AGU PUBLICATIONS

Tectonics

RESEARCH ARTICLE
10.1002/2014TC003570

Key Points:
- Exhumed lower crust and mantle make the distal Gulf of Lion margin
- Extraction of lower crust achieved by shallow-dipping detachments
- Slab retreat is the primary engine of extension and asthenospheric flow

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Citation:
Jolivet, L., C. Gorini, J. Smit, and S. Leroy (2015), Continental breakup and the dynamics of rifting in back-arc basins: The Gulf of Lion margin, *Tectonics*, *34*, 662–679, doi:10.1002/2014TC003570.

Received 1 MAR 2014
Accepted 2 MAR 2015
Accepted article online 4 MAR 2015
Published online 3 APR 2015

Continental breakup and the dynamics of rifting in back-arc basins: The Gulf of Lion margin

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Abstract Deep seismic profiles and subsidence history of the Gulf of Lion margin reveal an intense stretching of the distal margin and strong postrift subsidence, despite weak extension of the onshore and shallow offshore portions of the margin. We revisit this evolution from the geological interpretation of an unpublished multichannel seismic profile and other published geophysical data. We show that an 80 km wide domain of thin lower continental crust, the "Gulf of Lion metamorphic core complex," is present in the ocean-continent transition zone and exhumed mantle makes the transition with oceanic crust. The exhumed lower continental crust is bounded upward and downward by shallow north dipping detachments. The presence of exhumed lower crust in the deep margin explains the discrepancy between the amount of extension deduced from normal faults in the upper crust and total extension. We discuss the mechanism responsible for exhumation and present two scenarios: the first one involving a simple coupling between mantle extension due to slab retreat and crustal extension and the second one involving extraction of the lower crust and mantle from below the margin by the southeastward flow of hot asthenosphere in the back-arc region during slab rollback. In both scenarios, the combination of Eocene crustal thickening related to the Pyrenees, the nearby volcanic arc, and a shallow lithosphere-asthenosphere boundary weakened the upper mantle and lower crust enough to make them flow southeastward. The overall hot geodynamic environment also explains the subaerial conditions during most of the rifting stage and the delayed subsidence after breakup.

1. Introduction

The enigmatic nature of the ocean-continent transition (OCT) at magma-poor passive margins is a highly debated topic [*Emiliani*, 1965; *Sibuet et al.*, 2006]. Recent advances show a transitional domain between clearly continental and clearly oceanic characteristics [*Péron-Pinvidic and Manatschal*, 2009; *Leroy et al.*, 2010], and depending upon the chosen interpretation, the mechanisms explaining the final rupture of the continental lithosphere may strongly differ [*Lavier and Manatschal*, 2006; *Ranero and Pérez-Gussinyé*, 2010; *Reston*, 2010; *Huismans and Beaumont*, 2011; *Reston and McDermott*, 2011]. Several examples of Atlantic margins show a narrow zone of crustal thickness gradient (~50 km), followed seaward by a wide zone of strongly attenuated continental crust (~180 km wide) underlying a thick sag basin with an apparent lack of deformation that started to form in shallow water depth conditions [*Contrucci et al.*, 2004; *Moulin et al.*, 2005]. The presence of a localized weak lower crust may favor the early necking of the lithospheric mantle and spreading of the crust, leading to a final geometry close to that of Atlantic margins [*Huismans and Beaumont*, 2011]. Other examples show wide zones of exhumed mantle instead, and published scenarios involve the sequential development of several detachments, strongly controlled by inherited lithospheric structures [*Whitmarsh et al.*, 2001; *Manatschal*, 2004]. Rifting mechanisms may also vary along strike, and passive margins often show a segmentation such as in the Gulf of Aden [*Gueguen et al.*, 1998; *Leroy et al.*, 2010]. The amount of extension often varies with depth, and upper crustal extension can be much smaller than the extension required to explain tectonic subsidence, suggesting that the lower crust has accommodated much more extension [*Driscoll and Karner*, 1998]. This situation implies lateral migration of material within the crust and thus requires shear zones to accommodate this transfer. The transitional domain between the continental and oceanic crusts can then be composed of exhumed mantle, upper continental crust, or lower crust (see a discussion in *Sibuet and Tucholke* [2012]).

The Gulf of Lion margin shows many of those features typical of Atlantic-type passive margins [*Aslanian et al.*, 2009; *Bache et al.*, 2010], but it evolved in an entirely different tectonic setting associated with back-arc

