of the secondary welds.

Study of the Ibis and Landes faults is outside the scope of this paper and both structures have poor seismic coverage (Fig. 2). Nevertheless we provisionally interpret the Ibis and Landes faults as a large relay structure in which the Triassic salt and its diapirs are confined to the hanging wall of the Landes Fault but extend eastwards up the relay ramp into the footwall of the Ibis Fault (Fig. 3).

Initiation and growth of salt diapirs

Due to poor seismic resolution at depth, diapiric initiation is poorly documented. However, in some cases cutoffs of flanking reflectors suggest that salt diapirs and walls began to rise during the Late Jurassic. These diapirs could have been initiated reactively by mild extension of the carbonate platform during post-rift thermal subsidence (Fig. 11). In the most cases, the actual normal faults that could have initiated reactive diapirism are not clearly imaged, and so our interpretation is conjectural. Nonetheless, a large reactive diapir is preserved in the western Parentis domain between Euskal Balea and Izurde ridges (Fig. 9). Differential thicknesses of growth strata on its flanks suggest that it began rising in the Jurassic. Abundant Late Jurassic-Early Cretaceous normal faults between diapirs support an extensional setting (Fig. 9).

Some diapirs could have been triggered by differential sedimentary loading because the Upper Jurassic sequence varies slightly in thickness (Figs. 6–8). More severe tectonic instability is suggested by the major unconformity at the Jurassic-Cretaceous transition along both edges of the eastern Parentis domain (Mathieu, 1986; Mediavialla, 1987; Masse, 1997; Biteau *et al.*, 2006). Farther south, in the Alcyon or

Puffin zone, this erosion exposed Middle Jurassic strata (Figs. 6, 8, 12b & 12c). This erosive unconformity has been linked to the growth of salt anticlines during Triassic-Liassic deformation (Curnelle & Cabanis, 1989) but could also record uplift of a rift shoulder as the Bay of Biscay began to open in the Cretaceous. This rifting could have initiated reactive diapirism.

As subsidence peaked, a 5-km-thick Barremian to Albian sequence filled the Parentis Trough (Brunet, 1991) (Figs. 12b, 12c & 12d). Sediment loading in the Parentis Trough expelled Keuper evaporites towards the edges of the basin, where salt-cored anticlines formed (Eridan-Antares-Ibis in the southern and Céphée-Castor in the northern area) (Figs 11, 12c & 12d). Because Jurassic-Lower Cretaceous aggradation was faster in the west, salt diapirs there tended not to reach the surface or extrude. Deep diapirs grew in the footwall of extensional faults, forming broad but low diapirs. Exceptions were the relatively tall diapirs, Euskal Balea and Izurde. To the south the Basque-Cantabrian shelf remained high during the Lower Cretaceous (Mathieu, 1986; Bois *et al*., 1997). Some reefs grew above the salt-cored anticlines (e.g., Antares) of the northern Parentis edge during the mid-upper Albian (Mathieu, 1986; Biteau *et al*., 2006). The distribution of these carbonate buildups in the shelf margin was probably controlled by the interaction of eustasy, salt tectonics and paleogeography, as in the Mesozoic La Popa Basin (Mexico) and the modern Persian Gulf (Purser, 1973; Giles & Lawton, 2002).

During the Albian-Late Cretaceous, salt diapirs rose in chains of massive salt walls (Figs 12d & 12e). Once they pierced reactively to the surface, the salt walls would have continued to grow as passive diapirs (Fig. 11). Most diapirs have thickened peripheral sinks of Albian-Upper Cretaceous age around them (e.g., Puffin and Alcyon; Fig. 9, 12d & 12e). But generally these local responses to diapiric rise were masked by large changes in regional thickness controlled by crustal tectonics. As the source layer depleted and the salt was partially welded, many salt diapirs stopped growing in the mid-Cretaceous.

Effect of basement structures on early salt tectonics

North-dipping faults (Ibis and Landes faults) offsetting the basement with throws of more than 3 km are recognizable in the southern boundary of the basin. However, smaller faults are largely masked by the velocity effects of salt. Also, Pyrenean inversion may have removed early extensional offset on smaller faults. Despite poor imaging, the huge scale of faulting means that basement tilting and fault offsets evidently controlled the initial thickness of salt and overburden and hence affected the style of salt tectonics. Contrasting effects are shown by the Ibis Fault and the Landes Fault.

Ibis Fault

The most important structure of the eastern Parentis Basin is the Ibis Fault, which is overlain by the Eridan-Antares anticline. In the southern limb of the anticline the Jurassic-Cretaceous transition is marked by the southwards thinning and the onlap of Lower Cretaceous strata. These features suggest that the Ibis footwall was already elevated and tilted northward at the start of the Cretaceous (Fig. 6 shows the present-day subhorizontal attitude after Paleogene rotation of the basement during inversion). Farther south near the edge of the basin, this thinning changes to a strongly erosive regional unconformity. North of the Eridan-Antares anticline, thickening of the Neocomian to lower Aptian sequences suggests that the Parentis Trough began its main subsidence in Neocomian times.

Although the basement is poorly imaged, the Ibis Fault is interpreted as a half graben because its overburden forms a monocline verging northward (Figs. 4A & 6). A salt-cored anticline (Eridan-Antarés anticline) drapes over the northward-dipping master fault of the half graben. Generic physical modeling (e.g. Withjack and Callaway, 2000; Ferrer et al., 2008b) suggests that before Pyrenean shortening, this anticline was probably a monoclinal extensional drape fold responding to slip on the underlying Ibis Fault. The viscous salt layer partially decoupled the major sub-salt fault slip from the draping cover. Physical modeling (Jackson et al., 1994; Vendeville et al., 1995; Withjack and Callaway, 2000) suggests that in such systems, the displacement and displacement rate of the sub-salt normal fault and the thickness and strength of the salt layer and its overburden controlled the structural style of the cover sequence. Thrust faults similar to those in the northern limb of the fold also form in physical models of drape monoclines above a rapidly slipping normal fault in the basement (Withjack and Callaway, 2000).

A massive upper Aptian-Albian depocentre on the north flank of the Eridan-Antarés anticline onlaps markedly onto the Céphée-Aldebaran anticline to its north (Figs. 4A, 6 & 12d). The main sedimentary load of the Parentis Trough was thus confined between the two salt-cored anticlines. This differential load in the trough would have expelled salt laterally to the north and south and possibly to the east, where overburden was thinner. Salt expulsion would have allowed the depocenter to subside and eventually weld the autochthonous salt beneath the base of the trough, as inferred in Figures 4A and 6.

Landes Fault

The Landes Fault forms a half-graben at the southern edge of the Triassic salt basin and its overburden 5000-8500 m thick. Basement dips southwards in the hanging wall and farther north is near-horizontal or gently northward dipping (Fig. 4B). This change in hanging-wall dip strongly influenced the location and evolution of the Txipiroi salt-cored anticline (Ferrer *et al.*, 2008a) at the edge of the half-graben. Uppermost Cretaceous-Cenozoic strata rest on an angular unconformity, especially above extensional tilted blocks. Erosion truncated more than half the Lower Cretaceous sediment thickness below the severest parts of the unconformity (Fig. 9). Fault geometry suggests that the Txipiroi anticline grew from the migration and later accumulation of Keuper salt (Ferrer *et al.*, 2008a & Ferrer et al., 2008b).

As extension, basement tilting and sedimentation progressed during the Jurassic and Early Cretaceous, increased loading of the Landes half-graben expelled salt into early-formed diapirs such as Euskal Balea and Izurde and laterally to the north up the tilted hanging wall of the Landes Fault (Fig. 9). Lateral salt migration produced salt-cored anticlines where the dip of the basement becomes near-horizontal, forming the Txipiroi and Izurde ridges (Ferrer *et al.*, 2008b). Salt migration ended when the source layer welded. Examples of similar structural style have been described in the Jeanne d'Arc basin (Withjack & Callaway, 2000 & Ferrer et al., 2010).

