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Structural geometry in the eastern Pyrenees and western Gulf of Lion (Western Mediterranean)

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Abstract

We present new seismic data from the Gulf of Lion located east of the Pyrenees on the continental shelf of the Mediterranean Sea. The deep penetration LISA (Ligurian–Sardinia Sea) seismic lines, the shots of the LISA cruise recorded on land, and the high definition ELF seismic sections allow us to present a complete picture of the tectonics in this area from the surface to the Moho level, and also to document late Miocene–early Pliocene extensional tectonics in the area. Previous studies show a prominent thinning of the crust observed from the Pyrenees towards the Gulf of Lion. The Moho depth varies from 48 km beneath the Axial Range crust (thickened during the Pyrenean Eocene Orogeny) to 21 km below the Catalan Basin in the Gulf of Lion. This crustal thinning occurred mainly during the early Miocene extension of the Mediterranean Sea. Balanced reconstructed geological sections derived from reflection and refraction seismic data allow us to evaluate the stretching factors at the crustal level. A maximum extension of 25 km is computed for the Catalan Basin area. This extension is related to detachments that penetrate the crust as deep as 11 km to the base of the brittle crust. These intra basement detachments have been confused in the past with the Paleozoic acoustic basement. The detachments show a clear listric shape and the geometry of horst and grabens can be explained by a hanging wall and footwall configuration with isostatic rebound of the footwall. The uplift in the Eastern Pyrenees (Albères and Canigou Massifs), on the other hand, is related to the late Miocene–early Pliocene extension we mapped in the area. These elevated features, probably formed by isostatic rebound, are surrounded by deep basins such as the Roussillon and El Empordà depressions. A 1.7 km uplift during the late Miocene–early Pliocene is computed in the offshore part of the Albères Massif. The cause of this Late Miocene–early Pliocene extension is not well explained although an uplift related to the Messinian desiccation or a thermal anomaly in the mantle have been proposed. The relationship between the Eastern Pyrenees and the Gulf of Lion is governed by NE–SW transfer faults. These faults represent the southwestern limits of the Gulf of Lion basins. Although much seismicity is recorded in the Eastern Pyrenees, we do not see evidence of present tectonics in the Gulf of Lion as the extensional faults were active only until the early Pliocene. Therefore, the present stress is probably compressional, as shown by the focal mechanisms of the earthquakes, although the trends of compressional vectors are divergent in Spain and France. © 2001 Elsevier Science Ltd. All rights reserved.

Keywords: Seismic lines; Tectonics; Crustal thinning

1. Introduction

The formation of mountain ranges and their subsequent thinning and the relationship between plate motion and the net convergence vector are major questions addressed by researchers worldwide. The Mediterranean area is an excellent place to elucidate these kinds of problem. As mountain ranges are generally located at the boundaries between European countries, it concentrates efforts of different scientific communities, resulting in a large amount of available data. Besides that, the Eastern Pyrenees are cut at right angles by the Mediterranean coast and this disposition allows correlation of marine and on-land tectonic architectures. Moreover, the Gulf of Lion, east of the Eastern Pyrenees, is the target for periodic oil prospecting. More recently, the deep penetration multichannel LISA (Ligurian–Sardinia Sea) survey, carried out in 1995, and a

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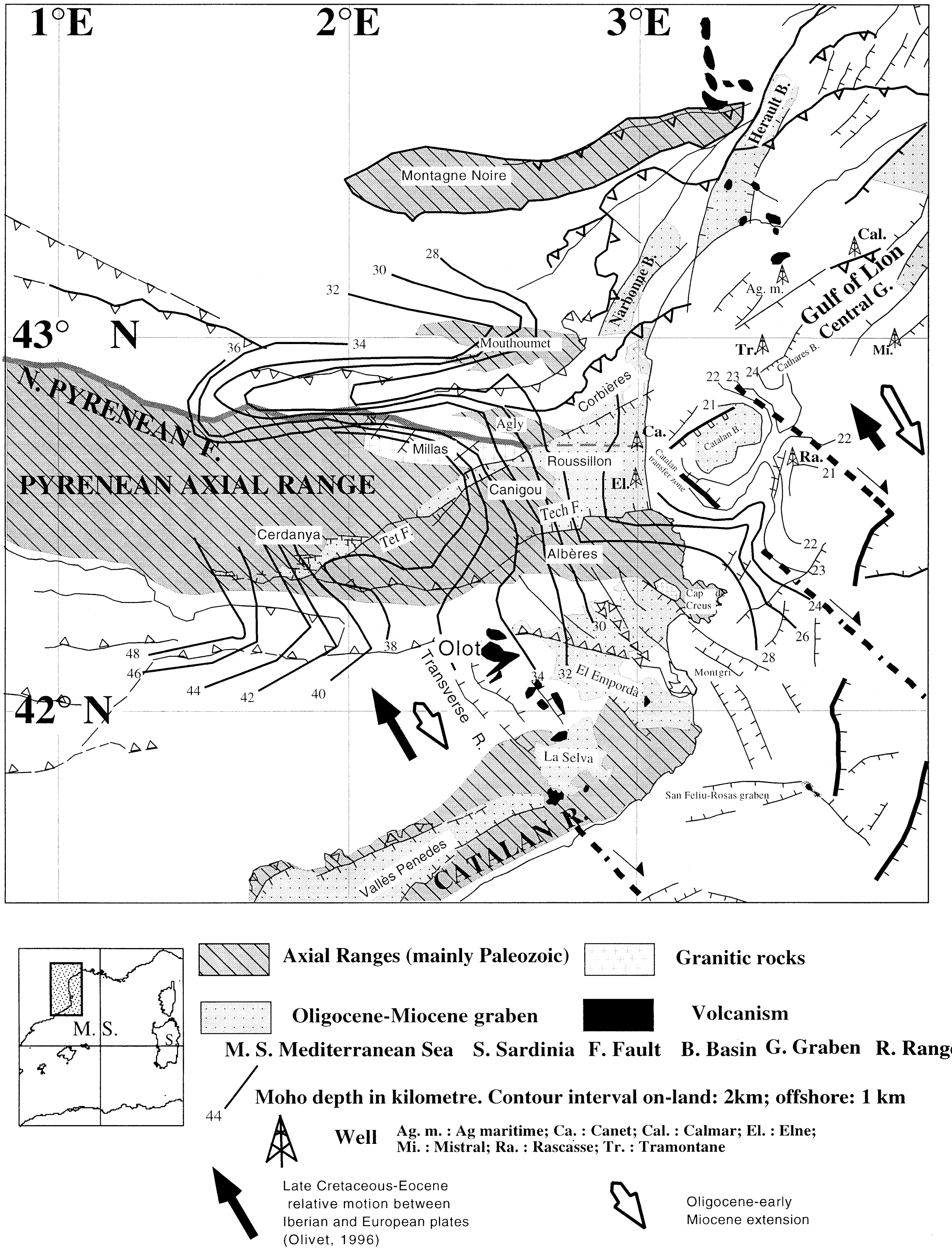

1°E
2°E
3°E
43° N
42° N
Montagne Noire
Herault B.
Narbonne B.
Gulf of Lion
Central G.
Cal.
Ag. m.
Tr.
Mi.
Cathares B.
Mouthoumet
N. PYRENEAN F.
PYRENEAN AXIAL RANGE
Corbières
Agly
Millas
Roussillon
Ca.
El.
Catalan B.
Catalan transfer zone
Ra.
Canigou
Tech F.
Tet F.
Cerdanya
Albères
Cap Creus
Olot
Transverse R.
El Empordà
Montgri
La Selva
San Feliu-Rosas graben
Vallès Penedes
CATALAN R.
M. S.
S
Axial Ranges (mainly Paleozoic)
Granitic rocks
Oligocene-Miocene graben
Volcanism
M. S. Mediterranean Sea S. Sardinia F. Fault B. Basin G. Graben R. Range
Moho depth in kilometre. Contour interval on-land: 2km; offshore: 1 km
Well
Ag. m. : Ag maritime; Ca. : Canet; Cal. : Calmar; El. : Elne;
Mi. : Mistral; Ra. : Rascasse; Tr. : Tramontane
Late Cretaceous-Eocene relative motion between Iberian and European plates (Olivet, 1996)
Oligocene-early Miocene extension

high quality oil industry seismic survey, performed in 1996 by ELF, provide data to construct a complete picture of the tectonic evolution from the Moho to the sea floor. In addition, on-land records (Nercessian et al., 2001; Vidal et al., 2001) of shots fired during the LISA cruise allow us to tie in the onshore and offshore tectonic structures at the crustal level. In the present study, we concentrate on the results of the upper tectonic levels, though considerations of deep crust structures (Nercessian et al., 2001; Vidal et al., 2001) will also be taken into account when deducing the tectonic evolution of this region and calculating the rate of extension.

2. Geological and geophysical setting

The Pyrenean Range forms the boundary between Spain and France running from the Atlantic Ocean to the Mediterranean Sea. From previous geophysical studies (Casas et al., 1997) it is clear that the mountain building varies from east to west and that the structure of the Eastern Pyrenees cannot be extrapolated westwards. We examine the Eastern Pyrenean Range and its relationships with the Gulf of Lion (Fig. 1). The formation of the Pyrenees follows a complex tectonic history that began in the Palaeozoic with compressional deformation and metamorphism in the Pyrenean Axial Range (Arthaud and Matte, 1975). During the Late Permian–Triassic periods, the Hercynian-thickened crust was thinned (Vissers, 1992). Nevertheless, this first extensional event has been largely obliterated by a prominent thinning that occurred during the Middle Cretaceous. At about 90 Ma, the Atlantic Ocean was deeply modified and the Iberian Plate moved along a left-lateral strike-slip fault relative to the European Plate (Olivet, 1996, and references therein). This plate boundary is represented by the North Pyrenean Fault that bounds the Paleozoic axial zone to the north (Fig. 1). The strike-slip motion along the North Pyrenean Fault was coeval with a prominent crustal thinning that occurred in the basins emplaced on the European crust. Although this process is mainly active along the western portion of the range (Casas et al., 1997) it is also recorded along the eastern part (Paquet and Mansy, 1991) in the Agly Massif (Fig. 1). During the Late Cretaceous–Eocene, the Iberian Plate had a convergent motion relative to the European Plate (Olivet, 1996). The rotation pole was located southwest of Iberia and consequently the compressional motion should increase eastwards. However, a decrease of the Eocene compression is observed (Gorini et al., 1993, 1994; Gorini, 1994; Guennoc et al., 1994; Mauffret and Gorini, 1996) from the Eastern Pyrenees to the Gulf of Lion (Fig. 1) and this is compatible with a transfer of the compression from the Eastern Pyrenees to Calabria (southeast of Sardinia) (Rehault et al., 1984; Mauffret and Gorini, 1996; Olivet, 1996). Geophysical studies based on refraction and deep penetration seismic ECORS profiles (Gallart et al., 1980, 1981; Daignieres et al., 1981, 1982; Surinach et al., 1993) show the crust to be much thicker on the Iberian (Spanish) side rather than the French side (48 km versus 36 km, Fig. 1). The geological observations also indicate that the Mesozoic sequence deformed by the subsequent Eocene compression extends further south in Spain than in France. The Montgri thrusted rocks (Bilotte et al., 1979; Pujadas et al., 1989) along the El Empordà coast (Fig. 1) show the large southwards extension of the Eocene transported terranes. The Montgri nappes may have been transported from the north of the Pyrenean Axial Range and originated from a Cretaceous basin located in the Gulf of Lion (Bilotte et al., 1979). The Corbières virgation (Arthaud and Mattauer, 1972), to the north of the Roussillon Basin (Fig. 1), suggests a larger implication of the sedimentary beds during the Pyrenean compression in SE France and the Gulf of Lion areas than in SW France. A flip in the Eocene transport has been proposed (Séranne et al., 1995) along a transfer fault between a northwards greater motion in the east (Gulf of Lion area) and southwards larger motion in the west (El Empordà region and eastern Spain). If an initial difference in crustal thickness from west to east may have existed during the Eocene compression in relation to the transfer motion between the Eastern Pyrenees and Calabria, the main crustal thinning from west to east (Fig. 1) is clearly related to the Oligocene–early Miocene rifting that preceded the opening of the Western Mediterranean Sea (Rehault et al., 1984; Gorini et al., 1993, 1994; Gorini, 1994; Guennoc et al., 1994; Mauffret et al., 1995; Mauffret and Gorini, 1996). Several exploratory wells (Fig. 1) show that the upper synrift sequence is middle Aquitanian in age (Cravatte et al., 1974; Lefebvre, 1981; Gorini et al., 1993). From refraction studies (Gallart et al., 1980, 1981) it is evident that the thick crust beneath the Pyrenean Axial Range and Northern Spain is progressively thinned from 48 to 23 km below the Mediterranean coast (Fig. 1), although some local irregularities may occur beneath particular structures (North Pyrenean Fault, Mouthoumet, Fig. 1). Where the Moho rises, the trend of its isobaths is NNW–SSE (Gallart et al., 1980), which is parallel to the western boundary of the Roussillon Basin (Fig. 1) and the main structures in the El Empordà region (Tassone et al., 1996; Lewis et al., 2000). In contrast, the Roussillon Basin is bounded by the Tet and the Tech faults that trend NE–SW, almost normal to the previous orientation. Also, the horsts drilled (Agde Maritime, Mistral, Tramontane,

Fig. 1. Tectonic framework of the Pyrenees and Gulf of Lion. On-land geology from Anadón et al. (1982), Calvet (1985), Clauzon et al. (1987b), ICC, Mapa geològic de Catalunya 1/250 000 (1989) and Arthaud and Pistre (1993) and offshore geology from Gorini et al. (1994), Mauffret et al. (1995) and Tassone et al. (1996). Moho depth redrawn from Gallart et al. (1981), Nercessian et al. (2001) and Vidal et al. (2001) for the continental shelf in the Gulf of Lion. Note the deepening of the Moho below the Canigou and Albères Massifs, Cap de Creus and Rascasse highs and the rise beneath the Catalan Basin.

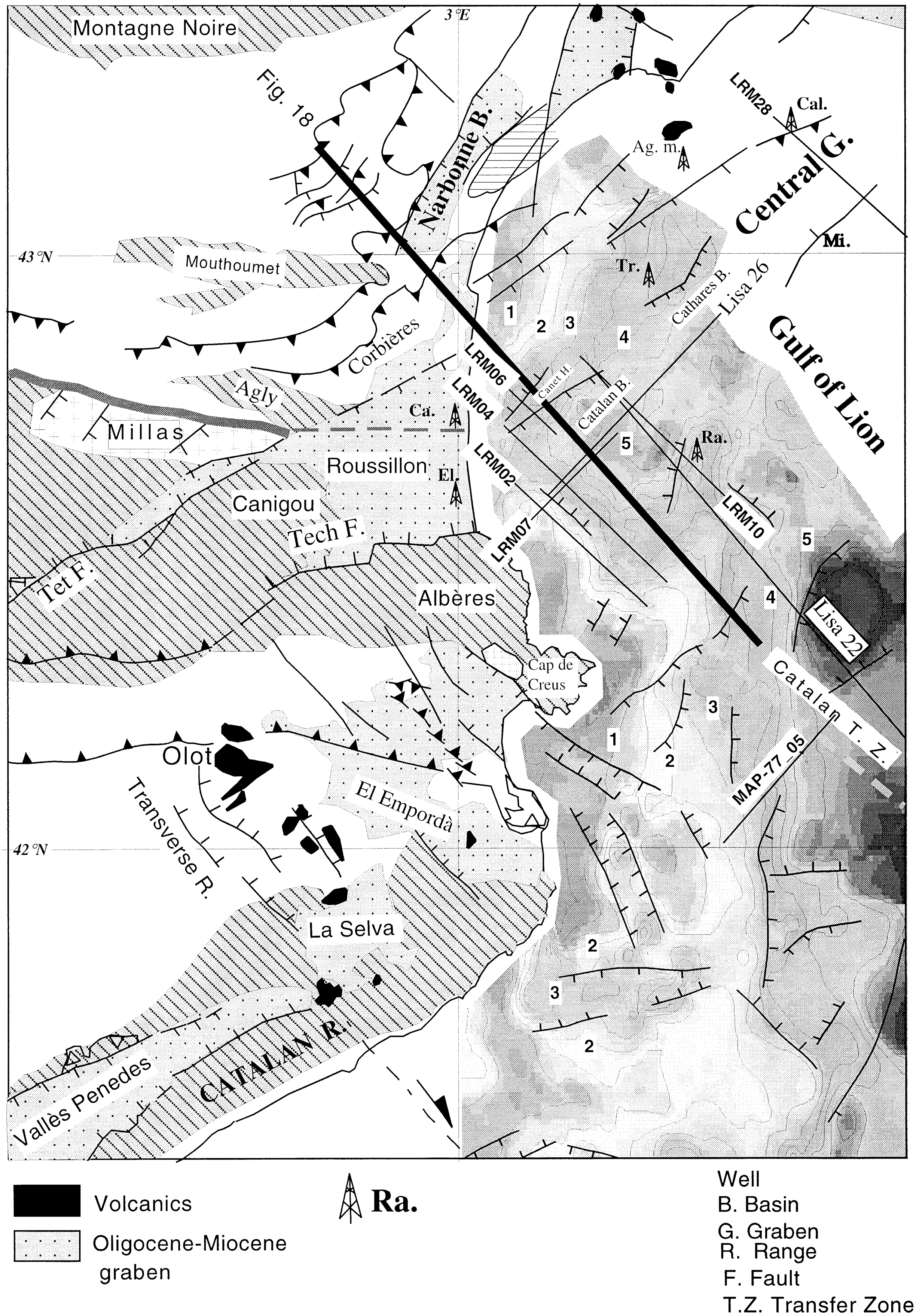


Fig. 2. Depth-to-basement map of the western Gulf of Lion and Catalan margin. Contour interval: 0.5 km. The location of the seismic profiles used in this study is indicated.