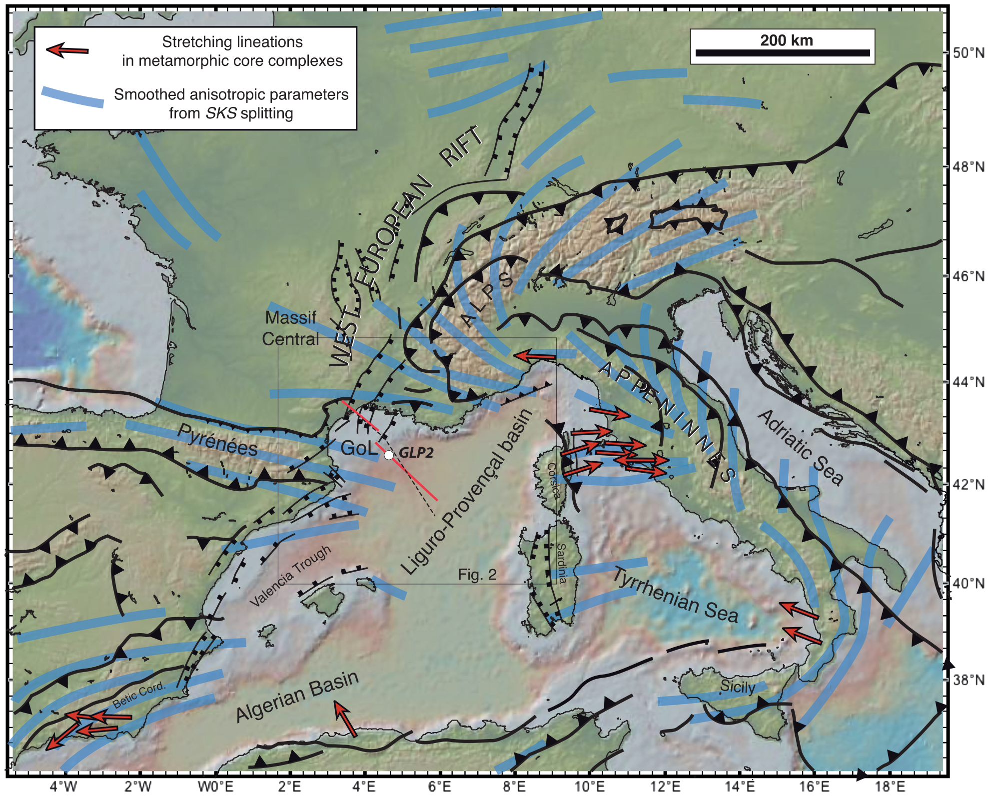

Figure 1. Topographic and bathymetric map of the Mediterranean region showing the position of the main structures, thrust front, subduction zones, the main strike slip, and normal faults. The red arrows show the directions of ductile stretching and shear senses within metamorphic core complexes. The blue thick lines show the smoothed fast direction of *SKS* waves in the back-arc domain and the blue bars in the vicinity of the subducting slab [*Jolivet et al.*, 2009]. The thin red line marks the location of the TGS-NOPEC MCS profile and the portion of the cross section of *Gorini et al.* [1994] used in this paper; the dashed black line marks the location of the ECORS profile used by *Gorini et al.* [1994]. The white dot along the profile marks the position of GLP2 drill hole. GoL: Gulf of Lion. This map shows that in back-arc domains, crustal stretching and asthenospheric flow due to slab retreat are partly coupled and that shear stresses can possibly be transmitted from the flowing asthenosphere to the lower crust.

extension [*Gorini*, 1993; *Séranne*, 1999]. The Gulf of Lion belongs to the Liguro-Provençal Basin (Figure 1) that formed during the Oligocene and Miocene in the back-arc region of the retreating Apennines subduction [*Réhault et al.*, 1984; *Gueguen et al.*, 1998; *Barruol and Granet*, 2002; *Lucente et al.*, 2006]. We propose here an interpretation of a deep-penetration multichannel seismic (MCS) profile that shows the ocean-continent transition with unprecedented details, suggesting that most of the crustal thinning has been accommodated by a viscous outward flow of the lower crust and mantle below shallow-dipping detachments, leading to the formation of 80 km wide metamorphic core complex: the Gulf of Lion metamorphic core complex (GoL MCC). The upper and lower contacts of the GoL MCC are two NW dipping detachments that form reflectors visible on the MCS profile. We discuss several scenarios of extension and coupling between crustal and mantle deformation during back-arc extension.

2. Geological and Geodynamic Setting

The Gulf of Lion margin belongs to the Liguro-Provençal Basin and its southern extension in the eastern Algerian Basin (Figures 1 and 2), the largest Mediterranean back-arc basin formed in the Oligocene and Miocene by the rotation of the Corsica-Sardinia block [*Auzende et al.*, 1973; *Dewey et al.*, 1973, 1989] as a consequence of the retreat of the Apennine subduction [*Réhault et al.*, 1984; *Gueguen et al.*, 1998]. After an Oligocene rifting episode (32–24 Ma), oceanic crust was formed in the early Miocene and the base of the middle Miocene until ~15 Ma [*Westphal et al.*, 1976; *Vigliotti and Kent*, 1990; *Gorini et al.*, 1993; *Mauffret et al.*, 1995; *Seranne et al.*, 1995; *Zarki-Jakni et al.*, 2004; *Chamot-Rooke et al.*, 1999; *Séranne*, 1999; *Speranza et al.*,