Pyrenean shortening of salt structures

The Pyrenean orogeny began during the Late Cretaceous and continued to the Eocene. As a result the central part of the Parentis Basin (Parentis Trough) was uplifted and inverted (Fig. 12e). The top of Cretaceous was strongly eroded, and the Paleocene unit is higher than south of the Ibis Fault (Figs. 6 & 12f). The Parentis Basin as a whole was only mildly shortened by this orogeny compared with the Pyrenean hinterland where the minimum total shortening is about 165 km (Beaumont *et al.*, 2000 & Muñoz, 2002).

The Landes High in the Parentis Basin appears to have provided a buttress against the Pyrenean compression, which protected the Parentis Basin from major inversion even though the underlying crust was severely thinned in the Mesozoic (Ferrer *et al.*, 2008a). The lack of significant inversion structures in the Parentis Basin is the main indication that the Landes High shielded the Parentis Basin from Pyrenean shortening from Late Cretaceous to Early Miocene time. Shielding explains why basement-involved Pyrenean shortening is only present south of the Landes High and concentrated along the northern margin of the Basque-Cantabrian Basin (Fig. 1C). The Landes High acted as a buttress probably because it had a stronger crust than in the adjoining Basque-Cantabrian and Parentis basins. The two basins were strongly extended in the Early Cretaceous, so that thin, warm crust was overlain by thick Mesozoic cover. By contrast in the Landes High, where Mesozoic extension was much less, the crust was thicker and colder and overlain by thin cover and thus significantly stronger and able to resist Pyrenean compression along the Iberian-Eurasia collision boundary.

The Pyrenean inversion of the Parentis Basin drastically deformed salt structures formed earlier by Triassic-Early Cretaceous extension and halokinesis. Most salt structures responded readily to shortening by rising and narrowing because they were weak and their ENE trend was favorably oriented to the north-directed Pyrenean compression (Figs. 11 & 12f). Salt structures responded in the following ways. (1) Previously buried and dormant salt walls near the southern boundary of the basin were rejuvenated by squeezing as the salt within them was displaced upwards and arched their previously flat-lying roofs (Figs. 8, 10, 11 & 12f). (2) Squeezed salt walls extruded so much that salt advanced glacially, beginning in the lattermost Cretaceous and accelerating during Eocene-early Miocene times (Figs. 7 & 11). (3) Large anticlines rose above regional during shortening as the salt-cored Ibis and Eridan-Antarés ridges above the Ibis Fault and the Txipiroi anticline near the Landes Fault (Figs. 4A & 4B). (4) In the most extreme regional shortening, both extruding and buried salt walls could have pinched off to form subvertical secondary salt welds (Figs. 10, 11, 12f & 12g). If diapirs welded shut, any further shortening could have been by inversion of pre-existing normal faults, by reverse fault welds or by short-cut thrusts that nucleated in the salt pedestals during the last stages of the Pyrenean contraction (Fig. 8). No new diapirs formed during the inversion because the source layer was largely depleted in salt. Salt diapirism ended with the close of the Pyrenean orogeny in the middle Miocene (Figs. 9, 10, 11 & 12g).

Pinching off of some diapirs is conjectural but is suggested by these three lines of evidence, each of which is disputable. (1) Pinch-off is mechanically likely as the culmination of shortening in a wide array of geological settings (e.g., Vendeville & Nilsen, 1995; Cramez & Jackson, 2000; Brun & Fort, 2004; Gottschalk *et al.*, 2004; Rowan *et al*. 2004; Roca *et al.*, 2006; Rowan & Vendeville, 2006; Sherkati *et al.*, 2006; Jackson *et al.*, 2008; Dooley *et al.*, 2009). However, shortening may not have been sufficient for complete pinch off. (2) In the western Parentis Basin, where the quality of seismic data is better, the strong reflector of the top of autochthonous salt curves up to a cuspate point below the crest of the salt wall. However, this geometry could be a velocity pull-up. (3) The roof of the diapirs is arched above the regional in some places, as for Puffin Diapir in Figure 6. Where one side of the diapir is uplifted higher than the other side, this is compatible with a thrust-weld resulting from closure of an inclined stem. Again, though the asymmetrical arching could reflect another cause: namely an inclined stream rising up the inclined stem of a diapir.

Conclusions

The evolution of the Parentis Basin during extensional and subsequent contractional crustal deformation was strongly influenced by salt tectonics. Two master faults (Ibis and Landes faults) separated two structural domains having a wide array of salt-related structures. These faults may form a large relay structure. Salt stocks and walls grew near the south edge of the salt basin in the hanging wall of the Landes Fault in the western domain and in the footwall of the Ibis Fault in the eastern. In the east, gentle drape anticlines cored by salt formed in the edges of the basin (Eridan-Antares-Ibis and Céphée-Castor ridges). In the west, larger salt-cored anticlines formed in the crest of basement antiforms (e.g. Txipiroi salt-anticline) or above flat basement (e.g. Marratxo and Izurde salt-anticlines).

Salt walls and diapirs rose from a Triassic source layer deposited during Pangea rifting. These structures may have been initiated in extension as reactive diapirs during early opening of the Bay of Biscay or by differential sedimentary loading as early as the Upper Jurassic. The thick sedimentary fill deposited in the Parentis Trough since the Barremian to Albian expelled Keuper evaporites towards the edges of the basin where salt-cored anticlines formed. The Eridan-Antares-Ibis anticline grew as a drape fold above the Ibis master fault in response to its slip. The Céphée-Castor salt-cored anticline grew in the northern edge.

Salt diapirs evolved to a passive mode during Albian times. Many of these walls stopped growing by the middle of the Late Cretaceous, but some continued to grow passively until the Miocene.

As Iberia and Eurasia collided and drove the Pyrenean orogeny in the Late Cretaceous, the Parentis Basin was mildly shortened. Pre-existing salt-cored anticlines were amplified so that their crests were uplifted and eroded. Because of their weakness and preferred orientation, most salt structures responded readily to compression and absorbed most of the Pyrenean shortening. Diapirs, some of which were dormant and buried, were rejuvenated by regional compression as their stems may have pinched off. Upward expulsion of salt displaced from the squeezed walls arched up their sedimentary roofs to form shallow anticlines. Locally the pre-shortening salt walls expelled salt which advanced as glaciers over the sea floor for up to 20 km. This salt sheet stopped advancing and was buried during the Oligocene. No new diapirs formed as the Parentis Basin was inverted because the autochthonous source layer was largely depleted.

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FIGURE CAPTIONS

Fig. 1. (A) Location map of the Pyrenees and the Bay of Biscay (modified from Muñoz, 2002). (B) Simplified tectonic map of the Pyrenees and adjoining basins (modified from Ferrer et al., 2008a). Location of Figure 3 is shown by the dashed rectangle. (C) Upper-crustal transect through the eastern Bay of Biscay and adjoining northern part of the Basque Pyrenees (modified from Ferrer et al., 2008a). Location is shown in (B).

Fig 2. Bathymetric map of the eastern Bay of Biscay (modified from Sibuet et al., 2004a) showing the seismic and well data used in this work. For clarity there are only included the names of the wells referenced in the manuscript: 1) Aldebaran, 2) Eridan, 3) Pingouin, 4) Ibis-2b and 5) Pelican.

Fig. 3. Simplified Cenozoic subcrop map of the offshore Parentis Basin showing the main salt structures and faults. Thick black lines and numbers show location of seismic profiles and line drawings in Figs. 4A, 4B, 6, 7, 8, 9 and 10.

Fig. 4. Line drawings of interpreted deep seismic lines through the Parentis Basin. (A) ECORS-Bay of Biscay seismic profile (modified from Pinet et al., 1987) and (B) MARCONI-3 seismic profile (modified from Ferrer et al. 2008a). Deep crustal structure is not to scale. See Fig. 2 for location.

Fig. 5. Tectonostratigraphic chart of the Parentis Basin showing ages, lithologies, tectonic events, and horizons interpreted in seismic lines (partially compiled from Mathieu, 1986; Mediavilla, 1987; Bourrouilh et al., 1995; Le Vot et al., 1996 and Biteau et al. 2006).