Rascasse wells) in the Gulf of Lion show the same NE–SW trend (Fig. 1), and so does the Vallès Penedes Graben, in the central part of the Catalan Range (Anadón et al., 1982). A rather similar trend (E–W) is observed offshore along the San Feliu–Rosas graben (Fig. 1).

The on-land recordings of the LISA shots (Nercessian et al., 2001; Vidal et al., 2001) and the LISA seismic lines allow us to contour the Moho depth beneath the coast and the continental shelf. The central part of the Gulf of Lion is occupied by a graben that is divided into the Cathares and the Catalan sub-basins (Gorini, 1994). Beneath the Catalan Basin, the Moho rises up to 21 km showing a local crustal thinning related to this basin formation (Fig. 1). This crustal thinning is less important than a previous estimation of 15 km depth, computed from gravity modeling (Guennoc et al., 1994). The basin is also well delimited by crustal thickening (Fig. 1), as particularly illustrated by a deepening of the Moho (28 to 23 km contours) along the Cap de Creus and Rascasse highs (Rascasse well: Ra, Fig. 1).

We will discuss in detail the recent tectonic evolution from the late Miocene to the present and the main results and hypothesis will be presented in the discussion.

3. Geological description of the seismic profiles in the Catalan Basin

These seismic profiles are placed in a general context with a depth-to-basement and structural map (Fig. 2). This map is a compilation of our own data base (Maillard, 1993; Gorini, 1994; Gorini et al., 1994; Mauffret et al., 1995) but also of Spanish data on the Catalan margin (Bartrina et al., 1992; Tassone et al., 1996). However, all seismic data in the western part of the Gulf of Lion (Catalan basin) have been

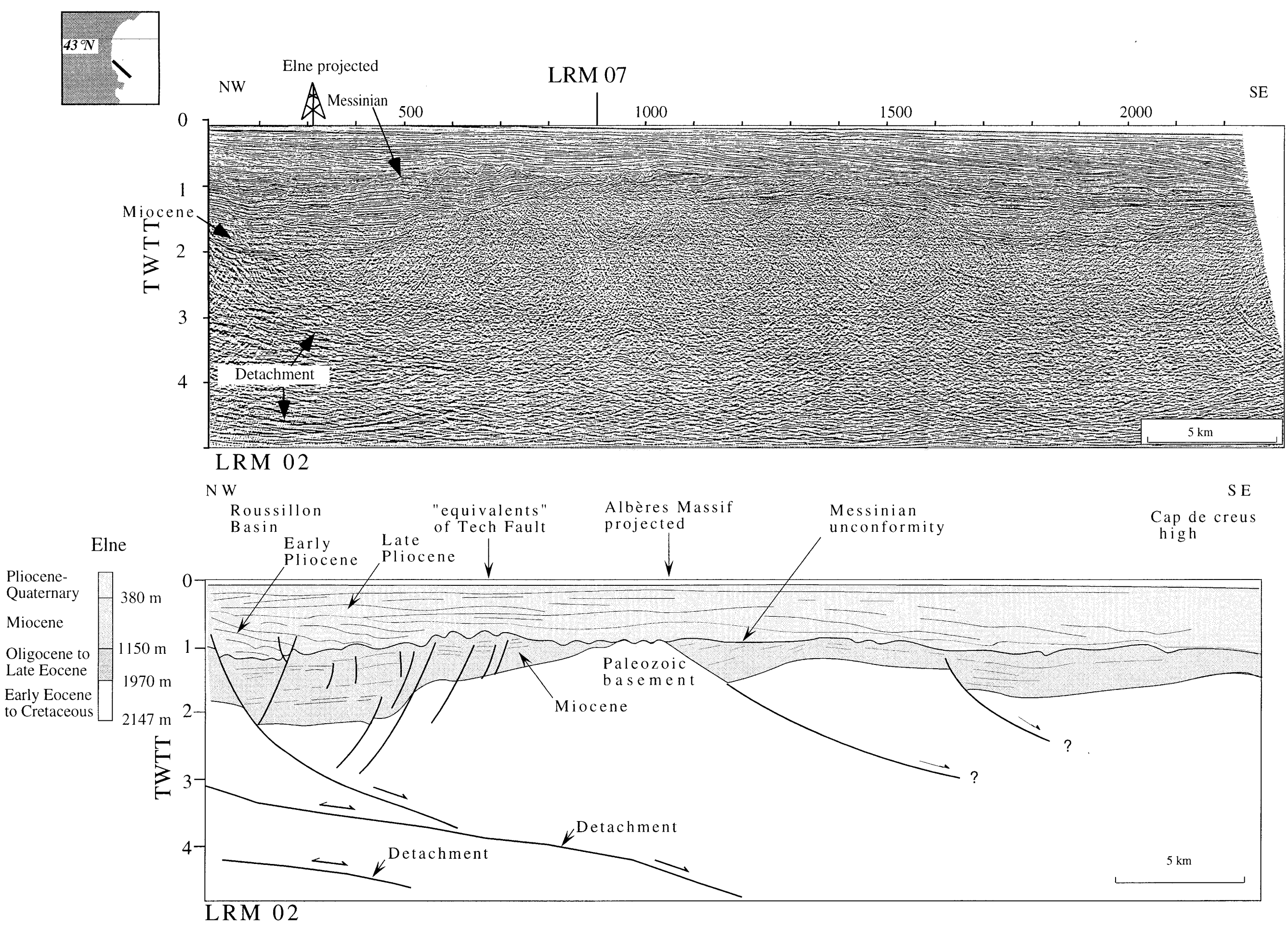


Fig. 3. Seismic profile LRM 02. Location in the inset and Fig. 1. This profile shows the extension on the continental shelf of the Albères Massif and Roussillon Basin. Note the gentle slope of the boundary between the two features. Although some faults show a 300 m throw, the eastern extension of the Tech Fault is not evident. Moreover, the Albères Massif plays a role of hanging wall during the extensional formation of the Roussillon Basin and a steep fault is not required along the northern boundary of this feature. The 'equivalents' of the Tech Fault could be antithetic faults of the main detachment. The Miocene is about 0.6 km tilted along the flank of the Albères Massif and on the top of this massif (shot point 1000). About 1 km of Miocene are eroded by the Messinian erosional surface. The Messinian erosional surface is about 1 km deep in the Roussillon Basin (1 s TWTT) and this result does not fit (380 m deep) with the interpretation (Gottis, 1958) of the Elne borehole because the stratigraphic attributions are probably erroneous. A basin separates the Albères Massif extension and the Cap de Creus high. Two deep detachments in the basement can be observed. These detachments, probable former Eocene thrusts, are not very active because the opening of the Roussillon Basin is minor compared with those of the Catalan Basin.

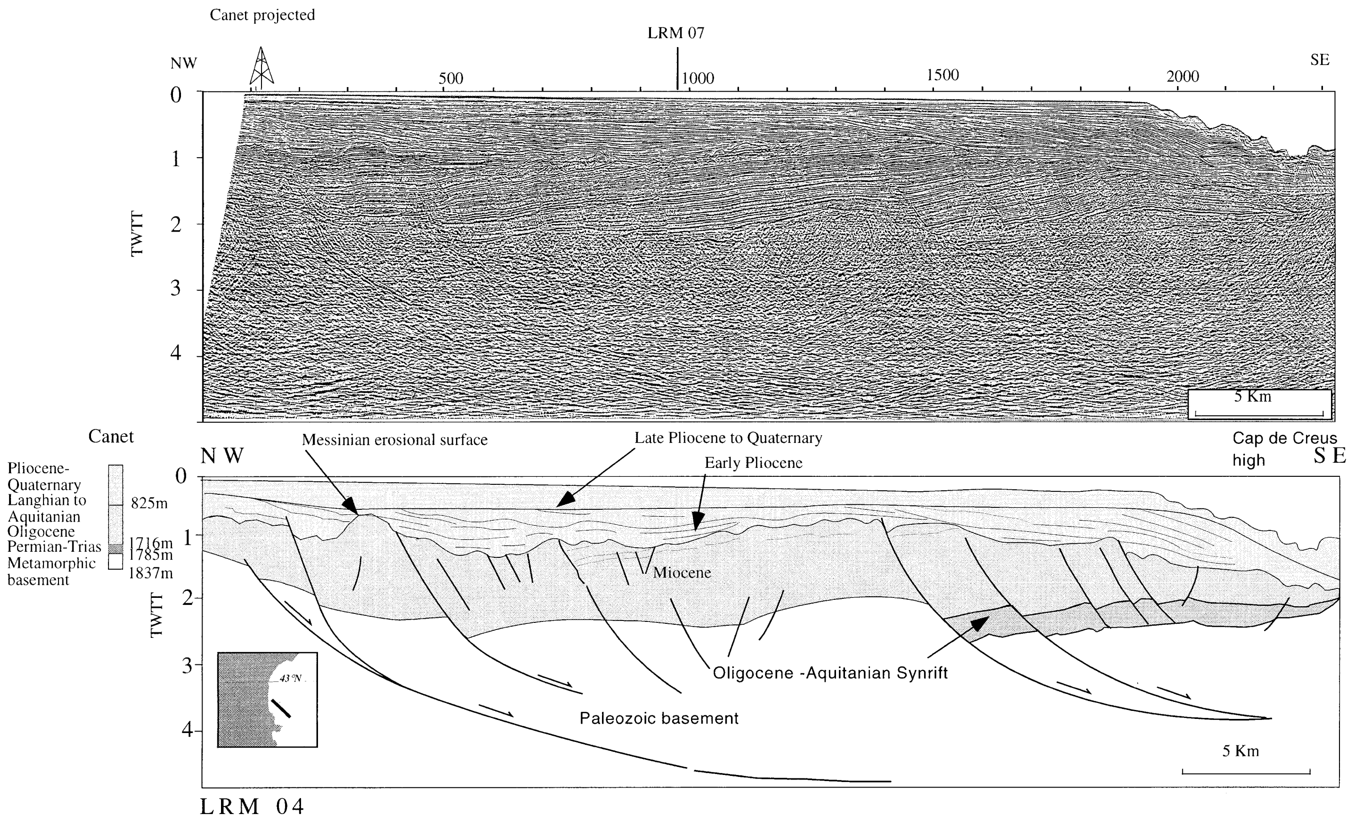


Fig. 4. Seismic profile LRM 04. Location in the inset and Fig. 1. When the Catalan transfer zone is crossed, a deep basin is formed and the Albères Massif disappears. The formation of the basin is related to a rotated block along a deep detachment fault particularly evident. Note that this detachment fault has been reactivated during the late Miocene and the early Pliocene but the onlaps of the late Pliocene indicate the end of activity. Late Miocene faults sealed by the Messinian unconformity offset the Miocene beds that are deeply eroded beneath the shelf break. The tilting of the Miocene layers is related to a late Miocene reactivation of the detachment fault.

studied again and the old profiles have been crossed with the new ELF seismic survey and LISA cruise. The main difference between the new map (Fig. 2) and the previous maps (Gorini, 1994; Gorini et al., 1994; Mauffret et al., 1995) is the maximum depth of the Catalan Basin (5 km in the new data and 7–8 km in the previous estimations). We will explain later why a large discrepancy was observed.

The seismic profile LRM 02 (Fig. 3) is located very close to the shoreline and shows the main tectonic features of the Roussillon Basin and adjacent easternmost Pyrenees (the Albères Massif and Cap de Creus). As the depth-to-basement map (Fig. 2) and the seismic profile LRM 02 (Fig. 3) show the similarity between the onshore and offshore structures, we give these the same name (Roussillon, Albères, Cap de Creus high, Fig. 3) on land and at sea. It is clear from previous geophysical data (gravity map; seismic profiles; Gottis, 1958) that, on land, the deepest part of the Roussillon Basin lies in its southern portion where the Elne well is located. In the offshore Roussillon Basin, the deepest part corresponds to the projection of the Elne borehole (Fig. 3). Nevertheless, the results from the Elne exploratory well (Gottis, 1958) are poorly correlated with the offshore data. The top of the late Miocene is only 380 m deep in the well, whereas the Messinian unconformity is about 1 km deep on the continental shelf (1 s two-way travel time, TWTT, Fig. 3). In the borehole, the base of the Miocene, 1150 m deep, overlies a layer supposed to be Oligocene–late Eocene in age. This horizon lies between 1150 and 1970 m. The layers drilled further (up to 2147 m, the borehole bottom) are interpreted as the Early Eocene to Late Cretaceous sandstone and black marls weakly metamorphosed and partially transformed into schist. In the offshore Roussillon Basin, the postrift Miocene layers are thick (Fig. 3) and the acoustic basement is about 2 s (more than 3 km) deep. The early Miocene–Oligocene synrift layer is absent in the offshore portion of the basin (Fig. 3). Besides that, the difference between the onshore and offshore depths of the Messinian unconformity (380 m versus 1 km) and the early Miocene–Oligocene sequence (1150 m versus 3 km) is too important. Therefore, we suspect an erroneous age for the stratigraphic column in the Elne exploratory well (Gottis, 1958) because the micropaleontological faunas were not yet well known in the Mediterranean area at that time (G. Clauzon, written communication). Therefore, the so-called Oligocene–Late Eocene horizon of the Elne exploratory well could be Miocene in age. Nevertheless, the on-land higher altitude of the Messinian unconformity may be explained by the