Fig. 6. (A) Seismic section and **(B)** and interpreted line drawing of a composite 2D seismic profile in the eastern Parentis Basin showing four salt structures (Puffin and Alcyon diapirs and Eridan-Antares and Céphée-Aldebaran ridges). Note the strong erosional unconformity in the top of Cretaceous in the Eridan-Antares Ridge and the lensoid Lower Cretaceous units. Welded feeders are conjectural. See Fig. 3 for location.

Fig. 7. (A) Seismic section and **(B)** and interpreted line drawing of a 2D seismic profile of the eastern Parentis Basin showing the Pelican Salt Sheet. Welded feeders are conjectural. See Fig. 3 for location.

Fig. 8. (A) Seismic section and **(B)** and interpreted line drawing of a seismic line in the eastern Parentis Basin showing two diapirs with different growth histories. Both diapirs grew passively during the Jurassic and Early Cretaceous. Puffin Ridge then stopped growing and was buried beneath a wedge of Upper Cretaceous-Eocene sediments probably because of thinner original salt closer to the margin of the salt basin. Later Puffin Ridge was rejuvenated by inversion in the early Oligocene. Alcyon Diapir continued to grow passively until the middle Miocene and must have also been squeezed during the Oligo-Miocene inversion. The differential rotation values within synkinematic strata are only for qualitative comparisons because the profile has not been depth converted. See Fig. 3 for location.

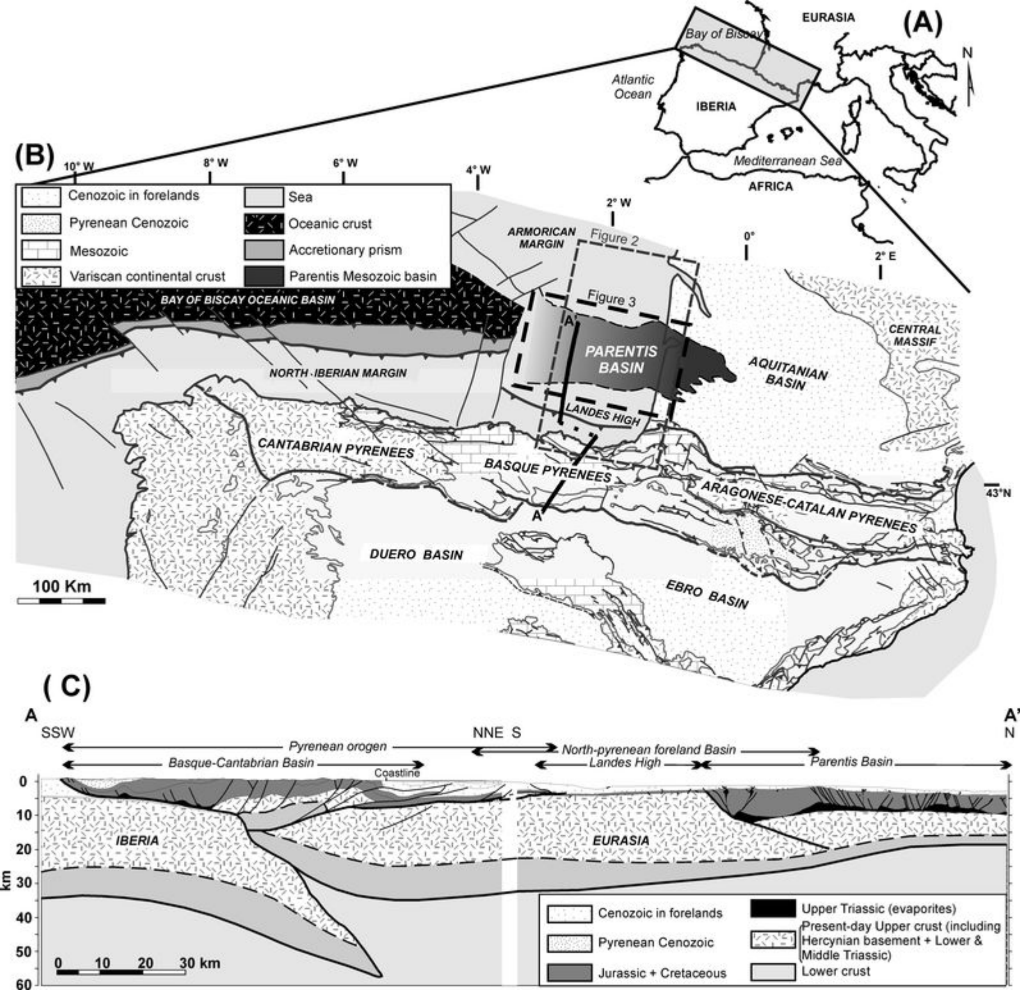
Fig. 9. (A) Seismic section and **(B)** and interpreted line drawing of a seismic line across the western Parentis Basin (deep offshore) showing two salt-cored anticlines and two possibly welded diapirs. See Fig. 3 for location.

Fig. 10. Geometry of Cretaceous and Cenozoic sedimentary units in two anticlines cored by squeezed diapirs in the hanging wall of the Landes Fault in the southern Parentis Basin. Both diapirs grew as passive diapirs during the Jurassic and Early Cretaceous. The diapirs then stopped growing and were buried beneath an Upper Cretaceous roof. They were rejuvenated by inversion in the early Oligocene and again uplifted during the Middle Miocene by contractional reactivation of the faults that controlled early growth of the diapirs. The differential rotation values within synkinematic strata are only for qualitative comparisons because the profile has not been depth converted. See Fig. 3 for location.

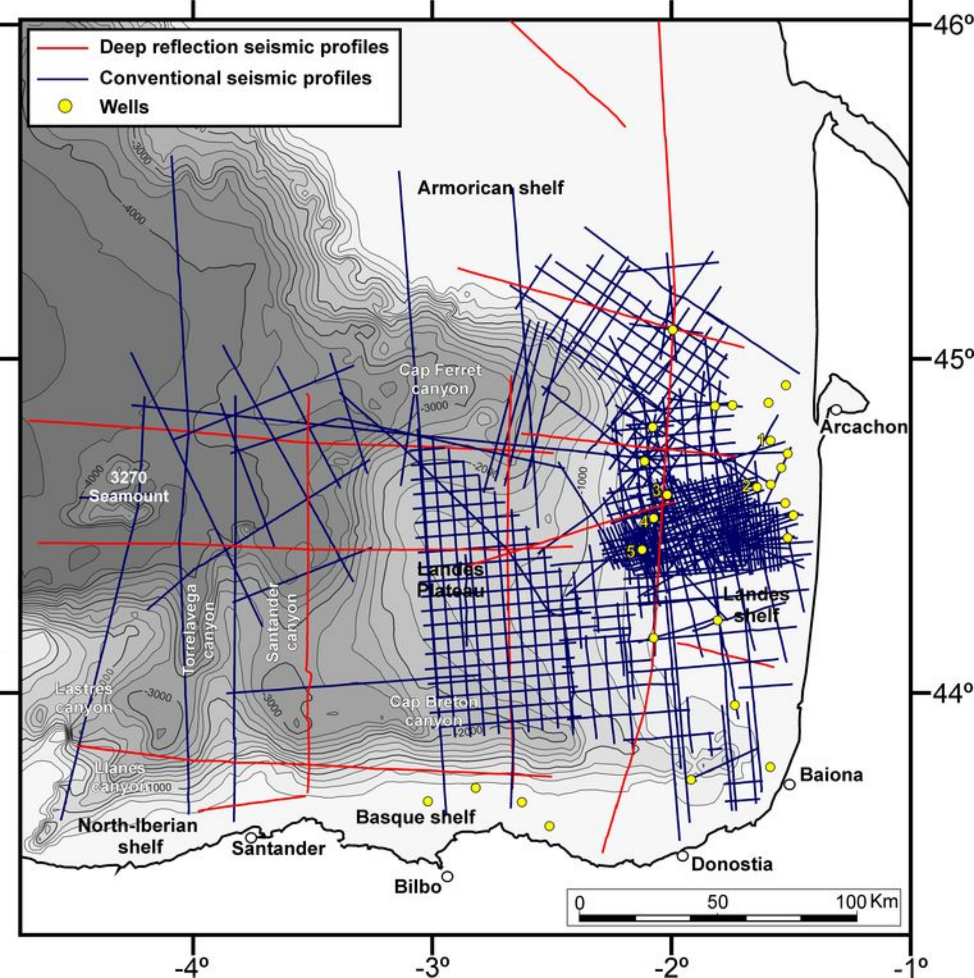
Fig. 11. Summary of the history of salt tectonics and regional tectonics in the offshore Parentis Basin**.**

Fig. 12. Qualitative restoration of the eastern Parentis Basin regional profile showed in Fig. 6 using 2DMove software (see Fig. 3 for location). (a) Evaporite deposition in the Parentis Basin was bounded in the south by a basement high (Landes High) which also controlled the thinning of the Jurassic package. (b-c) Extension produced a major Barremian-Albian depocenter in the Parentis Trough which expelled Keuper evaporites towards the north edge of the basin. A salt-cored anticline formed as a drape fold over the Ibis Fault. (d-f) During the Pyrenean compression the drape fold above the Ibis Fault was uplifted and its crest was eroded. In the south, previously buried, dormant salt walls were rejuvenated by squeezing and arched their previously flat-lying roof (e.g. Puffin Diapir). Later, buried salt walls may have pinched off by regional compression to form

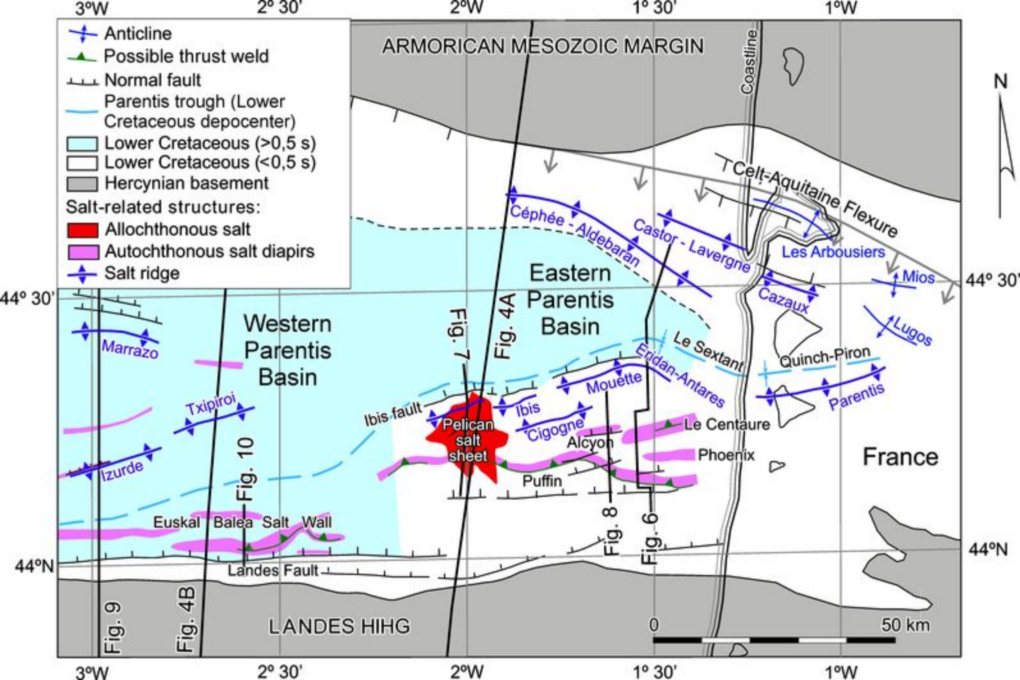
subvertical secondary salt welds (e.g. Alcyon Diapir). Salt tectonics ended during the Miocene.



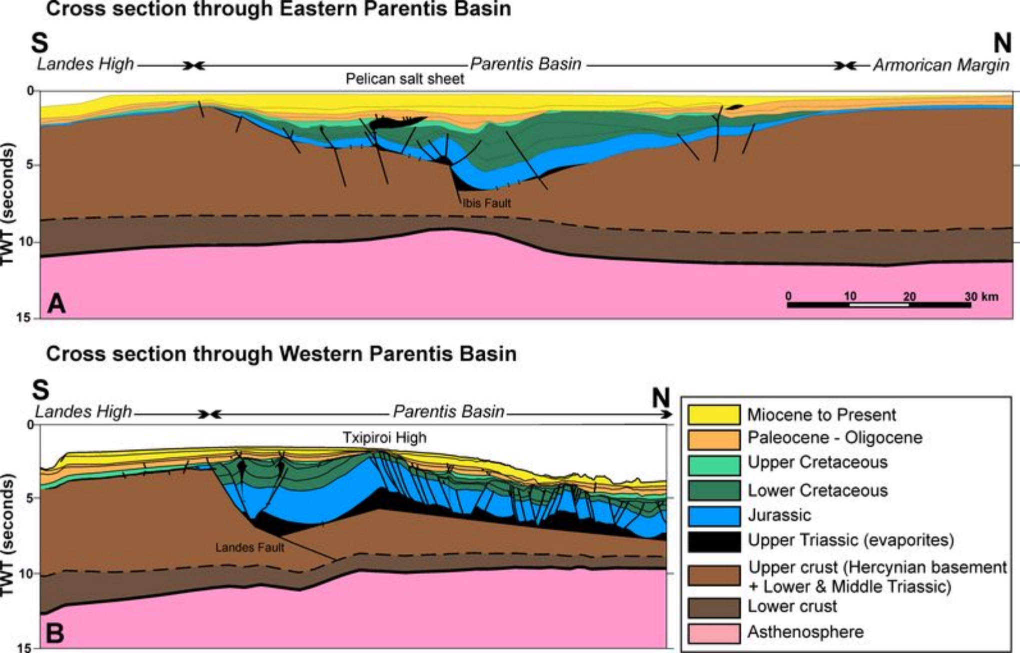
(A)
EURASIA
Bay of Biscay
Atlantic Ocean
IBERIA
Mediterranean Sea
AFRICA
N
(B)
10° W
8° W
6° W
4° W
2° W
0°
2° E
43°N
Cenozoic in forelands
Pyrenean Cenozoic
Mesozoic
Variscan continental crust
Sea
Oceanic crust
Accretionary prism
Parentis Mesozoic basin
ARMORICAN MARGIN
Figure 2
Figure 3
BAY OF BISCAY OCEANIC BASIN
NORTH IBERIAN MARGIN
PARENTIS BASIN
AQUITANIAN BASIN
CENTRAL MASSIF
LANDES HIGH
CANTABRIAN PYRENEES
BASQUE PYRENEES
ARAGONESE-CATALAN PYRENEES
DUERO BASIN
EBRO BASIN
A
A'
100 Km
(C)
A
A'
SSW
NNE S
N
Pyrenean orogen
North-pyrenean foreland Basin
Basque-Cantabrian Basin
Landes High
Parentis Basin
Coastline
IBERIA
EURASIA
km
0
10
20
30
40
50
60
0 10 20 30 km
Cenozoic in forelands
Pyrenean Cenozoic
Jurassic + Cretaceous
Upper Triassic (evaporites)
Present-day Upper crust (including Hercynian basement + Lower & Middle Triassic)
Lower crust