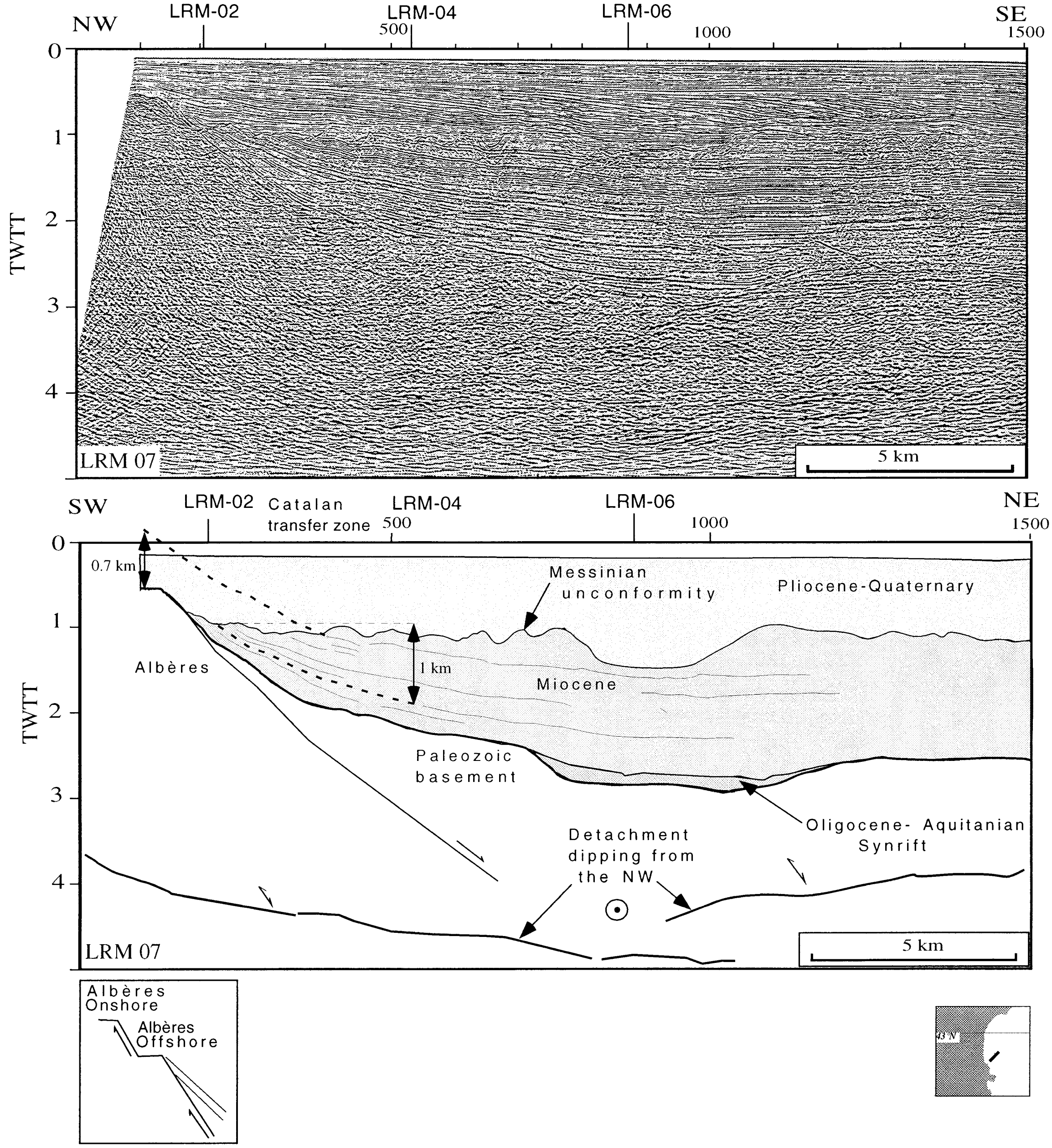


Fig. 5. Seismic profile LRM 07. Location in the inset and Fig. 1. This profile is normal to the Catalan transfer zone. The crossing points with LRM seismic profiles are indicated. On the left part of the profile the detachment level is not evident but it is correlated with other seismic profiles (LRM 02 and 04) where its image is clearer. The detachments are dipping from the NW to the SE and the seismic profile is almost normal to this direction. The extensional motion along the detachments is illustrated by the inclined arrows and the black point surrounded by a circle. Note that the deeper detachment, probably a former Eocene thrust, is almost inactive beneath the Albères Massif (see LRM 02, Fig. 3) where the crust is thick and active below the Catalan Basin (see LRM 04, Fig. 4). The deep detachment is progressively concealed (shot point 1000) by a shallow one. Observe the 1 km tilting of the pre-Messinian layer along the Albères Massif and the 0.7-km-thick Miocene eroded during the Messinian desiccation on the top of the offshore Albères Massif. This configuration suggests a 1.7 km uplift of the offshore Albères Massif along a ramp that is located along the Catalan transfer zone.

location of Elne well on the northern flank of the Tech Messinian canyon (Clauzon and Cravatte, 1985; Clauzon et al., 1987a,b). The offshore extension of the Roussillon Basin is underlain by two deep reflectors (3 and 4.5 s TWTT) beneath the acoustic basement (Fig. 3). The deepest one cannot be traced with confidence because it is confused with the migration artifacts (smiles) and masked by the shallowest horizon that dips to the SE (from 3 to about 5 s TWTT below the shot point 1200) (Fig. 3). From previous studies (Mauffret and Gennesseaux, 1989; Gorini et al., 1991; Mascle et al., 1996) it is clear that these intra-basement reflectors are Eocene thrusts reactivated as extensional detachments during the Oligocene–early Miocene extension, although such reflectors have not yet been identified beneath the Easternmost Pyrenees (Albères). From the seismic profile LRM 02 (Fig. 3) it is evident that the formation of the Roussillon depression is related to the extensional motion along the upper detachment level during the formation of the Mediterranean Basin. The boundary between the Albères Massif and the Roussillon Basin is supposed to be a large normal fault, named the Tech Fault (Figs. 1 and 2), reactivated several times (Calvet, 1985, 1986; Clauzon et al., 1987b). According to Calvet (1985, 1986) the red beds described in the area of the Tech Fault could be lower Miocene continental beds dipping to the south as a result of the normal faulting. According to Clauzon et al. (1987a,b), who consider these layers to dip eastwards, these same beds could represent a Pliocene Gilbert-type fan delta. However, these contradictory hypotheses apply to the western part of the Roussillon Basin and may not be valid in the eastern part of the basin. Our seismic profile, located on the eastern part of the same basin, shows a moderate dip (about 7° towards the northwest, Fig. 3) of Miocene layers overlying the offshore Albères flank, although several faults ('equivalents' of the Tech Fault, Fig. 3) affect the offshore Albères flank and the overlying Miocene bed with a clear throw of about 300 m. Nevertheless, according to our present study, the northern boundary of this massif is not a fault similar to the Tech Fault. Therefore, the Albères Massif is not a horst between the Roussillon and the El Empordà Grabens (Calvet, 1985) but the hanging wall that has moved southwards along the detachment level. The Tech Fault may just be an antithetic fault affecting the rollover. As for the Miocene beds, they are gently tilted, with a throw of about 0.5 km, along with the Albères flank and subsequently affected by the Messinian unconformity (Fig. 3), with more than 1 km erosion on top of the Albères Massif. This massif crops out at sea level very close to the coast and rises on land (Calvet, 1985, 1986) to an altitude of almost 1.5 km.

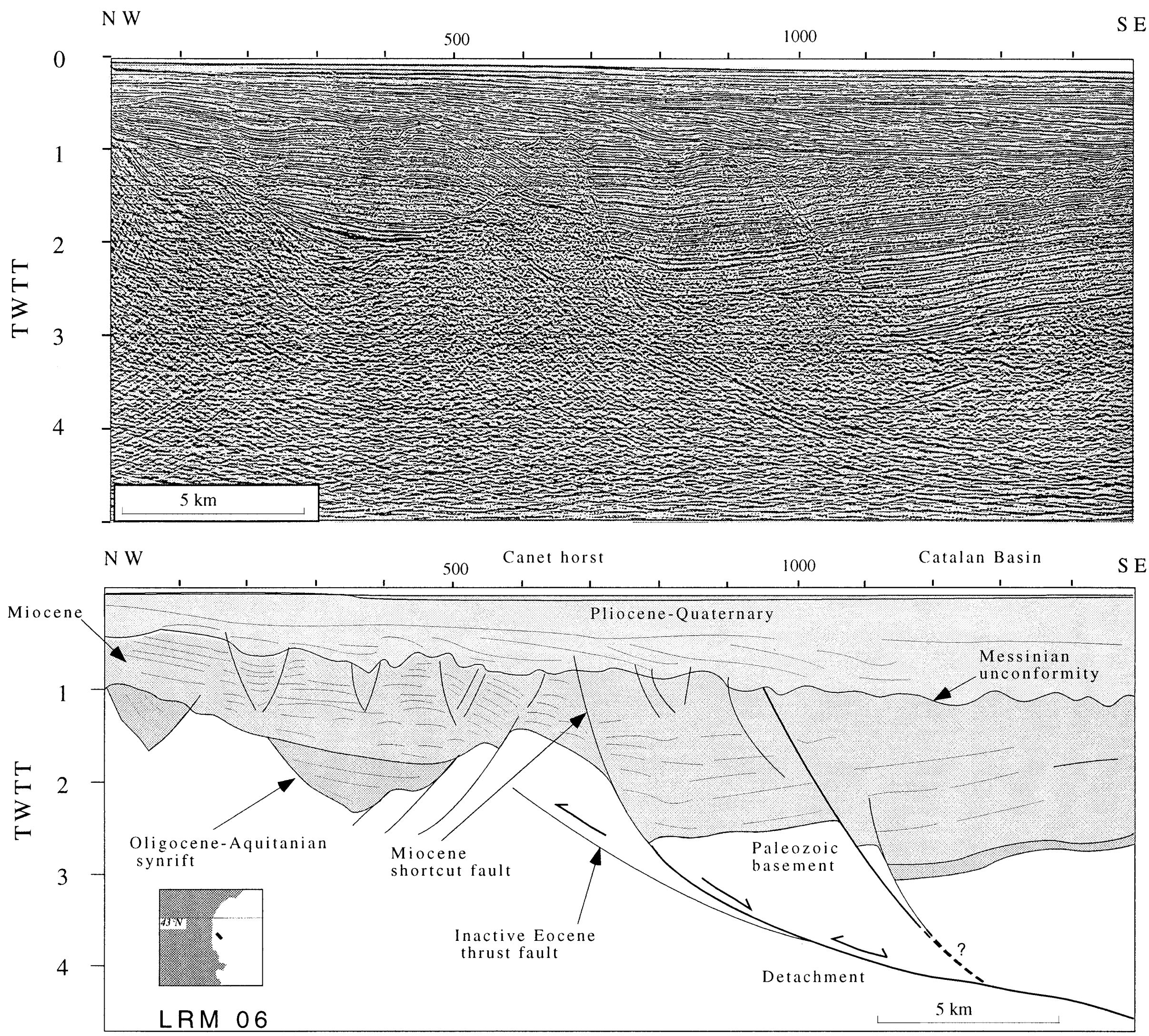


Fig. 6. Seismic profile LRM 06. Location in the inset and Fig. 1. The deep detachment observed on the profile 04 (Fig. 4) limits a horst and disturbs the Miocene layers. An antithetic fault shows a rotational listric motion in a small graben. Note the fan-shaped synrift formation that indicates that a first displacement occurred during the early Miocene. Therefore, the late Miocene rotation is a reactivation of a first motion. The detachment is in part inactive beneath the Canet horst and a late Miocene listric fault (shortcut) soles out into the deep detachment.

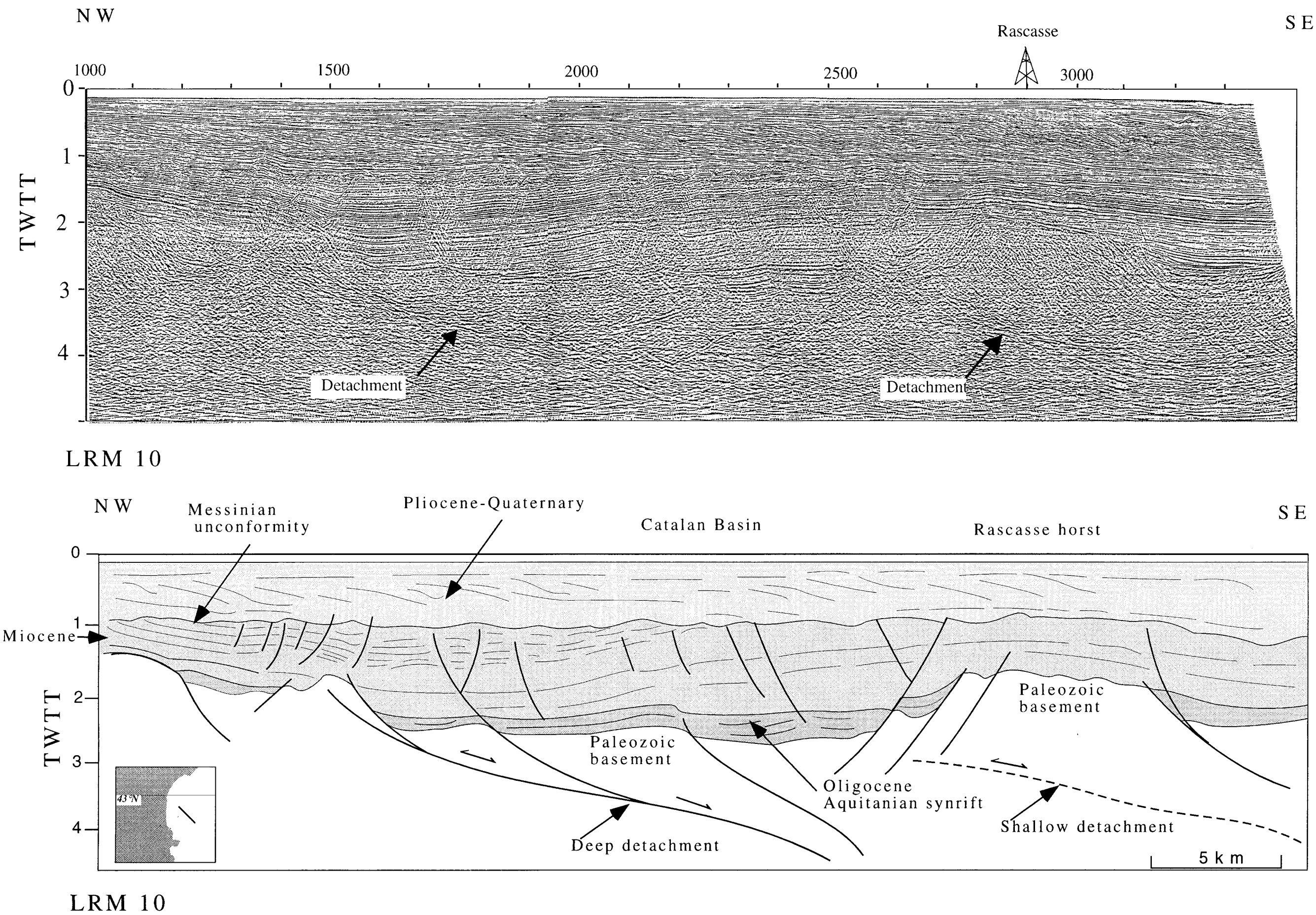


Fig. 7. Seismic profile LRM10. Location in the inset and Fig.1. This profile crosses the Catalan Basin and the horst where the Rascasse well has been drilled. A deep detachment in the basement has been confused (Mauffret and Gennesseaux, 1989; Gorini et al., 1993) with the acoustic basement covered by a thick synrift formation. A second shallower detachment is observed beneath the Rascasse well (Mauffret and Gennesseaux, 1989; Guennoc et al., 1994). Note the prominent late Miocene faults near the Rascasse horst.

Southeast of the Albères Massif a new basin is developed on the continental shelf before the offshore extension of the Cap de Creus high (Fig. 3).