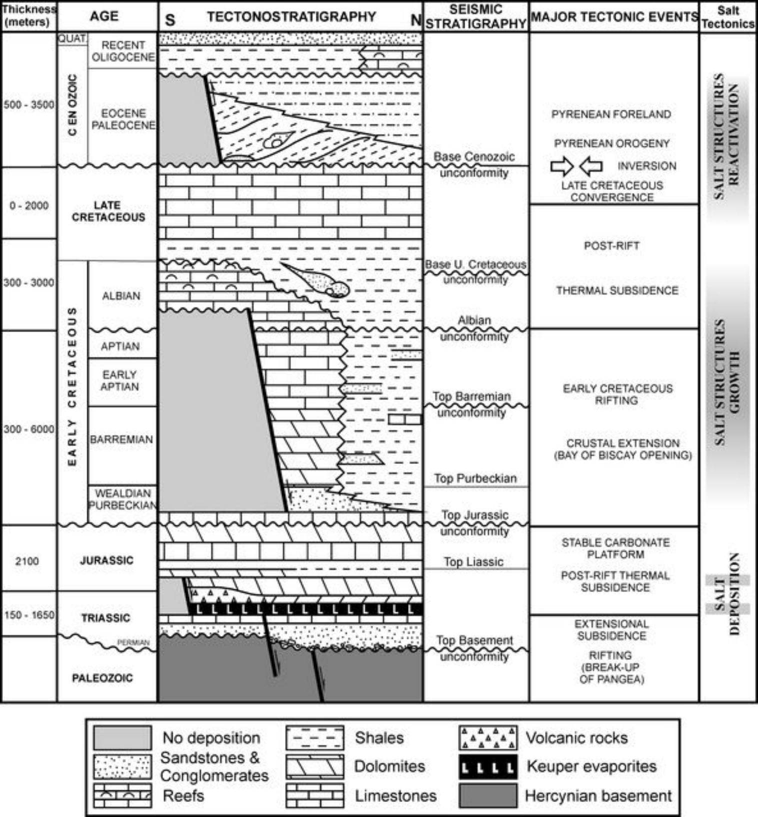


Deep reflection seismic profiles
Conventional seismic profiles
Wells
Armorican shelf
Cap Ferret canyon
3270 Seamount
Landes Plateau
Landes shelf
Torrelavega canyon
Santander canyon
Lastres canyon
Cap Breton canyon
Llanes canyon
North-Iberian shelf
Basque shelf
Santander
Bilbo
Donostia
Baiona
Arcachon
0
50
100 Km
46º
45º
44º
-4º
-3º
-2º
-1º

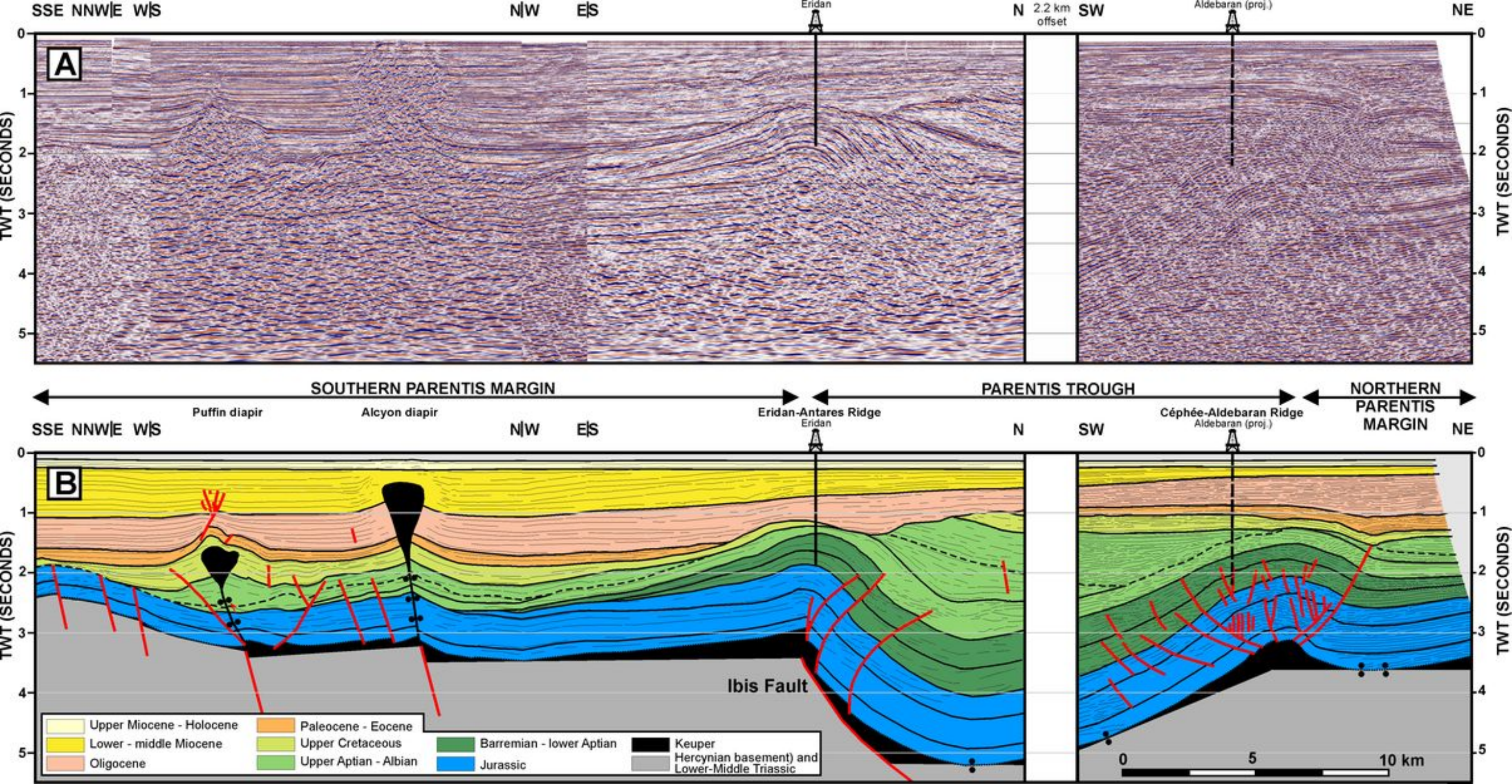


Anticline
Possible thrust weld
Normal fault
Parentis trough (Lower Cretaceous depocenter)
Lower Cretaceous (>0,5 s)
Lower Cretaceous (<0,5 s)
Hercynian basement
Salt-related structures:
Allochthonous salt
Autochthonous salt diapirs
Salt ridge
3ºW
2º 30'
2ºW
1º 30'
1ºW
44º 30'
44ºN
ARMORICAN MESOZOIC MARGIN
Coastline
N
Celt-Aquitaine Flexure
Céphée - Aldebaran
Castor - Lavergne
Les Arbousiers
Mios
Cazaux
Lugos
Quinch-Piron
Parentis
Le Sextant
Eridan-Antares
Mouette
Ibis
Cigogne
Alcyon
Le Centaure
Phoenix
Puffin
Pelican salt sheet
Ibis fault
Eastern Parentis Basin
Western Parentis Basin
Marrazo
Txipiroi
Izurde
Euskal Balea Salt Wall
Landes Fault
LANDES HIHG
France
Fig. 4A
Fig. 7
Fig. 10
Fig. 8
Fig. 6
Fig. 9
Fig. 4B
0
50 km