The seismic profile LRM 04 (Fig. 4) clearly shows the evolution of the offshore area. The chronostratigraphy of the exploratory well Canet 1 has been revised (Clauzon and Cravatte, 1985) for the Pliocene section and is much more precise than the previous identification (Gottis, 1958). The Messinian unconformity, drilled at 825 m depth, is placed between marine early Pliocene (Tabianian) marls and marine Langhian marls. The hiatus of the Serravallian and Tortonian layers corresponds to the Messinian erosional unconformity. To the layers between 825 and 1716 m, a Burdigalian–Aquitanian to Oligocene age was attributed. Between 1716 and 1785 m, a Permian Triassic layers overlies the metamorphic basement (Gottis, 1958). A siliceous mylonitic breccia at this level suggests a faulted contact between the sedimentary cover and the basement. The depths of the Messinian and the acoustic basement on the seismic profile are about 1 and 2 km, respectively (Fig. 4), which fits with the information of the Canet well, although the so-called Aquitanian–Oligocene layer in the well corresponds to the Miocene on the seismic profile LRM 04 (Fig. 4). Projection of the Canet 1 on the seismic profile LRM 04 suggests that the well is located at the northwestern shallow boundary of the Roussillon–Catalan Basin. More precisely (see later), the borehole may be placed (Fig. 2) just above the eastern extension of the North Pyrenean Fault concealed by the recent sediments of the Roussillon Basin. The Roussillon–Catalan Basin is bounded by a detachment fault that flattens deep into the acoustic basement (Fig. 4). If the main extensional event is related to the early Miocene opening of the Mediterranean Sea, it is evident (Fig. 4) that this fault was reactivated and offsets the Miocene, the Messinian unconformity and the early Pliocene. The onlap of the late Pliocene and Quaternary indicates that this fault is now inactive. However, in the early Pliocene, the throw along this fault was small and most of the faults shown at the right of Fig. 4 are sealed by the Messinian unconformity. The Albères Massif is absent in this profile that shows thick Miocene beds tilted and dipping northwestwards (Fig. 4). It is evident that the tilting of the basement and the Miocene beds is related to the detachment fault activity and that the Cap de Creus high area forms its hanging wall. Furthermore,

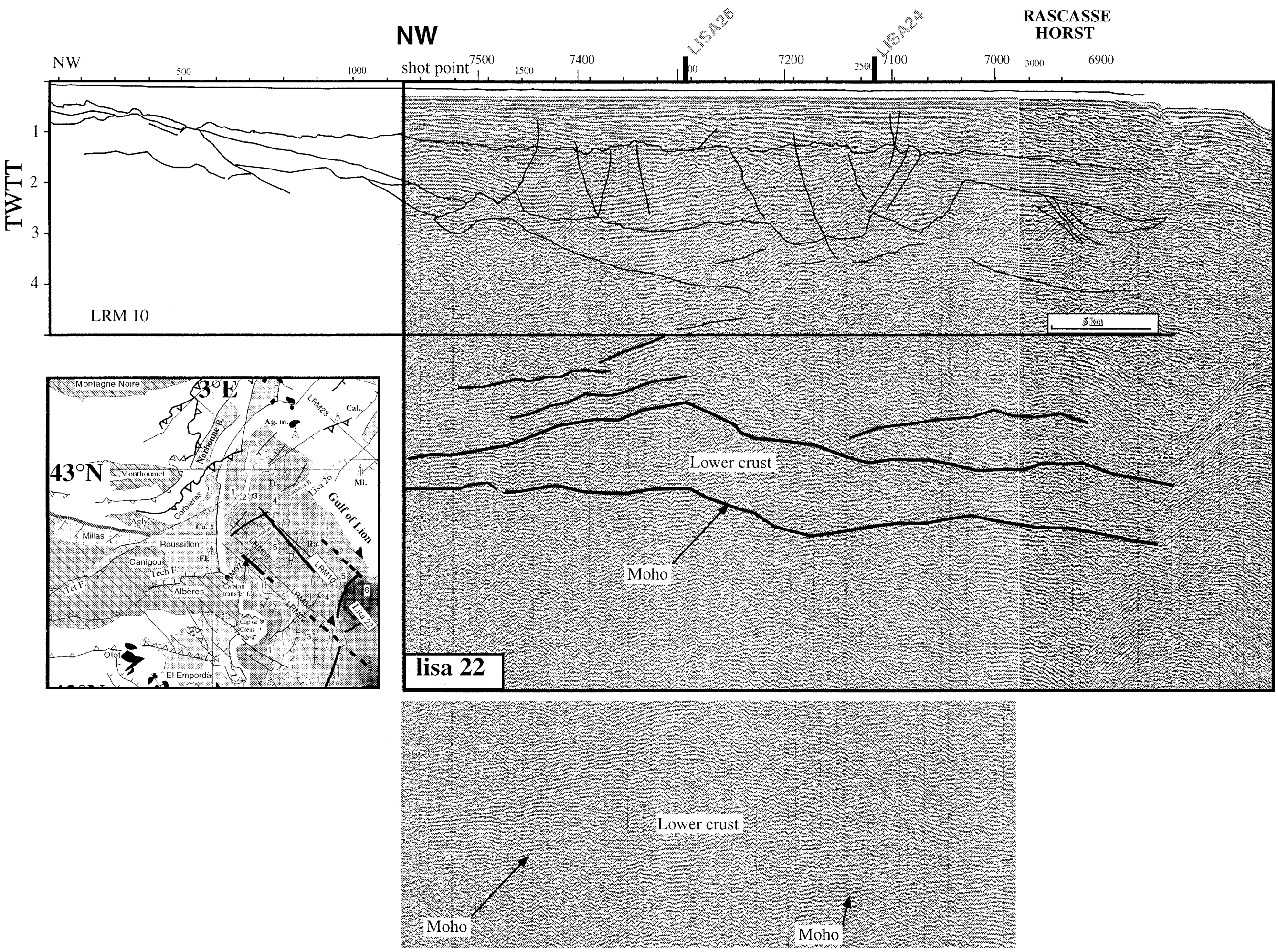


Fig. 8. Seismic profile LISA 22. Location in the inset and Fig. 1. This profile is not accurate for the shallow structure and is completed by the result of LRM 10 (Fig. 7). In addition, we correlated the profile with all the existing lines particularly with other LISA seismic lines (LISA 24 and 26). The results are good for the deep parts (below 5 s TWTT). This profile shows a prominent rise of the Moho and the lower crust below the Catalan Basin.

it is obvious that the extension is more important in the Gulf of Lion than in the Roussillon Basin and this increase in the extensional activity has been linked to the Catalan transfer zone (Lefebvre, 1981; Gorini, 1994; Gorini et al., 1994; Mauffret et al., 1995). This right-lateral fault is located between LRM 02 and LRM 04 and consequently the Albères Massif may have been cut by this fault and transported into the Cap de Creus high. In this area, the maximum of the Messinian erosion is observed and the faults are reactivated during the late Miocene, whereas the early Pliocene is unaffected and onlaps the Messinian unconformity (Fig. 4).

The seismic line LRM O7 illustrates the Catalan transfer zone and the absence of the Albères Massif in the basin (Fig. 5). The acoustic basement in the Catalan transfer zone area is not a steep feature but shows a listric geometry. A close examination of this seismic profile suggests that the upper crust beneath the acoustic basement is crossed by a relatively steep fault that soles out into a deep detachment (Fig. 5). This fault corresponds to the Catalan transfer zone and wide angle seismic profiles, performed during the LISA experiment, reveal that a thickening of the crust towards the west begins beneath this transfer zone (Nercessian et al., 2001; Vidal et al., 2001). The basement and the overlying Miocene beds show a prominent 30° southeastwards dip and a coeval Messinian erosion. The Miocene tilting on the northern flank (about 0.5 km, LRM 02, Fig. 3) and on the eastern flank (about 1 km, LRM 07, Fig. 5) of the Albères Massif suggests a late Miocene uplift illustrated by the inclination of Miocene reflectors. A projection of the Miocene beds through the erosion surface allows us to estimate a large (0.7 km) erosion of the Miocene sequence that followed, or is contemporaneous with, the uplift. This erosion is related to the major sea level drop during the Messinian salinity crisis. The uplift of the offshore Albères Massif is coeval with the formation of the NE–SW trending normal faults shown by the seismic survey (Figs. 3 and 4). This extensional tectonics occurred during the late Miocene and persisted in some areas until the lower Pliocene (Fig. 3). These facts fit very well with the conclusions of a recent

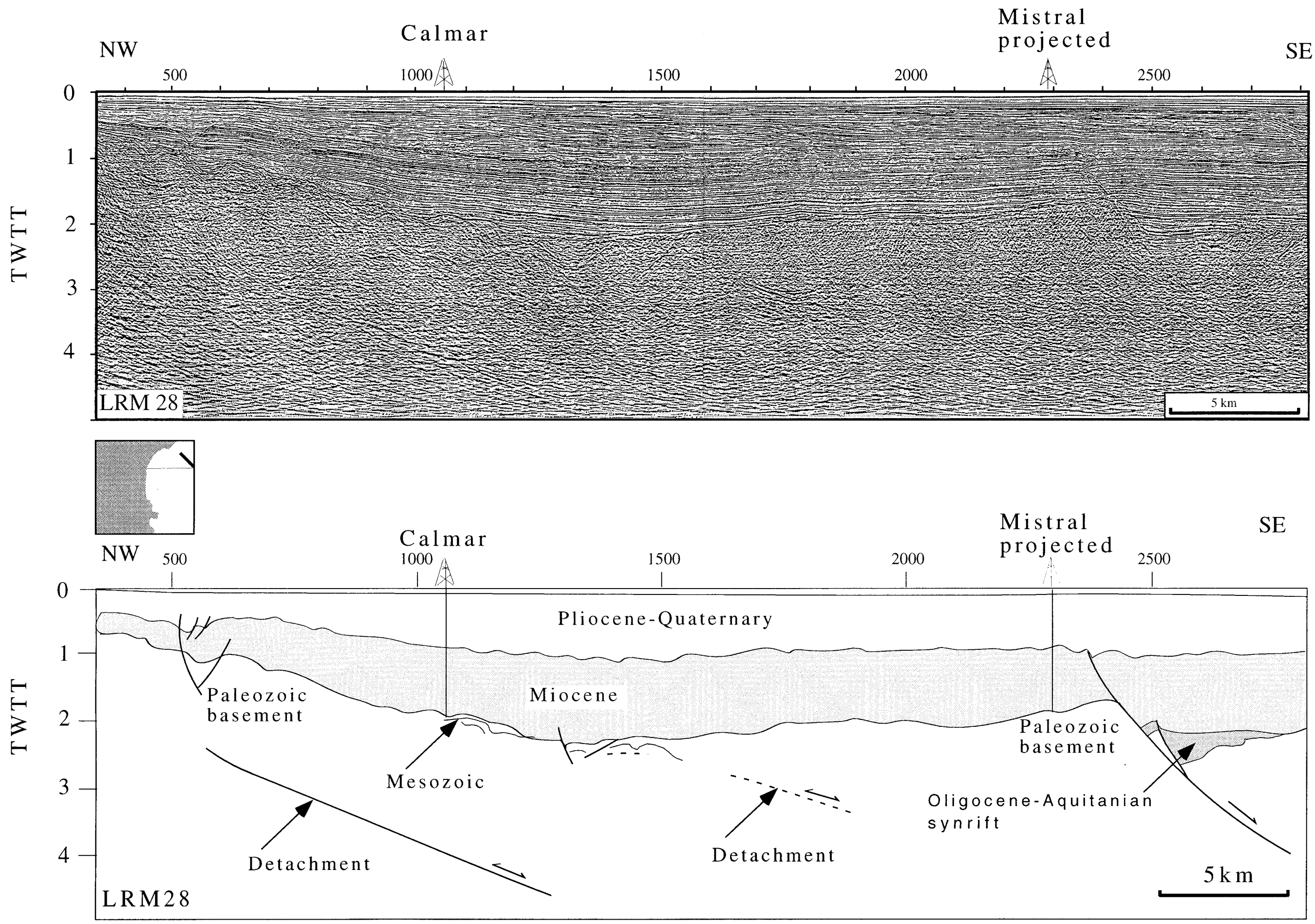


Fig. 9. Seismic profile LRM 28. Location in the inset and Fig. 1. This profile is outside of the detailed study but shows different levels of detachment faults in the area where the Calmar borehole has been drilled. A prominent late Miocene fault that bounds the Mistral horst is illustrated. This observation indicates that the late Miocene extension is not restricted to the southwest of the Gulf of Lion but affected the entire region.

structural activity of the Albères Massif, as proposed by Calvet (1985). However, the geometry of the massif uplift is not simple because the Tech Fault and the Catalan transfer zone are at right angles to each other. We observe the same tilt of the sedimentary beds as the Albères Massif and not a dip towards the faults, as would be expected in the case of a normal fault. A tilt along an uplift conforms more to a ramp and flat geometry (left lower inset, Fig. 5) than a normal extensional fault characterized by a fan-shaped geometry and a dip of the synrift beds towards the footwall (Benedicto et al., 1996, 1999; Mauffret et al., 1999). If the Albères Massif is effectively the hanging wall of the Roussillon Basin, it is probably the footwall of the El Empordà Basin (Fig. 2) and the right-lateral motion of the Catalan transfer zone that may also contribute to the uplift of the Albères Massif by thinning of the crustal load in the Gulf of Lion. In fact the crust is thinned on three sides of the Albères Massif and this particular situation explains the almost circular uplift of the massif. The crossing between LRM 07 (Fig. 5), LRM 02 (Fig. 3) and LRM 04 (Fig. 4) suggests that the detachment level identified on these profiles is here (Fig. 5) almost horizontal because the seismic profile is a strike section almost normal to the dip from the NW to the SE of the detachment (Fig. 5). Towards the southeast, the identified detachment level dips and a new shallower level can be seen, whereas the deeper detachment is concealed and disappears beneath the shot point 1000 (Fig. 5).

In Fig. 6 we can observe that a deep detachment at the northwestern limit of the Catalan Basin (shot point 700, Fig. 4) has a listric shape. It penetrates deep into the crust but was reactivated during the late Miocene along a steep listric fault that bounds the Canet horst (Fig. 6, shot point 700). We observe a northwest extension of the detachment beneath the horst that has been inactive during the extension because neither the basement of the horst nor the sedimentary strata of the graben are displaced where the detachment meets the acoustic basement (shot point 500, Fig. 6). The inactive part of the detachment between shot points 500 and 1000 (Fig. 6) is clearly an Eocene compressional detachment. The deep part of the detachment has been reutilized during the extension. The fault that bounds the early Miocene Canet horst and has been reactivated during the late Miocene (shot point 700; Fig. 6) is a shortcut that soles out into the detachment. This prominent fault that limits the Catalan Basin is the

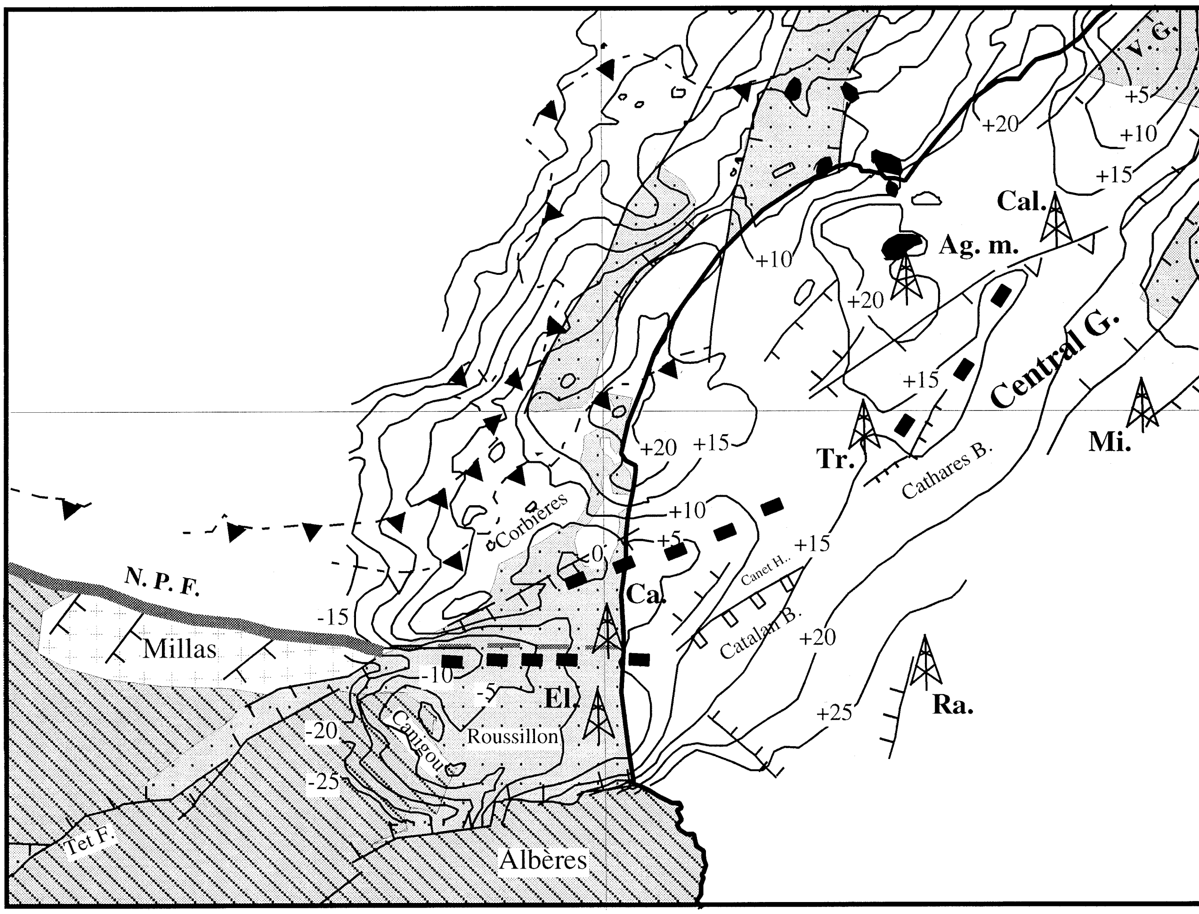


Fig. 10. Bouguer gravity map (Arthaud et al., 1981). The deepening of the Moho (see Fig. 1) beneath the Canigou Massif is well illustrated by a strong negative gradient. An E–W gradient in the Roussillon Basin may correspond to the buried North Pyrenean Fault (N.P.F.). A small gravity low (0 milligals), north of the Canet borehole, may be correlated with the graben north of the Canet horst shown in Fig. 6. This low (0 milligals) may be correlated with the minimum in the central Gulf of Lion (minus black solid line surrounded by the 15 milligals contour) and the Vistrenque Graben (V.G.) where the minimum value is 5 milligals. The Corbières Massif and the Palaeozoic high, where the Agde Maritime (Ag. m.) has been drilled, are correlated with a maximum of 20 milligals.

apparent prolongation of the North Pyrenean Fault, which is probably located beneath the Canet well in the Roussillon Basin (Fig. 2, see later). Another fault around the shotpoint 1000 shows a throw in the Miocene but the difference in thickness of these layers across the fault suggests that the first activity was coeval with the extensional formation of the basin in the early Miocene. A typical listric geometry is observed at the shotpoint 600 (Fig. 6) beneath the Messinian but this time the throw is opposite (towards the northwest) to the large faults previously described in this profile. Again the fault has been reactivated up to the Messinian but its formation is related to Oligocene–Aquitanian extension.

The Rascasse well reached the Paleozoic basement on the top of a block (Gorini et al., 1993). A detachment beneath this block has been identified (Fig. 7) and described in several publications (Mauffret and Gennesseaux, 1989; Gorini, 1994; Gorini et al., 1994; Guennoc et al., 1994). However, a major mistake must be noted in these publications because the deep detachment in the Catalan Basin has been confused with the acoustic basement. The previous seismic profiles in the Catalan Basin were not clear and the chaotic body beneath the well-bedded Miocene was misinterpreted as a synrift formation. Now it is evident that the chaotic zone belongs to the acoustic basement overlying the deep detachment. The late Miocene activity of the faults is clearly observed, particularly near the Rascasse horst (Fig. 7).

The LISA 22 seismic profile (Fig. 8) has been correlated with all the data in this area and particularly with other LISA profiles (LISA 24 and 26) that clearly show the deep structure down to the Moho (Nercessian et al., 2001; Vidal et al., 2001). On the profile LISA 22, the correlation with the previous seismic line (LRM 10, Fig. 7) shows that the upper part of LISA 22 is poorly processed because the refracted arrivals severely damaged the reflections that have been muted. In contrast, the image is much better below 5 s TWTT and a lower reflective crust is identified from 6 to 8 s TWTT and a Moho reflection between 8 and 9 s TWTT. The Moho is shallower (8 s TWTT) beneath the Catalan Basin and is deeper (9 s TWTT) below the Rascasse horst, which is clearly a rollover (Fig. 8).

The observations made in the Catalan Basin can be

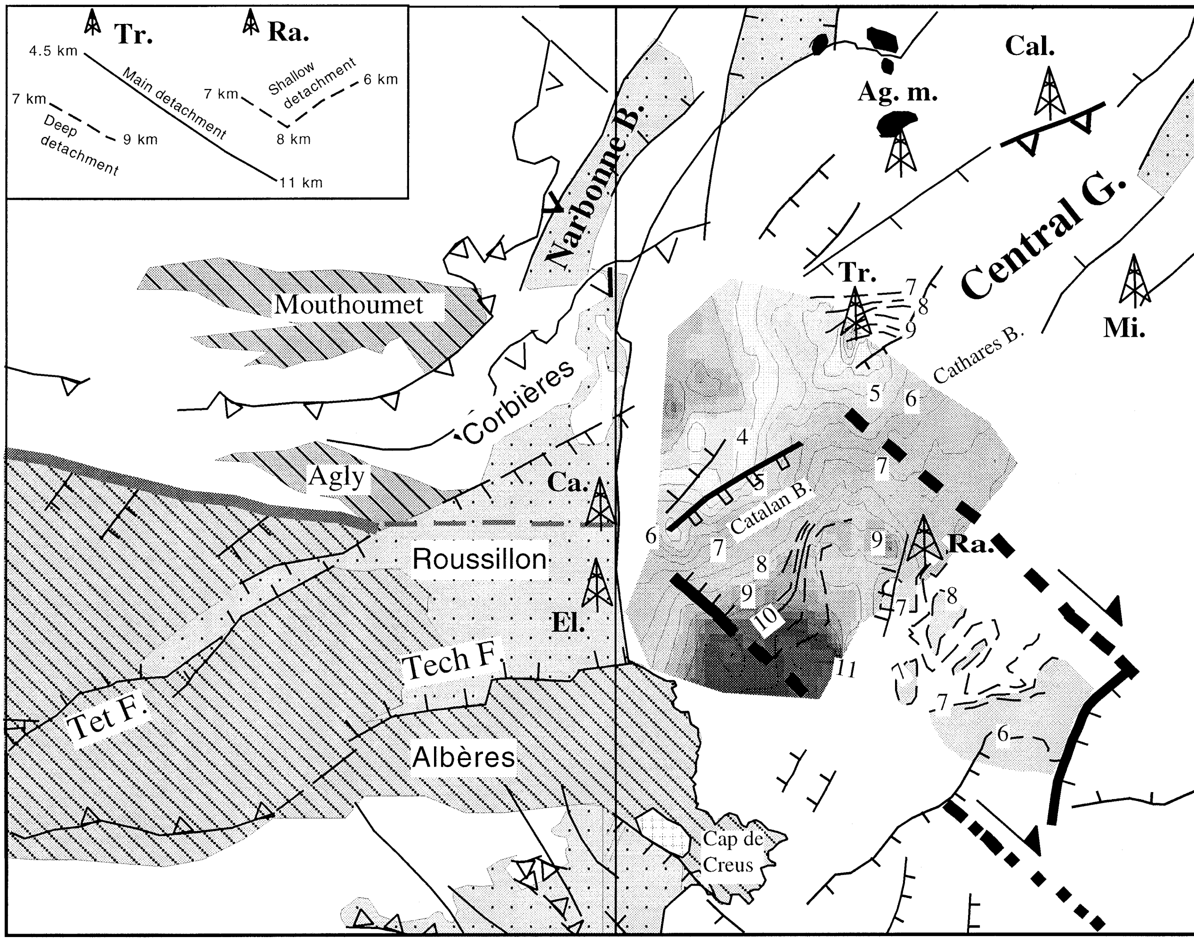


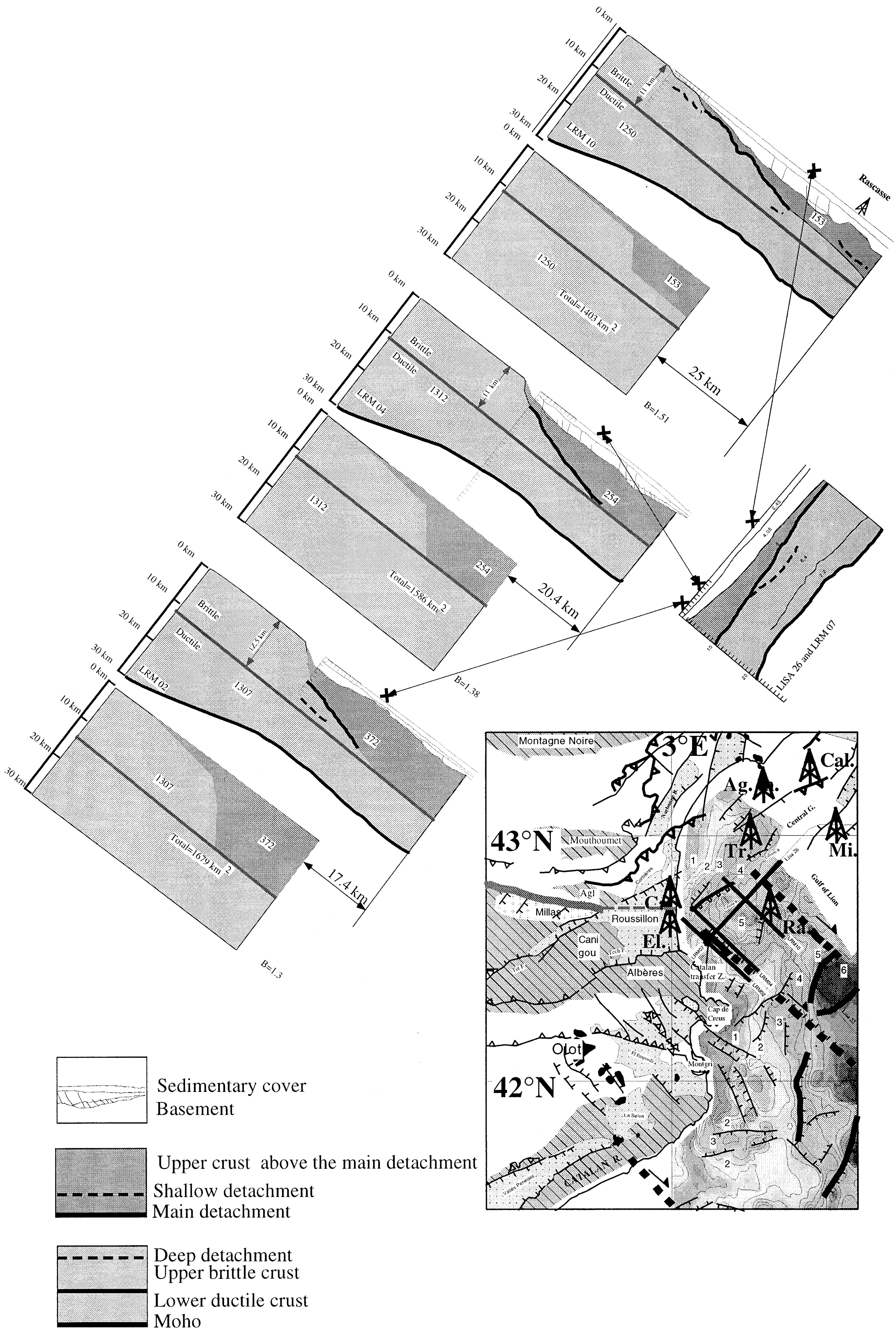
Fig. 11. Depth-to-detachments. In the southwest of the Gulf of Lion a detachment can be followed in the basement as deep as 11 km. This deep detachment is concealed progressively towards the northeast by a shallower one (see Fig. 5). The other detachments, below the Tramontane and Rascasse wells, are represented by dashed lines (see the inset). Note the dip of the mapped detachment that is normal to the Corbières Massif. This geometry suggests that the detachment was a reverse deep thrust during the Eocene Pyrenean Orogeny that have been subsequently reactivated in extension during the Oligocene–early Miocene formation of the Catalan Basin.

extended to the entire Gulf of Lion. An Eocene thrust sheet formed of Mesozoic rocks has been reached by the Calmar well (Gorini et al., 1993; Gorini, 1994). A southeastwards-dipping detachment is observed below the well (Fig. 9). A shallower detachment is also evident beneath the basin. The Mistral horst, where the Paleozoic has been drilled (Cravatte et al., 1974), is bounded by a large fault that has been obviously reactivated up to the Messinian.

4. Discussion

4.1. Moho shape and gravity maps

The shape of the Moho is derived from the refraction studies on land (Gallart et al., 1980, 1981; Daignieres et al., 1981, 1982) and at sea (Nercessian et al., 2001; Vidal et al., 2001) and a complete description of the results is beyond the scope of this study. We have already discussed the eastwards thinning of the crust that may be related to a possible right-lateral transfer of the Eocene compression from the eastern Pyrenees to Calabria (Rehault et al., 1984; Mauffret and Gorini, 1996; Olivet, 1996) and certainly from the extensional formation of the Gulf of Lion during the Oligocene–early Miocene. On land, this thinning is located to the north along the North Pyrenean Fault (Fig. 1) with a step of about 10 km (38–28 km). Towards the east, the thinning shows a NNW trend and is parallel to the western boundary of the Roussillon Basin (Fig. 1). The 2787-m-high Canigou Massif is related to this thinning and has probably a footwall position relative to the Roussillon Basin and an isostatic rebound of the footwall may explain this elevation. The Albères Massif and the Cap de Creus promontory correspond to a thickening of the crust. The offshore data show that this thickening extends to the Rascasse high and that the Catalan Basin is underlain by a moderately thin crust (Fig. 1) but well delineated with a subcircular shape. The Bouguer gravity map (Fig. 10)



LRM 10
LRM 04
LRM 02
Brittle
Ductile
Rascasse
25 km
20.4 km
17.4 km
Total=1403 km 2
Total=1586 km 2
Total=1679 km 2
B=1.51
B=1.38
B=1.3
LISA 26 and LRM 07
Montagne Noire
3°E
43°N
42°N
Cal.
Ag.
Tr.
Mi.
Mouthoumet
Central G.
Gulf of Lion
Roussillon
Millas
Agl
El.
Ra.
Cani gou
Albères
Catalan transfer Z.
Cap de Creus
Olot
Montgri
La Selva
Sedimentary cover
Basement
Upper crust above the main detachment
Shallow detachment
Main detachment
Deep detachment
Upper brittle crust
Lower ductile crust
Moho

(Arthaud et al., 1981) shows negative values perfectly correlated to the deepening of the Moho beneath the Axial Pyrenean Range (Canigou Massif) with the same N–S to NNE–SSW trend as the Moho map (Fig. 1). The E–W trend in the northern Roussillon Basin is probably linked to the extension of the North Pyrenean Fault and the Canet may well have been drilled on the footwall of the fault or a horst that limits the Catalan Basin on the continental shelf (Fig. 6). North of this area, a small gravity low (0 milligal, Fig. 10) may be related to a basin that has an eastern offshore extension (Fig. 6). Therefore, we suggest that the northern boundary of the Roussillon and Catalan Basins is not the Tet Fault (Figs. 1 and 2) but the North Pyrenean Fault and its offshore eastern extension. The relative rise of the gravity values towards the north from 0 to 20 milligals (Fig. 10) is related to the Corbières that has been thrusted to the north and thickened during the Eocene Pyrenenan Orogeny, and then subsequently thinned by extension during the rifting in the Gulf of Lion (Gorini et al., 1991; Mascle et al., 1996). The maximum value of the gravity high (20 milligals) corresponds to a Paleozoic slice involved into a thrust (Viallard and Gorini, 1994). The relative gravity high (20–25 milligals) extends offshore beneath the Agde Maritime well (Ag. m., Fig. 10) where the Paleozoic has been drilled (Gorini et al., 1994). This high is flanked to the south by a gravity low surrounded by the 15 milligal contour (Fig. 10). This low is located along the strike of the Vistrenque Graben (V. G., Fig. 10) but in the central part of the Gulf of Lion it delineates a small graben located north of the Tramontane well (Tr., Fig. 10) and in the study area north of the Canet horst (Fig. 6) and not the Central Graben as supposed by Gueguen (1995) on a map derived from altimetry. We confirm the observations made by Guennoc et al. (1994) that, in the Central Graben, the negative effect on the gravity of the deep sedimentary basin is compensated for by a positive effect of the Moho rising. The North Pyrenean Fault is not expressed offshore in the gravity map but the aeromagnetic maps (Galdeano and Rossignol, 1977; Guennoc et al., 1994) show an E–W trend north of Rascasse high that suggests (see later) that this structure is limited by the fault. A complete study of the North Pyrenean Fault in the offshore area has already been presented (Séranne et al., 1995; Mauffret and Gorini, 1996) and is beyond the scope of this paper.