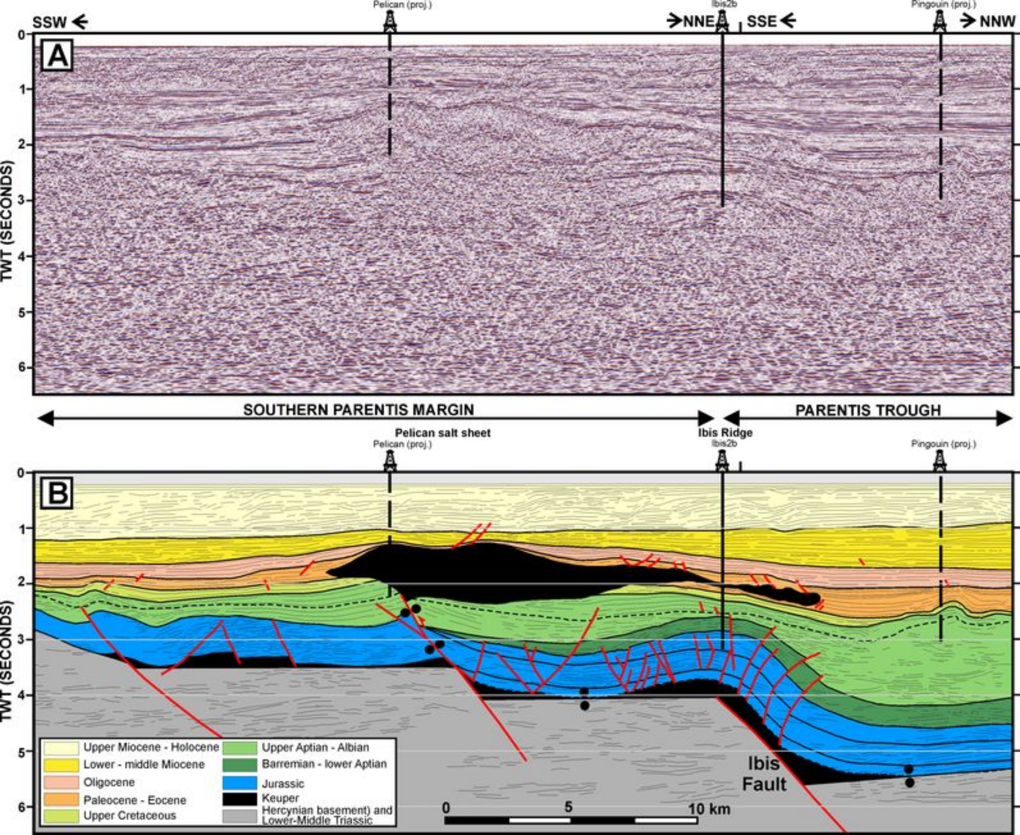

Cross section through Eastern Parentis Basin
S
N
Landes High
Parentis Basin
Armorican Margin
Pelican salt sheet
Ibis Fault
TWT (seconds)
0
5
10
15
A
0
10
20
30 km
Cross section through Western Parentis Basin
S
N
Landes High
Parentis Basin
Txipiroi High
Landes Fault
TWT (seconds)
0
5
10
15
B
Miocene to Present
Paleocene - Oligocene
Upper Cretaceous
Lower Cretaceous
Jurassic
Upper Triassic (evaporites)
Upper crust (Hercynian basement + Lower & Middle Triassic)
Lower crust
Asthenosphere

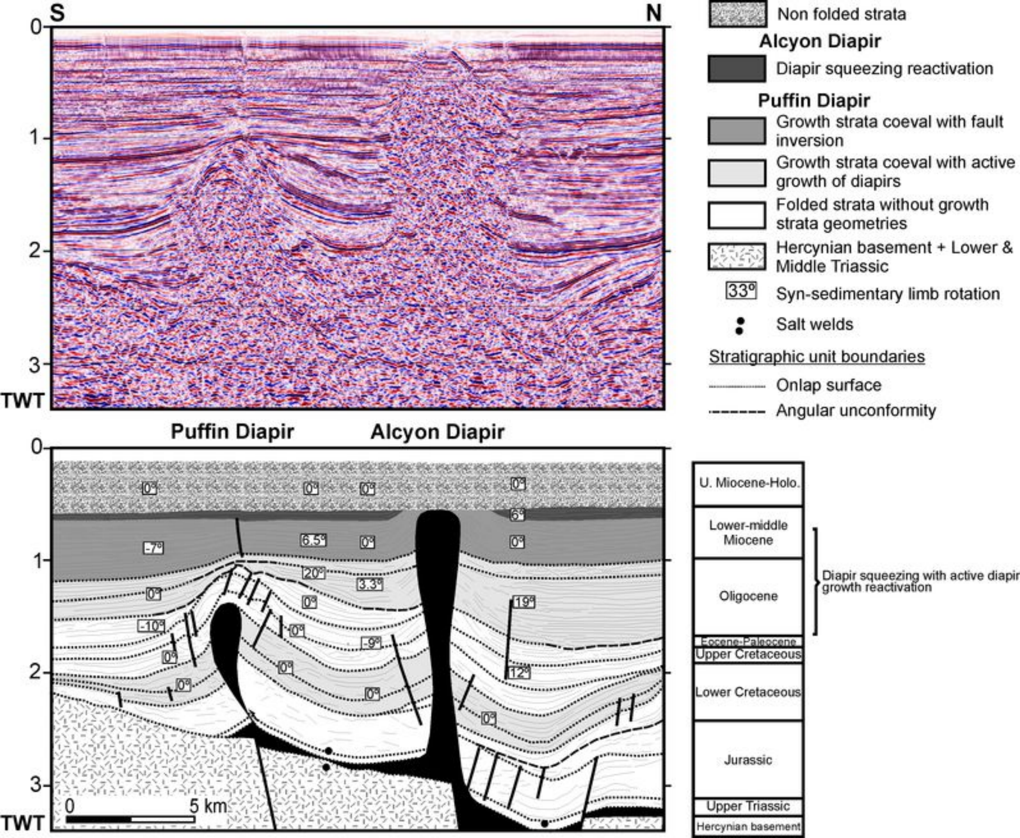


Thickness (meters)
AGE
S
TECTONOSTRATIGRAPHY
N
SEISMIC STRATIGRAPHY
MAJOR TECTONIC EVENTS
Salt Tectonics
QUAT.
RECENT
OLIGOCENE
500 - 3500
CENOZOIC
EOCENE
PALEOCENE
0 - 2000
LATE CRETACEOUS
300 - 3000
ALBIAN
APTIAN
EARLY APTIAN
300 - 6000
EARLY CRETACEOUS
BARREMIAN
WEALDIAN
PURBECKIAN
2100
JURASSIC
150 - 1650
TRIASSIC
PERMIAN
PALEOZOIC
Base Cenozoic unconformity
Base U. Cretaceous unconformity
Albian unconformity
Top Barremian unconformity
Top Purbeckian
Top Jurassic unconformity
Top Liassic
Top Basement unconformity
PYRENEAN FORELAND
PYRENEAN OROGENY
INVERSION
LATE CRETACEOUS CONVERGENCE
POST-RIFT
THERMAL SUBSIDENCE
EARLY CRETACEOUS RIFTING
CRUSTAL EXTENSION (BAY OF BISCAY OPENING)
STABLE CARBONATE PLATFORM
POST-RIFT THERMAL SUBSIDENCE
EXTENSIONAL SUBSIDENCE
RIFTING (BREAK-UP OF PANGEA)
SALT STRUCTURES REACTIVATION
SALT STRUCTURES GROWTH
SALT DEPOSITION
No deposition
Sandstones & Conglomerates
Reefs
Shales
Dolomites
Limestones
Volcanic rocks
Keuper evaporites
Hercynian basement