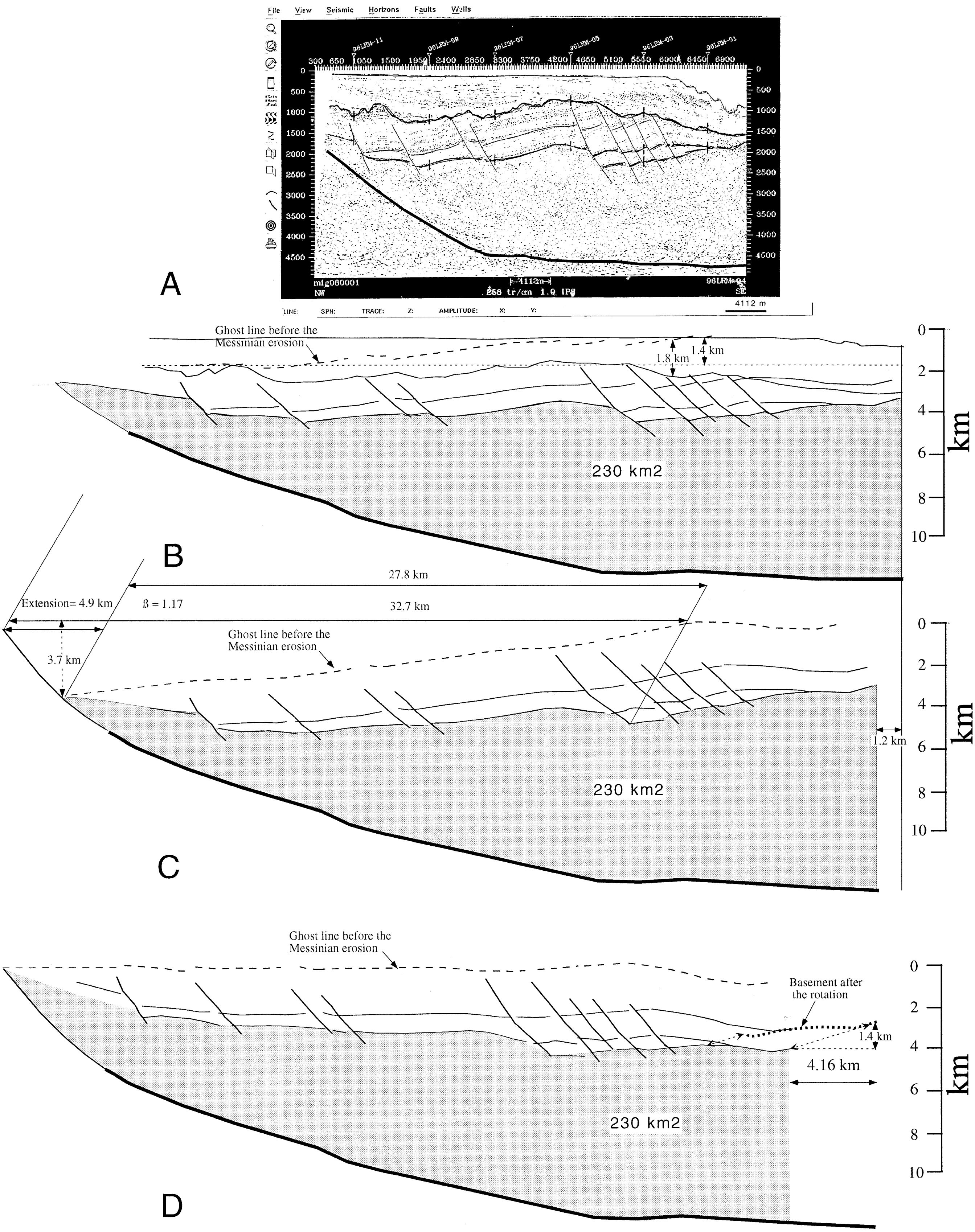
4.2. Detachments

In the Corbières Massif, which shows a prominent virgation of the Eocene Pyrenean structures and a NE–SW trend, it is now very well known that the previous Eocene thrusts have been reactivated in extension along detachment faults (Gorini et al., 1991; Mascle et al., 1996). On land, these faults use the Triassic evaporites in the shallowest part of the detachment, but they progressively cut the Paleozoic basement towards the southeast in the offshore area (Mauffret and Gennesseaux, 1989; Séranne et al., 1995; Benedicto et al., 1996, 1999; Séranne, 1999). However, the Paleozoic basement on land may already be affected by deep detachments (Viallard and Gorini, 1994). Offshore detachments have been described beneath the Tramontane well (Gorini et al., 1994), the Rascasse well (Guennoc et al., 1994) and beneath the eastern Gulf of Lion (Gorini et al., 1993). However, the LRM profiles clearly show new detachment faults never described in the southwestern part of the Gulf of Lion and sometimes confused with the acoustic basement. Moreover, several detachment faults have been identified at different levels. The Rascasse deep detachment previously described between 8 and 6 km (Fig. 11, inset) is placed above the new imaged feature (Figs. 7, 8 and 11), whereas the opposite situation is observed beneath the Tramontane well where the previously identified detachment is deeper (between 7 and 9 km; Fig. 11, inset) than the new feature contoured in Fig. 11 (4.5 km, inset). In the basin, the upper detachment may conceal a deeper one (Fig. 5) and it is clear that we map (Fig. 11) the deepest detachment in the southwest of the Gulf of Lion and, in the northeast of the study area, a shallower feature that overlies and conceals the deeper detachment. The depth map (Fig. 11) shows the dip to the southeast of the deepest detachment and a NE–SW trend parallel to the Corbières Massif. The maximum depth is 11 km that corresponds to the lower limit of the upper brittle crust as determined by the depth of the earthquakes, whereas some epicenters are located in the lower ductile crust (11–32 km; Souriau and Granet, 1995). In our map (Fig. 11) the detachments are shallower towards the northeast and terminate at a listric fault that cuts the sedimentary cover up to the Messinian (Fig. 6). This shortcut merges with the detachment that is a former thrust inactivated in the upper part beneath the Canet horst (Fig. 6). This ramp, before the late Miocene reactivation, may have been connected to a flat along the Paleozoic basement before reaching the flat and ramp geometry of the Corbières Triassic detachment level (Gorini et al., 1991; Mascle et al., 1996) or a deeper level in the Paleozoic basement (Viallard and Gorini, 1994).

4.3. Balanced sections and stretching factor

To generate the crustal sections (Fig. 12) we use the refraction studies for the velocities, the LISA seismic profiles for the deep part (5 to 9 s TWTT) and the LRM profiles for the shallower structures (0 to 5 s TWTT). Because of the choice of the LRM lines, that are the best seismic sections, we do not have a complete picture of the

Fig. 12. Balanced cross-section at crustal scale. Location given in Fig. 1 and in the inset. The stretching factor β varies from 1.3 (in SW) to 1.51 (in NE). LISA 26, correlated with LRM 07, shows the thinning of the crust beneath the Catalan Basin. The different detachments and the boundary between the ductile and brittle crusts are drawn.


A
File View Seismic Horizons Faults Wells
mig080001
NW
SE
4112 m
B
Ghost line before the Messinian erosion
1.4 km
1.8 km
230 km2
km
C
Extension= 4.9 km
ß = 1.17
27.8 km
32.7 km
3.7 km
Ghost line before the Messinian erosion
1.2 km
230 km2
D
Ghost line before the Messinian erosion
Basement after the rotation
1.4 km
4.16 km
230 km2

margin in its southeastern part, towards the deep oceanic domain. Nevertheless, for comparison, we decided to draw the same length for the three presented transverse cross-sections (Fig. 12). Consequently our sections cannot be completely balanced because the seaward parts of the roll-over structures (Cap de Creus and Rascasse highs) are missing. Nevertheless, our restorations are very suggestive and based on a complete data set. In particular the seismic profile LISA 26 has been correlated with LRM 07 and the on-land refraction and wide reflection records of the shots fired during the LISA experiment (Nercessian et al., 2001; Vidal et al., 2001). From this profile (Fig. 12) it is evident that the Catalan Basin is underlain by a Moho high as shown by the Moho depth (21 km, Fig. 1). The profile LRM 07 also shows (Figs. 5 and 12) two different detachment levels that we contoured in Fig. 11, a deep one in the southwest and a shallow one in the northeast. It is quite possible that these different reflectors may be related to duplexes (Saula et al., 1994) formed during the Eocene compression but we do not have enough clear data yet to confirm this hypothesis. A similar geometry has been described in the Valencia Trough where the detachments have been reactivated several times (Sàbat et al., 1997; Verges and Sàbat, 1999). In the LRM 02 profile, the transition between ductile and brittle crust was adjusted to 12.5 km to fit with the depth of the observed detachment. In the others profiles (LRM 04 and 10) the transition has been fixed at 11 km, i.e. the depth observed in the seismicity data (Souriau and Granet, 1995). However, we know that the depth of the transition between the ductile and the brittle crust may vary between 11 and 15 km. Moreover, a depth of 15 km is deduced from the LRM 02 pre-extensional reconstruction (Fig. 12). In this reconstruction we balanced the area of the entire crust from the basement to the Moho, the lower crust below the main detachment and the upper crust above this detachment. We can see in Fig. 12 the other detachments (dashed lines) above the main detachment (LRM 10) and beneath (LRM 02, LISA 26 and LRM 07). In Fig. 12, to respect the area of the upper crust overlying the detachment, we draw pre-extensional ramps much steeper than the extensional listric faults. We do not know if these steep, probable Eocene, thrusts are real or drawing artifacts. If the steep compressional ramps are real, the Oligocene extension may use new shortcut faults in the upper crust. The stretching factor along the NW–SE trending seismic lines increases from $\beta = 1.3$ (LRM 02) to $\beta = 1.51$ (LRM 10). A displacement of 25 km on LRM 10 juxtaposes the Rascasse structure and the fault that bounds to the northwest the Catalan Basin and closes this basin. This value is more important than the extension in the Corbières Massif (less than 10 km) but equivalent to the Eocene compressional motion towards the north (about 25 km) (Mascle et al., 1996). From these different estimates, it can be deduced that the Corbières Massif was located in the Catalan Basin before the Pyrenean compression and the Rascasse structure placed behind it towards the southeast. During the Pyrenean Orogeny the Corbières was transported on land and the Rascasse structure moved into the Catalan Basin. Moreover, the thrusted Montgri terranes on the El Empordà coast (Figs. 1 and 12) probably comes from a Mesozoic basin located north of the Axial zone in the Gulf of Lion (Bilotte et al., 1979). During the extension, the Rascasse horst was displaced above deep detachments to its present position. From the estimates of compression (25 km) and extension (10 km) in the Corbières Massif, we can propose an initial position of the Rascasse structure southeast of its present location. In conclusion, the Catalan Basin may be a Mesozoic basin formed during the Jurassic extension or the Cretaceous pull-apart motion where the thick sedimentary beds have been expelled to the south and the north by compression then extension.

4.4. Late Miocene–early Pliocene extension

We calculated the total stretching factor along the NW–SE trending seismic lines (Fig. 12). Nevertheless, from the observations of seismic profiles it is quite clear that the extension occurred during two periods: the first event is coeval with the rifting in the Mediterranean region during the Oligocene–early Miocene, whereas the second one is late Miocene–early Pliocene in age. The total stretching factor results from the two extensional events. The youngest normal faults are particularly evident in the LRM 04 (Fig. 4) and we choose this profile to evaluate the late Miocene–early Pliocene extension (Fig. 13). The line drawing has been stretched to show a picture with no vertical exaggeration. The area of the upper detached crust obtained is slightly smaller (230 km^2) than the area calculated on drawing using a velocity function of the same section (254 km^2, Fig. 12). To show a complete picture of the margin, the level of detachment was extended towards the coast in Fig. 12 more than in Fig. 13 and this difference may explain the difference of area. However, we do not have a complete picture of the detachment in the upper level (left side of Fig. 13) and the left tip of the detached block was deformed to respect the slope of the detachment and the surface (compare C and D, Fig. 13). It is evident that the largest

Fig. 13. (A) LRM 04 seismic profile with line drawing superimposed. (B) Line drawing stretched and adjusted to a depth section with no vertical exaggeration. Dashed line: ghost of the top of the Miocene before the Messinian erosion that is assumed weak on the upper continental shelf and maximum (1.82 km) beneath the present shelf break. The uplift of the ghost line is estimated to be 1.4 km. (C) Restored shape of the ghost before the throw of the normal faults and estimation of the extension to 4.9 km with the Faure and Chermette (1989) method. The collapse of the tip of the block is estimated to be 3.7 km and the uplift of the top of the same block to be 1.4 km (compare (C) and (D)). (D) Initial position of the rotated block before the late Miocene extension with conservation of area (230 km^2). The extension is slightly less important (4.16 km) than with the first method (4.9 km, (C)). The dashed line represents the top of the basement block after the rotation (C).

Messinian erosion occurred above the present shelf break. To estimate this erosion we assumed that the upper Miocene sequence was constant in thickness before the erosion and we drew a ghost line (dashed line, Fig. 13) parallel to a Miocene horizon located below the Messinian erosional surface. An erosion of 1.8 km was deduced from this reconstruction (Fig. 13B) and a 1.4 km tilt can be evaluated from the beginning of the section to the end of the ghost line (Fig. 13B). Then we removed the offset caused by the Messinian normal faulting to restore the geological transect (Fig. 13C). The extension is moderate (1.2 km) because we draw relative steep normal faults in the Miocene, although some weak traces in the basement suggest that the faults are listric and sole out into the detachment level. However, the main observation about the extensional motion is the tilting of the Miocene sequence and the subsequent Messinian erosion. With the new geometry of the ghost line, the tilt is up to 3.7 km from the tip of the collapsed block to the top of the pre-Messinian ghost line (Fig. 13C). The uplift of the top of the block is estimated to be 1.4 km from the position after the rotation (Fig. 13C) and before (Fig. 13D). An evaluation of the extension by the method developed on the tilted blocks in extension (Faure and Chermette, 1989) indicates a value of 4.9 km and $\beta = 1.17$. If we rotate the hanging wall with a conservation of surface (230 km^2, Fig. 13) and a horizontal Miocene sedimentary cover (Fig. 13D) the distance is 4.16 km. It is now evident that the tilt of the Cap de Creus high is caused by a late Miocene–early Pliocene rotation of block along the deep detachment level. The maximum Messinian erosion on the shelf break is observed on the top of this tilted block.