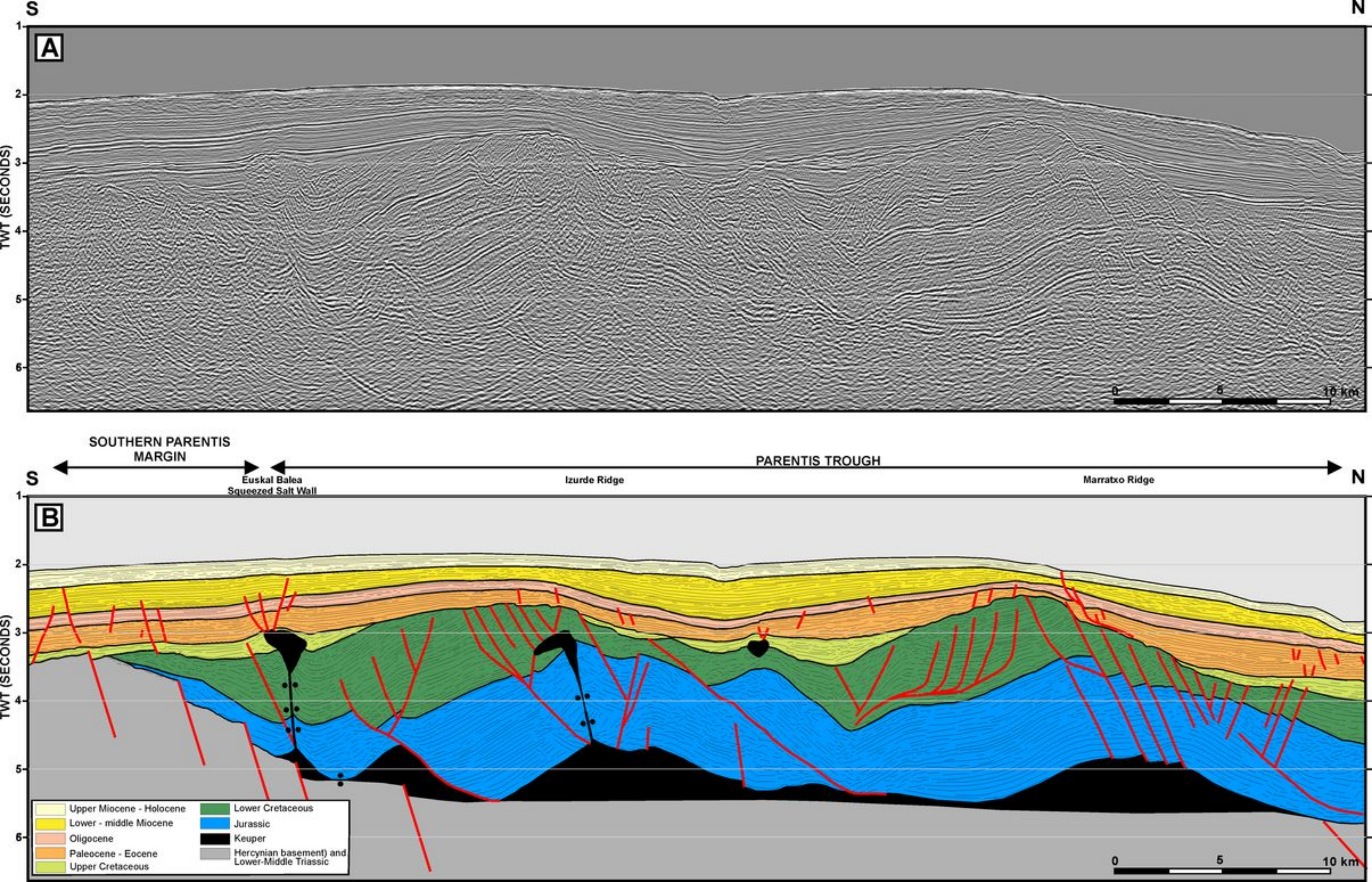


SSE NNW|E W|S
N|W E|S
Eridan
N
2.2 km offset
SW
Aldebaran (proj.)
NE
A
TWT (SECONDS)
SOUTHERN PARENTIS MARGIN
PARENTIS TROUGH
NORTHERN PARENTIS MARGIN
Puffin diapir
Alcyon diapir
Eridan-Antares Ridge
Céphée-Aldebaran Ridge
B
Ibis Fault
Upper Miocene - Holocene
Lower - middle Miocene
Oligocene
Paleocene - Eocene
Upper Cretaceous
Upper Aptian - Albian
Barremian - lower Aptian
Jurassic
Keuper
Hercynian basement) and Lower-Middle Triassic
0 5 10 km

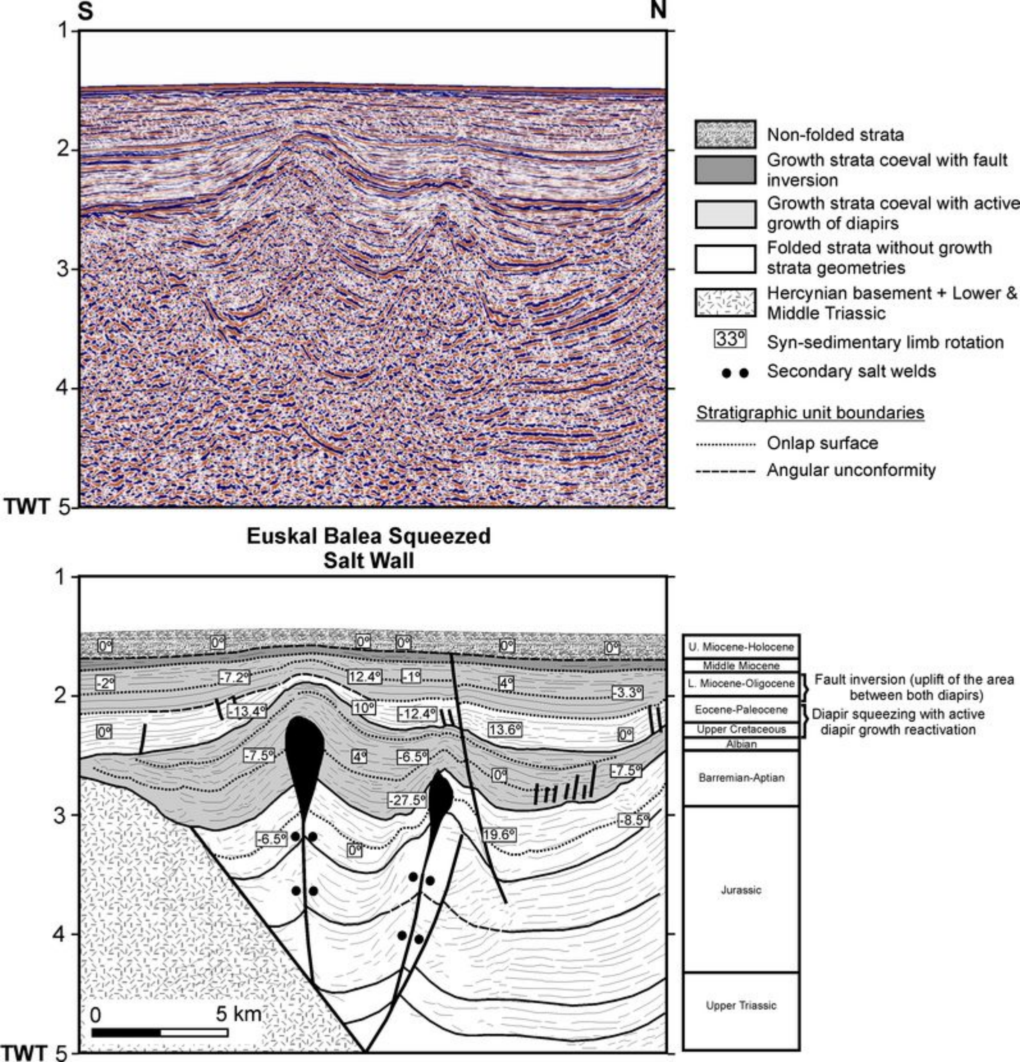

Pelican (proj.)
Ibis2b
Pingouin (proj.)
SSW
NNE
SSE
NNW
A
TWT (SECONDS)
SOUTHERN PARENTIS MARGIN
PARENTIS TROUGH
Pelican salt sheet
Ibis Ridge
B
Ibis Fault
Upper Miocene - Holocene
Lower - middle Miocene
Oligocene
Paleocene - Eocene
Upper Cretaceous
Upper Aptian - Albian
Barremian - lower Aptian
Jurassic
Keuper
Hercynian basement) and Lower-Middle Triassic
0
5
10 km

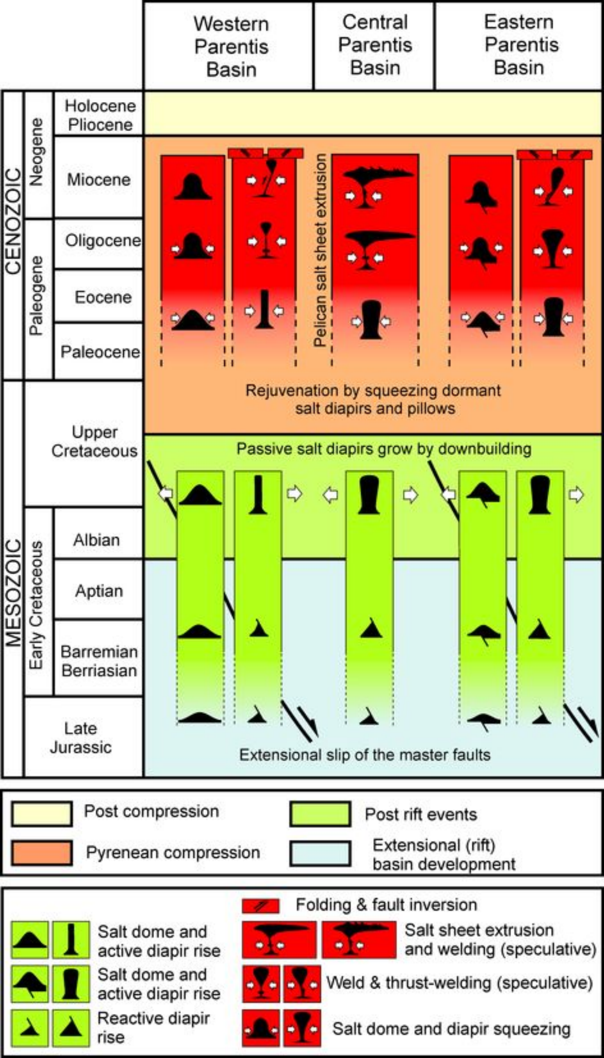


S
N
0
1
2
3
TWT
Non folded strata
Alcyon Diapir
Diapir squeezing reactivation
Puffin Diapir
Growth strata coeval with fault inversion
Growth strata coeval with active growth of diapirs
Folded strata without growth strata geometries
Hercynian basement + Lower & Middle Triassic
33º
Syn-sedimentary limb rotation
Salt welds
Stratigraphic unit boundaries
Onlap surface
Angular unconformity
Puffin Diapir
Alcyon Diapir
0º
6º
-7º
6.5º
20º
3.3º
19º
-10º
-9º
12º
0 5 km
U. Miocene-Holo.
Lower-middle Miocene
Oligocene
Eocene-Paleocene
Upper Cretaceous
Lower Cretaceous
Jurassic
Upper Triassic
Hercynian basement
Diapir squeezing with active diapir growth reactivation

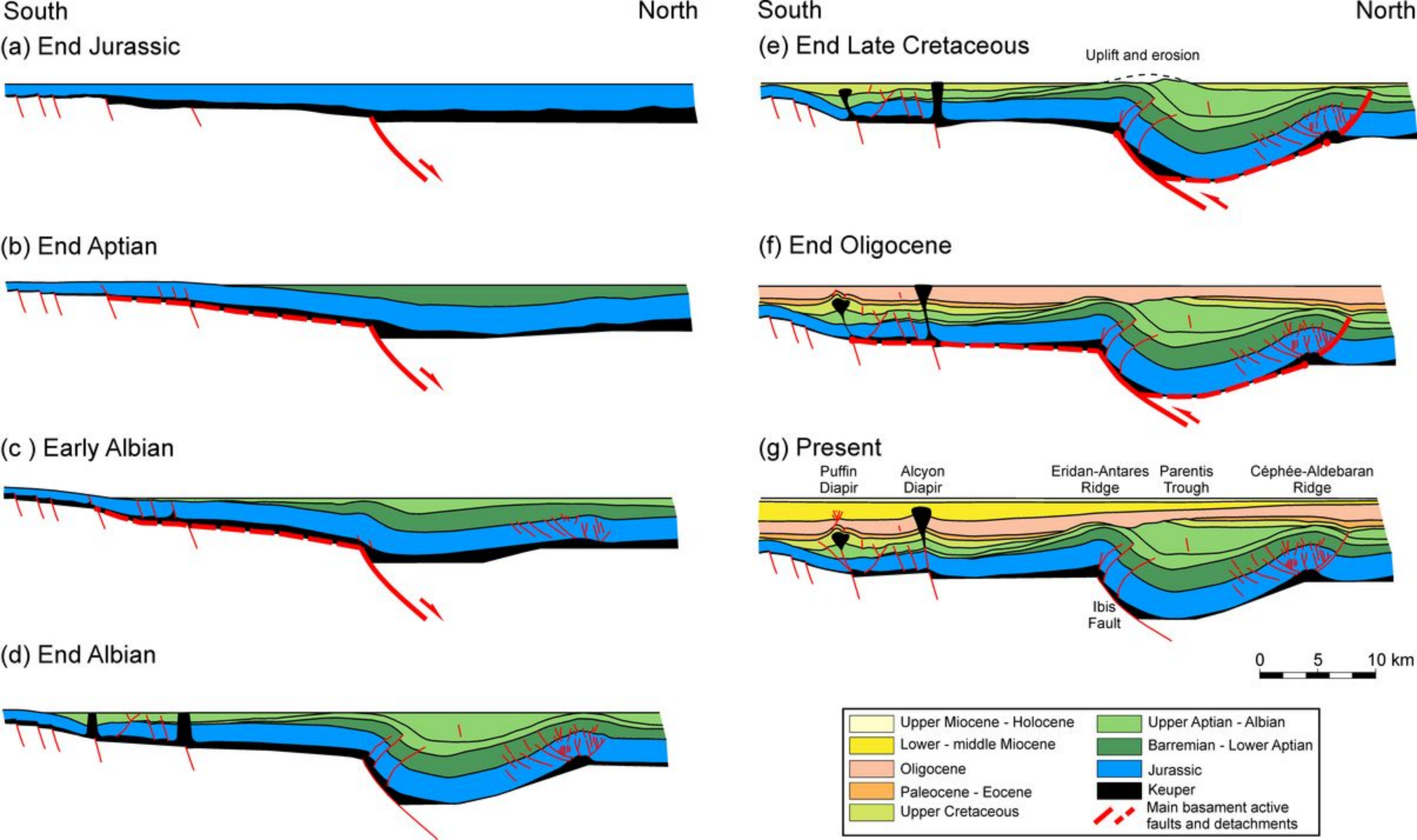


S
N
A
TWT (SECONDS)
1
2
3
4
5
6
0
5
10 km
SOUTHERN PARENTIS MARGIN
PARENTIS TROUGH
Euskal Balea Squeezed Salt Wall
Izurde Ridge
Marratxo Ridge
B
Upper Miocene - Holocene
Lower - middle Miocene
Oligocene
Paleocene - Eocene
Upper Cretaceous
Lower Cretaceous
Jurassic
Keuper
Hercynian basement) and Lower-Middle Triassic


S
N
TWT
Non-folded strata
Growth strata coeval with fault inversion
Growth strata coeval with active growth of diapirs
Folded strata without growth strata geometries
Hercynian basement + Lower & Middle Triassic
33º Syn-sedimentary limb rotation
Secondary salt welds
Stratigraphic unit boundaries
Onlap surface
Angular unconformity
Euskal Balea Squeezed Salt Wall
0 5 km
U. Miocene-Holocene
Middle Miocene
L. Miocene-Oligocene
Eocene-Paleocene
Upper Cretaceous
Albian
Barremian-Aptian
Jurassic
Upper Triassic
Fault inversion (uplift of the area between both diapirs)
Diapir squeezing with active diapir growth reactivation



Western Parentis Basin
Central Parentis Basin
Eastern Parentis Basin
CENOZOIC
MESOZOIC
Neogene
Paleogene
Early Cretaceous
Holocene Pliocene
Miocene
Oligocene
Eocene
Paleocene
Upper Cretaceous
Albian
Aptian
Barremian Berriasian
Late Jurassic
Pelican salt sheet extrusion
Rejuvenation by squeezing dormant salt diapirs and pillows
Passive salt diapirs grow by downbuilding
Extensional slip of the master faults
Post compression
Pyrenean compression
Post rift events
Extensional (rift) basin development
Folding & fault inversion
Salt dome and active diapir rise
Salt dome and active diapir rise
Reactive diapir rise
Salt sheet extrusion and welding (speculative)
Weld & thrust-welding (speculative)
Salt dome and diapir squeezing


South
North
(a) End Jurassic
(b) End Aptian
(c) Early Albian
(d) End Albian
South
North
(e) End Late Cretaceous
Uplift and erosion
(f) End Oligocene
(g) Present
Puffin
Diapir
Alcyon
Diapir
Eridan-Antares
Ridge
Parentis
Trough
Céphée-Aldebaran
Ridge
Ibis
Fault
0
5
10 km
Upper Miocene - Holocene
Lower - middle Miocene
Oligocene
Paleocene - Eocene
Upper Cretaceous
Upper Aptian - Albian
Barremian - Lower Aptian
Jurassic
Keuper
Main basament active
faults and detachments