4.5. Age of the tectonics

The formation of the Mediterranean Basin is well dated from the Oligocene–Aquitanian (early Miocene) by the exploratory wells and related seismic profiles that show the synrift sediments deposited in the half grabens (Cravatte et al., 1974; Gorini et al., 1993; Gorini, 1994). However, the synrift deposits are very thin and a prominent tectonic activity occurred during the late Miocene. Most of the faults are sealed by the Messinian unconformity but some extensional activity persisted into the early Pliocene (Fig. 4). This extensional tectonics is well correlated with the Catalan tectonic history (Saula et al., 1994; Tassone et al., 1996; Lewis et al., 2000). In the El Empordà Basin, volcanic activity is 7.5–10 Ma old (Tortonian) (Donville, 1973a), whereas in the La Selva depression, the main volcanic activity occurred at 7.7 Ma (Tortonian) and 5.1 Ma (Pliocene) (Donville, 1973b). However, in this closed valley, weak volcanic activity persisted until the late Pliocene (3.5 and 2 Ma) (Donville, 1973b). The tectonic activity in the Cerdanya Basin, probably related to the right-lateral strike-slip Tet Fault (Cabrera et al., 1988), occurred mainly during the middle to late Miocene time. It has been clearly demonstrated (Calvet, 1985) that the uplift of the Albères Massif and the correlated Messinian erosion occurred mainly during the late Miocene. North of the Millas granitic Massif the normal faulting is mainly 21–13 Ma old according to fossiliferous infilling of karstic fissures (Faillat et al., 1990). However, the Messinian erosion may have eroded the late Miocene filling and the cracks caused by the late Miocene extension. Moreover, there is a difficulty in dating the sedimentary rocks that are mainly continental azoic red beds formed by erosional activity (Calvet, 1985, 1986) in the Roussillon region. On another hand, a Pliocene normal faulting is proposed for a 1.8 km uplift of the Millas granitic Massif (Arthaud and Pistre, 1993). The detachment faults that limit the half grabens on land, in the north of the Gulf of Lion, have a Neogene activity, possibly as late as Pliocene–Quaternary (Raymond et al., 1994). In the Central Pyrenees, the Axial Zone has been uplifted and re-excavated by a 2–3 km exhumation since the middle Miocene (10–5 Ma) and perhaps to the present (Fitzgerald et al., 1999). Nevertheless, we assume that the extensional tectonics have the same age in the El Empordà Basin, the Central and the Eastern Pyrenees and in the Gulf of Lion, where (this study) the late Miocene–early Pliocene tectonics are evident. Therefore, the on-land geology, the offshore geophysics and borehole results together suggest extensional tectonics during the late Miocene–early Pliocene time. However, we can observe that in NE Spain, the normal faults are oriented NW–SE, whereas the faults have a NE–SW strike in SW France (Philip et al., 1992). The best explanation for the NW–SE trend is a right-lateral transfer between the Gulf of Lion and the Valencia Trough that trends NE–SW with two pulses of activity separated by a middle Miocene tectonic quiescence. It is assumed in some publications (Briais et al., 1990), that the extensional activity in the Eastern Pyrenees may have been active during the Quaternary up to the present, whereas a compressional environment with reverse faulting is preferred for the Quaternary epoch in others studies (Philip et al., 1992). For our study it is clear that the extension is active up to the early Pliocene but no later. However, the seismicity (Souriau and Pauchet, 1998; Goula et al., 1999) suggests a strong tectonic activity at the present day (Fig. 14). The NE–SW faults of El Empordà, where the volcanic injections are located (Olot, Fig. 14), are still active. The Tet and Tech Faults are not very active but several earthquakes are located in the Corbières folded region. It is evident that the Corbières Eocene compressional zone has been reactivated in negative inversion during the Oligocene–early Miocene extension (Gorini et al., 1993; Mascle et al., 1996) and the meandering of the rivers suggest that the present deformation is again compressional with positive inversion of the same structures first folded during the Eocene compression then extended during the subsequent extension (Genna et al., 1997). The focal mechanism of an earthquake in the Agly Massif indicates a strike-slip fault that trends E–W and compressional stress oriented NE–SW (Rigo et al., 1997; Pauchet et al., 1996). The

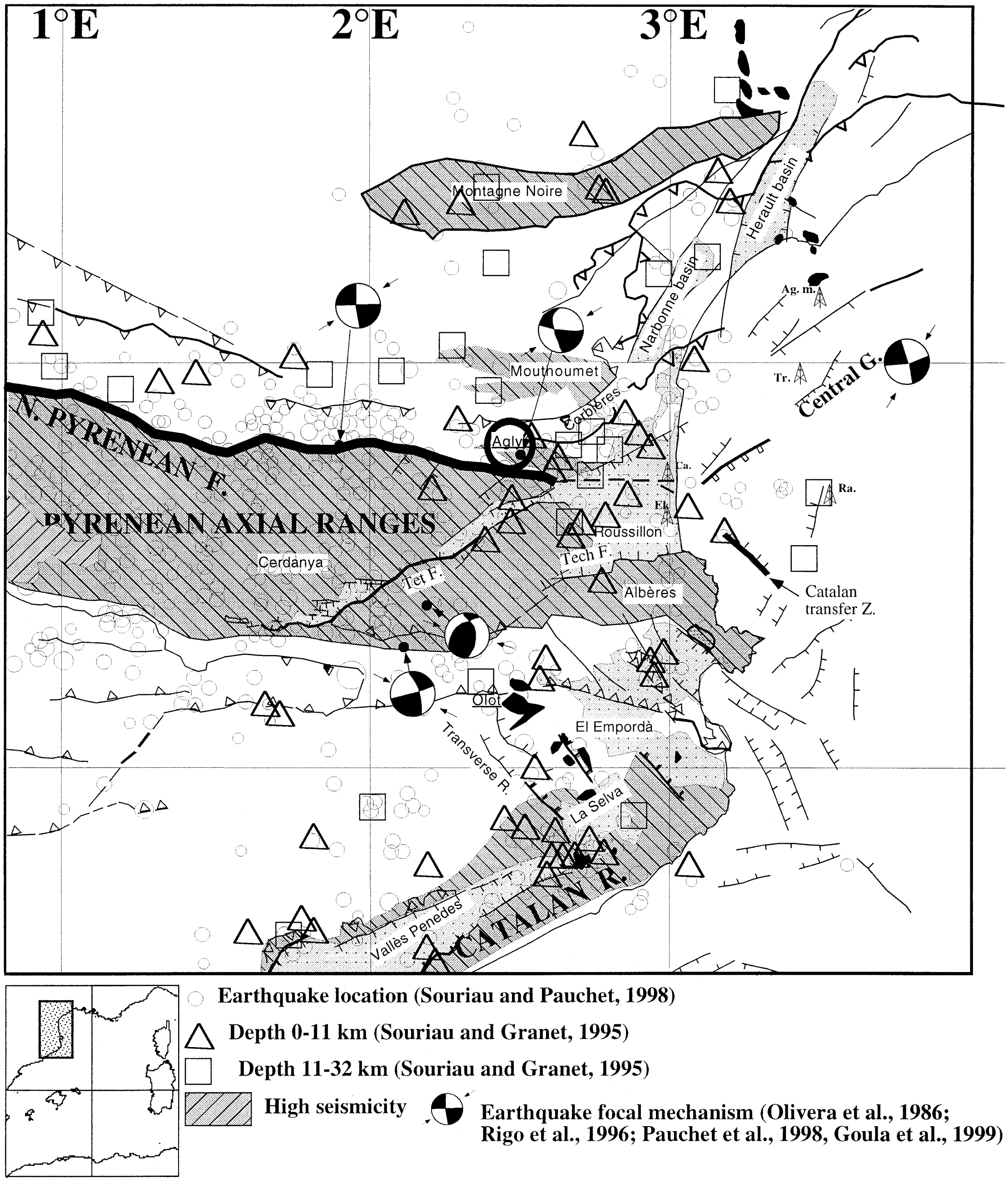


Fig. 14. Seismicity in the study area. Most of the earthquakes are located in the upper crust (11 km) but some are deeper in the lower ductile crust (11–32 km). Two earthquakes of this type are localized below the Rascasse horst. The Agly earthquake corresponds to a fault parallel to and with the same focal mechanism as the North Pyrenean Fault (Rigo et al., 1997; Pauchet et al., 1996). Note the major seismicity below the Corbières Massif and the transverse structures (transfer faults) in El Empordà and in the Catalan Range between the Vallès Penedes and la Selva depression. Note the relationship between the seismicity in the Catalan Range, the trend of the Transverse Range and the Olot volcanism. The focal mechanisms determined by Olivera et al. (1986) are compatible with a NNW–SSE right-lateral fault.

same mechanism has been determined (Olivera et al., 1986) along the North Pyrenean Fault (Fig. 14). In the Gulf of Lion, the same NE–SW compressional stress was also determined (Goula et al., 1999) by the focal mechanism of an earthquake located near the Mistral well (Fig. 14). The NE–SW trend shown in this region contrasts with the NW–SE orientation of the stress calculated from focal mechanisms in the Western Pyrenees and in Spain, south of the Pyrenean Axial Ranges (Fig. 14; Olivera et al., 1986; Goula et al., 1999). However, the contemporary stress deduced from the borehole breakout analysis shows again a NE–SW trend on the Catalan margin (Jurado and Müller, 1997). This discrepancy between the NW–SE stress that affects the Iberian Peninsula and most of the European region, and the NE–SW trend north of the Pyrenean Axial Range and on the Catalan margin is not well explained and may be related to local reactivation of deep faults at the northern and southern boundaries of the Pyrenean Axial Ranges. In particular, the North Pyrenean Fault may have been reactivated with a left-lateral strike-slip mechanism. From our study and on-land studies, we conclude that the extensional tectonics are late Miocene–early Pliocene in age and that the present stress is probably compressional, although the offshore evidence of compression is not yet shown on the regional deep penetration seismic profiles.

4.6. Transfer zones

The Gulf of Lion extensional basin is separated from the mainland towards the east (Provence) and the west (Pyrenees and Catalan Region) by major transfer zones. The western limit is named the Catalan transfer zone (Lefebvre, 1981; Gorini et al., 1993). This feature is parallel to the 120–140° trending faults that segment the Gulf of Lion (Séranne et al., 1995) and the Valencia Trough (Maillard and Mauffret, 1999). The Catalan transfer zone is located along the Cap de Creus canyon (Berné et al., 1999; Fig. 15A) and shows a strong gravimetric gradient on the free air gravity map (Pascal et al., 1993; Mauffret et al., 1995, Fig. 15B) that is related to the abrupt thickening of the crust beneath the Catalan margin. A seismic profile (Fig. 15C and D) shows a step in the sea floor in the Messinian unconformity and in the acoustic basement and this disposition suggests that the Catalan transfer zone has a major control on the tectonics and the sedimentation in this area. The refraction results (Nercessian et al., 2001; Vidal et al., 2001) show, below the Catalan transfer zone, a thickening of the crust and a correlative dipping of the Moho towards the Pyrenees. The Catalan transfer zone guided the extensional motion of the Rascasse high and the coeval

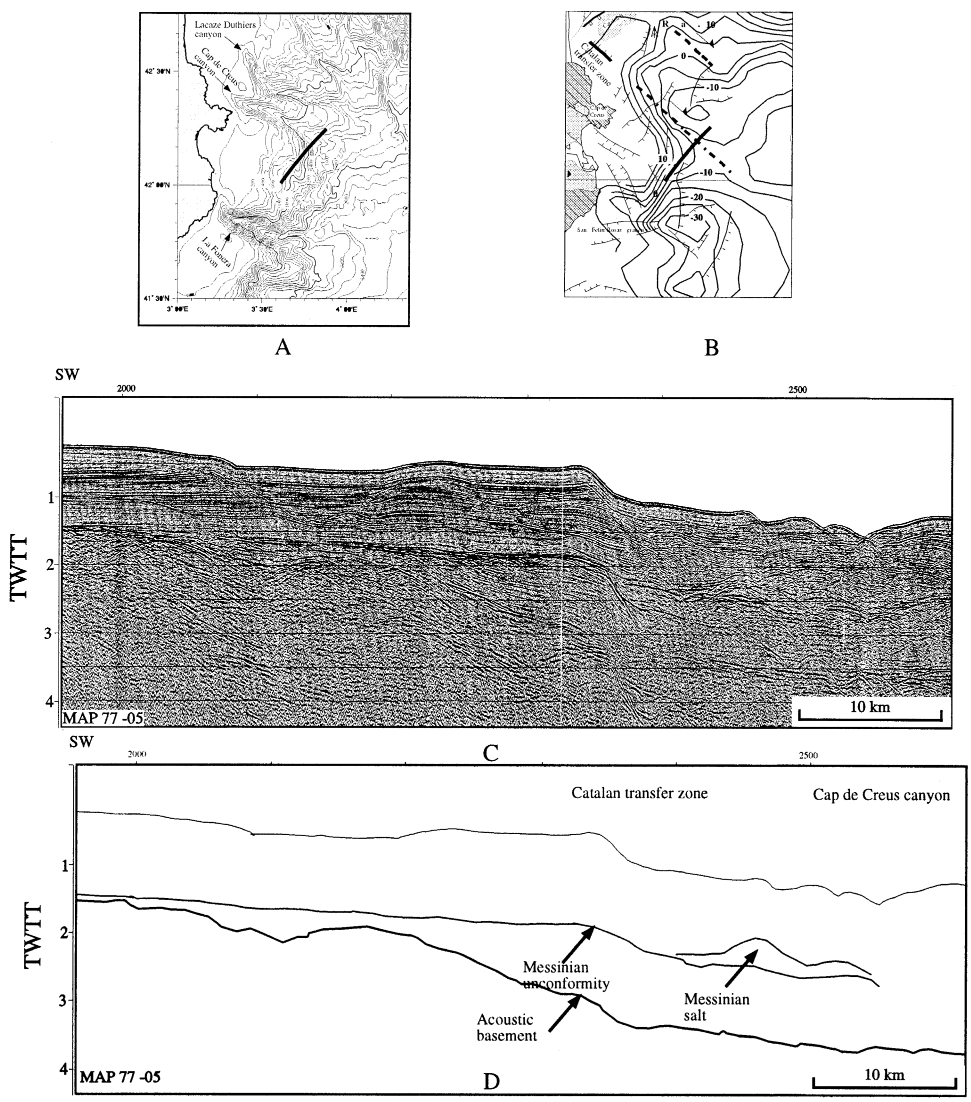


Fig. 15. (A) Bathymetric map of the Catalan region (Berné et al., 1999). Note the trend of the Cap de Creus canyon. (B) Free air gravity map (Pascal et al., 1993; Mauffret et al., 1995). Note the NW–SE trend of the gravity gradient beneath the Catalan transfer zone. (C) and (D) Seismic profile MAP 77-05 and its line drawing. Location Fig. 1 and in (A) and (B). Note the slopes of the seafloor, the Messinian unconformity and the acoustic basement correlated with the Catalan transfer zone.

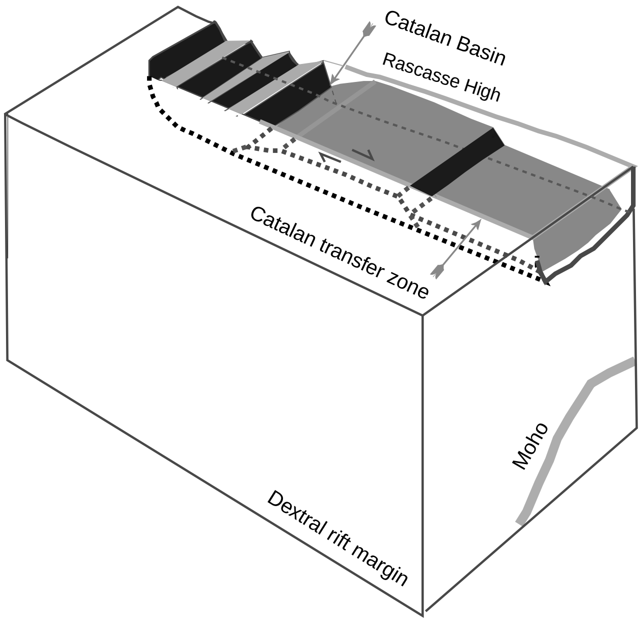


Fig. 16. Sketch of the Catalan transfer zone modified from Faulds and Varga (1998). Observe the listric geometry of the transfer zone and the deepening of the Moho towards the continent.

opening of the Catalan Basin (Fig. 16). The boundary between the Gulf of Lion and the Pyrenees is a typical dextral rift margin (Faulds and Varga, 1998). On the continental shelf, the Albères Massif is limited by a steep slope but the contact between the acoustic basement and the sedimentary infilling shows a listric geometry and the Miocene strata are tilted in relation to the uplift of the Albères (Fig. 5). The main orientation of the extensional structures in the Gulf of Lion is NE–SW (Gorini et al., 1993), whereas the trend of the Rascasse horst is almost N–S (Fig. 2). This trend can be explained by an addition of the NW–SE extensional vector predominant in the Gulf of Lion and the NE–SW vector related to the normal motion along the Catalan transfer zone (see the inset, Fig. 17B). In addition, we may suppose a counter-clockwise rotation of the Rascasse high in relation to the dextral motion of the Catalan transfer zone (inset, Fig. 17B). In this region of the Gulf of Lion, the Catalan and the Cathares Basins are separated by another transfer zone with a step in the Moho depth from 21 to 24 km (Fig. 1). The major boundary between the Valencia Trough and the Gulf of Lion is the North Balearic fracture zone that is the trace of Sardinia during its Miocene rotation (Maillard and Mauffret, 1999). The volcanism in the Olot region and the Transverse Range (Fig. 1) is the probable on-land extension of this fracture zone (Maillard, 1993). We propose that the region between Olot and the Roussillon is a major transfer zone between the Gulf of Lion and the Valencia Trough. The faults of the Transverse Range and El Empordà have a normal component (Saula et al., 1994; Lewis et al., 2000) but a right-lateral component, like the Catalan transfer zone, is also probable.

5. Structural framework and summary of the tectonic evolution

We suggest that the North Pyrenean Fault is the northern boundary of the Roussillon Basin. This fault probably extends in a westerly direction in the Gulf of Lion (Arthaud and Mattauer, 1972; Olivet, 1996). The aeromagnetic map (Fig. 17A) of the Mediterranean Sea (Galdeano and Rossignol, 1977) shows this E–W orientation north of a very large magnetic anomaly that strikes NW–SE. It is clear (Muon, 1984; Guennoc et al., 1994) that the E–W trend is related to the North Pyrenean Fault, whereas the NW–SE orientation is linked to the transfer faults and the extensional motion of the crustal blocks towards the ocean. The relationships between the rocks and the prominent magnetic anomaly are not known because the Rascasse well sampled Palaeozoic rocks (Gorini, 1994) and the magnetic rocks linked with the magnetic anomalies are deeper than this Palaeozoic basement (Muon, 1984; Guennoc et al., 1994). It cannot be excluded, however, that a piece of upper mantle was emplaced during the Mesozoic extension then squeezed and intruded into the crust during the Pyrenean Orogeny (Casas et al., 1997). Then this piece of deep material could have been extended by rifting towards the southeast during the Oligocene–early Miocene epoch. The North Pyrenean Fault is active to the present day with a left-lateral motion as shown (Fig. 14) by the focal mechanisms in the Agly Massif (Pauchet et al., 1996) and along the fault (Olivera et al., 1986). The zigzag pattern of this fault (Fig. 17B) and the same feature along the northern boundary, the so-called Tech Fault, of the Albères Massif (Calvet, 1985, 1986) may be related to a left-lateral motion in the Eastern Pyrenees. However, we show that a fault along the northern boundary of the Albères Massif is not evident and that this feature is probably the hanging wall relative to the footwall located north of the Roussillon Basin. The minor faults that correspond to the Tech Fault are in an antithetic position relative to the main detachment. The Albères Massif has been displaced towards the southeast during the formation of the basin. Nevertheless, the uplift of the Albères (Calvet, 1985) the Canigou Massif and a part of the Catalan Range (Lewis et al., 2000) are clearly caused by a late Miocene–early Pliocene flexural isostatic rebound caused by the thinning of the crust. The Albères Massif is the hanging wall relative to the Roussillon Basin but it may be the footwall of the El Empordà Basin. The Canigou uplift, which has a normal component, may be related to the thinning of the crust beneath the Roussillon Basin and the transfer fault, which trends NW–SE (Fig. 17B) and is located at the boundary between the two. We do not know if the Albères Massif is cut by the fault but a similar fault system exists in the El Empordà (Tassone et al., 1996). The geometry of the Conflent, Roussillon and Catalan Basins shows the role of the transfer faults that also transform the extensional motion from the Gulf of Lion to the Valencia Trough. The sketch in

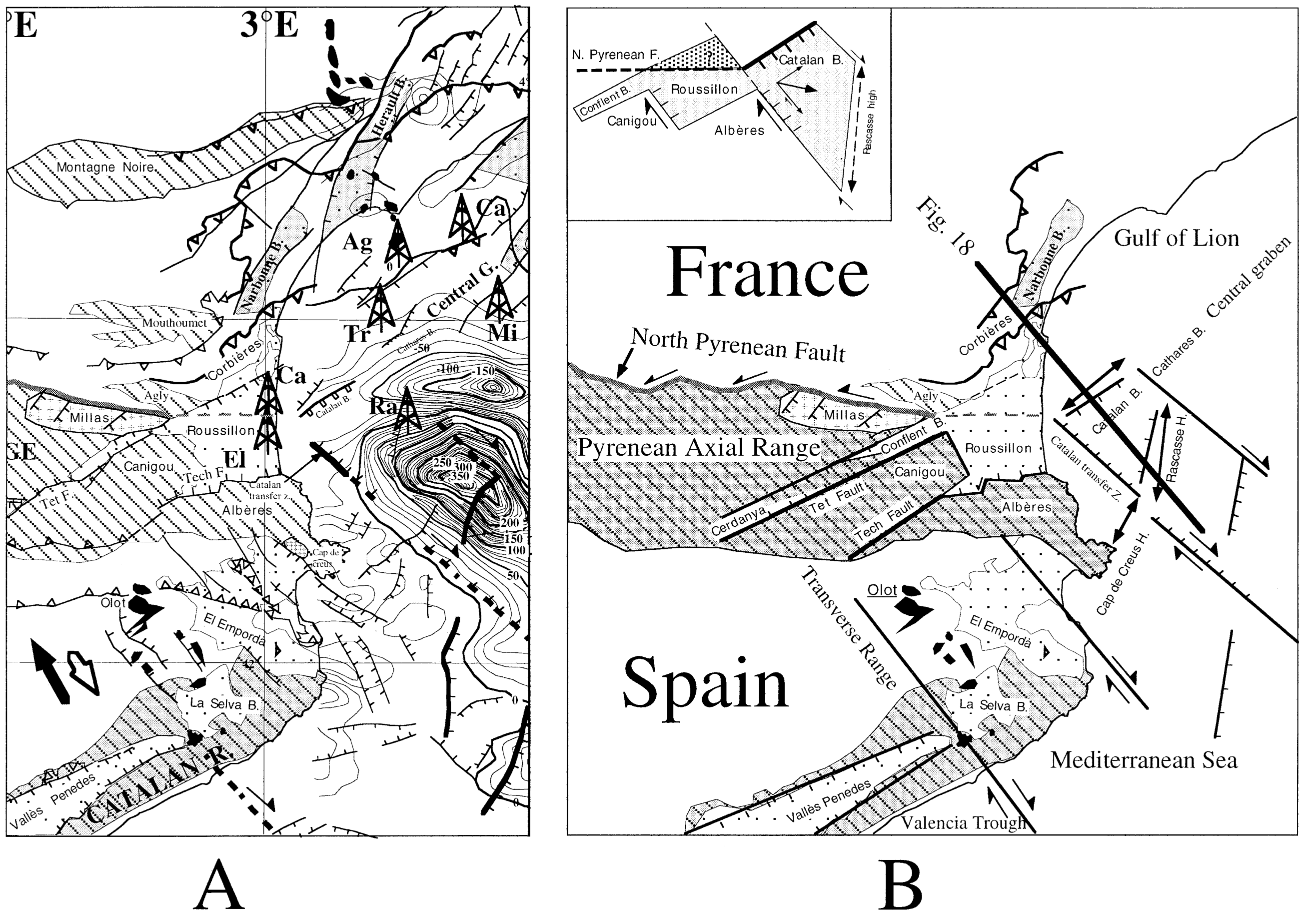


Fig. 17. (A) Aeromagnetic map of the Mediterranean Sea (Galdeano and Rossignol, 1977). Observe the E–W trend that limits to the north the prominent positive (350 nT) magnetic anomaly. This E–W trend has been correlated with the North Pyrenean Fault (Muon, 1984). The NW–SE trend beneath the Rascasse structure may correspond to an extensional block of dense rocks (peridotites from the mantle) limited by the transfer faults. (B) Simplified structural map of the study area. The location of Fig. 18 is indicated. Observe the possible recent activity of left-lateral strike-slip faults along the North Pyrenean Fault and the Tech Fault north of the Albères Massif (Calvet, 1985; 1986). In the upper left inset we show the role of the transfer zones in the opening of the basins. Note the almost N–S trend of the Rascasse high that is explained by an addition of the extensional vectors and a counter-clockwise rotation. We assume that the Roussillon Basin is closed to the north by the North Pyrenean Fault. Nevertheless this basin may be closed by the Tet Fault as proposed in several publications (Clauzon et al., 1987a,b). The large difference in altitude between the high massifs (Canigou, Albères) and the Roussillon and Catalan Basins is explained by normal motions and uplift by isostatic rebound along the NE–SW faults and the perpendicular transfer zones. The Transverse Range in the Catalan region may be related to a major transfer fault and not a simple footwall in a NE–SW extensional regime as proposed by Saula et al. (1994) and Lewis et al. (2000).

Fig. 18 (modified from Viallard and Gorini (1994)) shows the formation of the half grabens and horsts in the Gulf of Lion above deep detachments that penetrate deeper and deeper in the crust. We detect two extensional periods with rotation of crustal blocks along the same deep listric detachment. The younger one, late Miocene–early Pliocene in age, could be related to the unloading and correlative uplift shoulders caused by the Messinian desiccation crisis (Norman and Chase, 1983) then the loading caused by the deposition of thick Messinian evaporites and salt layers and the subsequent Pliocene reflooding of the basin. However, the volcanism in El Empordà, the tectonic activity in the Pyrenees and adjacent regions suggest a deep influence at the crustal level and the lithosphere may have been thinned by a deep process not yet fully understood. Tomographic and geochemical studies (Hoernle et al., 1995) suggest a large-scale mantle upwelling beneath western Europe with low velocity regions at a depth of 150–250 km. This low velocity can be related to a hot mantle upwelling that induces a thermal anomaly, a volcanic activity, a thinned lithosphere and extension. Late Miocene to Quaternary volcanism in the Catalan region (Bartrina et al., 1992), in the Valencia Trough (Marti et al., 1992; Maillard, 1993; Maillard and Mauffret, 1993) and in the Massif Central (Granet et al., 1995) can be an expression of the hot mantle upwelling and related plumes. In the Western Mediterranean Sea, the hot mantle can be related to a detached slab of the North African subduction zone (Hoernle et al., 1995; M. Granet, written communication). In the Catalan Range, it is proposed that the mountainous topography is related to a deep lithospheric thinning (Lewis et al., 2000). However, our study shows that, in the Gulf of

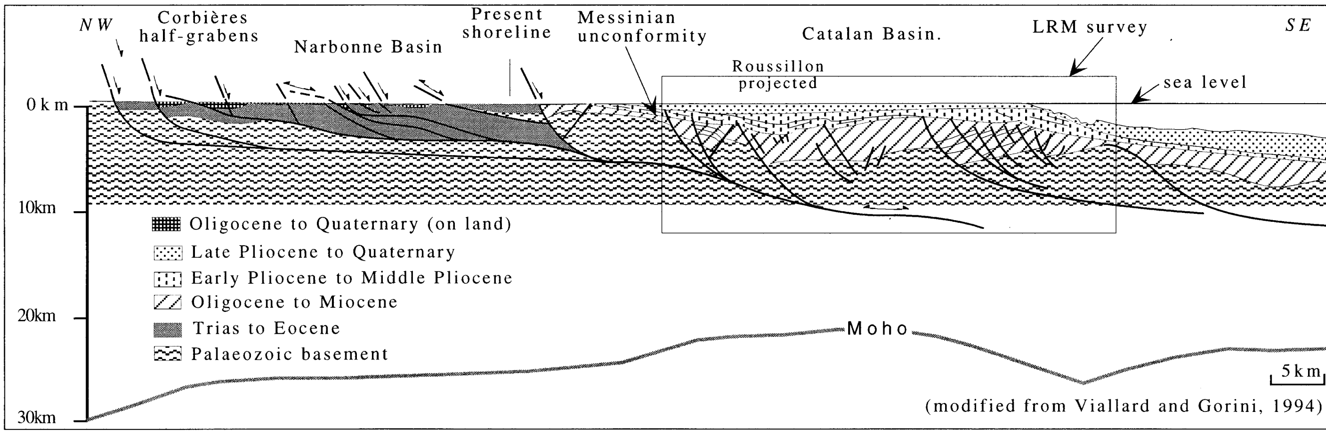


Fig. 18. The geological transect at the crustal scale (modified from Viallard and Gorini (1994)) shows the rise of the Moho beneath the Catalan Basin and the detachments faults at different levels. On land, the main detachment is located at the boundary between the Paleozoic basement and the Triassic evaporites. However, some detachments may penetrate deeper into the basement (Viallard and Gorini, 1994) and merge with the deep detachments beneath the continental shelf. The detachments are former Eocene thrusts well known in the Corbières Massif, reactivated in extension during the Oligocene–early Miocene then late Miocene–early Pliocene epochs.

Lion, the extension, tilting of the sedimentary layers and uplift of the rift shoulders are late Miocene–early Pliocene, whereas the volcanism was still active during the Quaternary in the Massif Central (Granet et al., 1995) and in the Valencia Trough (Marti et al., 1992). The low velocity in the asthenosphere is a present day feature (Hoernle et al., 1995) and the uplift around the Valencia Trough is mainly Pliocene (Janssen et al., 1993). Therefore, the relatively rapid tectonic event detected during the late Miocene and the early Pliocene in the Gulf of Lion cannot be directly related to the hot mantle upwelling that is active up to the present. On the other hand, the uplift caused by the Messinian desiccation (Norman and Chase, 1983), that fits in time with our observations, cannot be directly correlated with the mantle upwelling that begins before the Messinian and is active to the present day (late Miocene to Quaternary volcanism).

6. Conclusions

The study of the complete seismic data located on the western continental shelf of the Gulf of Lion allows us to determine the tectonic evolution and the processes of the crustal thinning of the Eastern Pyrenees. We do not think that the crust beneath the Gulf of Lion has been thickened during the Pyrenean Orogeny. However, in contrast to previous studies, we find the synrift sequence to be very thin and an extension above sea-level may be an explanation. Extensional tectonics in the region are evident with deep detachments in the basement and motions of hanging walls followed by isostatic uplift of the footwalls. In addition, the transfer faults that are normal to extensional faults play an important role in the opening of the half grabens. The extension is mainly related to the formation of the Mediterranean Sea during the early Miocene–Oligocene times but a second pulse of extension is clearly recorded by the sedimentary cover during the late Miocene–early Pliocene. The high massifs of the Eastern Pyrenees (Canigou, Albères) result from this late extension. The extension is no younger than early Pliocene and mainly active during the late Miocene and the Pyrenean region may be undergoing a compression at the present day. We do not explain either the late Miocene–early Pliocene extension or the stress deduced from the focal mechanisms of the present seismicity. The main conclusion of this study is the importance of the crustal thinning in the transfer zones and that this thinning can be confused with the normal extension along the horst and graben structures. The Transverse Range in Catalonia and the Canigou Massif in France show a high position relative to the basins that are related to NW–SE transfer faults. On the other hand, culmination of the Albères Massif, as well as other structures, is the result of isostatic rebound due to the unloading of the footwalls caused by the collapse of the hanging walls towards the Mediterranean Sea. The uplift of the Albères Massif is evaluated to be 1.7 km in its offshore portion. A mantle upwelling may be an explanation of the uplift and late Miocene–early Pliocene extension. However, our study demonstrates that this extensional tectonic event in the Gulf of Lion is restricted in time and cannot be extended to the present. A correlation between the brief extensional tectonic episode and the Messinian salinity crisis was also proposed but we do not know the relation between this Messinian event and the mantle upwelling. The Messinian salinity crisis is probably caused by a tectonic event that closed the connection between the Mediterranean Sea and the Atlantic Ocean. The Messinian tectonic event may be a pulse of activity related to the mantle thermal anomaly but this hypothesis has to be tested.

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