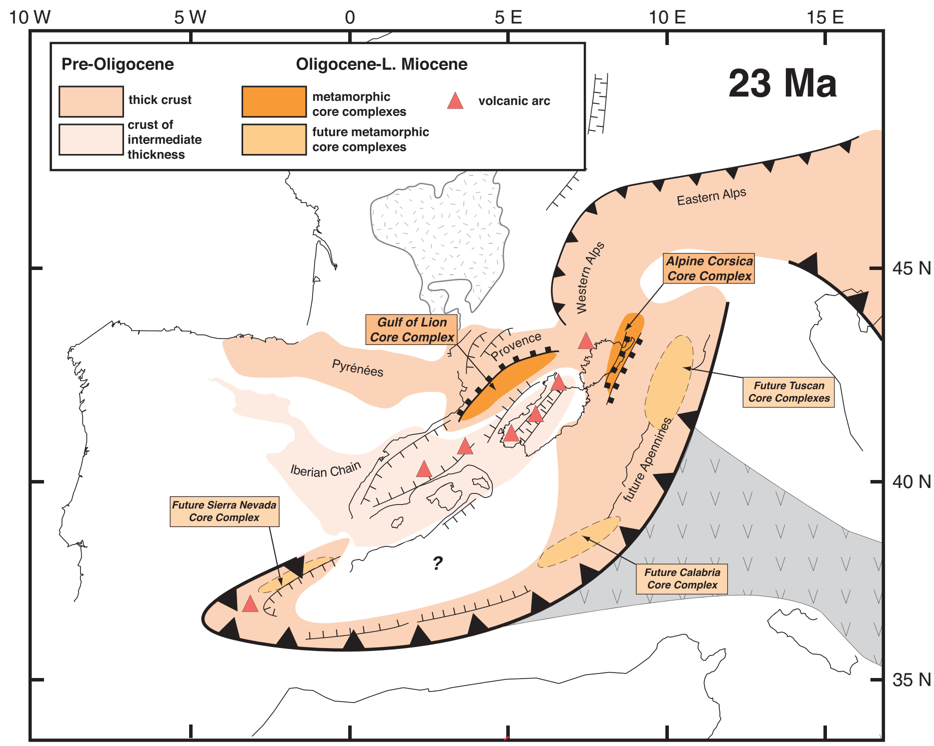

Figure 2. Depth to prerift map, color map (depth shown in meter) compiled by *Mauffret et al.* [1995], grey scale map (depth in seconds, two-way travel time) from *Rollet et al.* [2002]. GLP2: Drill hole Golfe du Lion Profond 2. The red dotted line represents the seismic refraction profile of *Gailler et al.* [2009], the thin red line marks the location of the TGS-NOPEC MCS profile and the portion of the cross section of *Gorini et al.* [1994] used in this paper, and the dashed black line marks the location of the ECORS profile used by *Gorini et al.* [1994].

2002; *Bellahsen et al.*, 2012]. The inception of rifting in this region occurred contemporaneously with other Mediterranean back-arc basins [*Jolivet and Faccenna*, 2000], but in those other cases, back-arc extension only led to the collapse and thinning of a previously thickened crust (Aegean, Alboran, and Northern Tyrrhenian) or the formation of small oceanic basins with a very short oceanic ridge (southern Tyrrhenian) [*Le Pichon and Angelier*, 1979; *Dewey*, 1988; *Kastens and Mascle*, 1990; *Nicolosi et al.*, 2006].

Figure 3 shows a possible reconstruction in map view at the Oligocene-Miocene transition. The volcanic arc that was active in Sardinia in the Oligocene during the formation of the West Sardinia Graben results from the northward subduction of the Ionian lithosphere below the Corsica-Sardinia block and the future Apennines that form a thickening orogenic wedge. This arc runs obliquely to the Pyrenean-Provence mountain belt that results from the Eocene collision between Iberia and Europe. In the present situation, the Pyrenees end abruptly in the Mediterranean Sea where they are replaced by the Gulf of Lion margin. Remnants of the Eocene belt are found to the north of the Gulf of Lion in the Provençal foreland that was mostly preserved from extension.

The study of *SKS* shear wave seismic anisotropy below the central and western Mediterranean [*Barruol and Granet*, 2002; *Lucente et al.*, 2006] and the comparison of the obtained fast directions with the stretching directions observed in the metamorphic core complexes exhumed during the Oligocene and Miocene [*Jolivet et al.*, 2009] suggest that the crust and the asthenospheric mantle were partly coupled during extension and slab retreat and that the kinematics of extension was controlled by asthenospheric flow toward the retreating slab (Figure 1). The mantle fabric cannot be directly dated of course, and it is the parallelism with the well-dated Oligocene-Miocene crustal fabric in all Mediterranean back-arc domains and the compatibility with the flow direction expected from the reconstructions involving slab retreat that led *Barruol et al.* [2004] and *Jolivet et al.* [2009] to attribute the mantle fabric to a Neogene phase of asthenospheric flow.

The Liguro-Provençal Basin is bordered to the east by the Corsica-Sardinia block and to the north by the Provence margin. Both margins are segmented [*Gueguen et al.*, 1998] with narrow portions (offshore Nice and Corsica) and wider portions (Gulf of Lion and Sardinia) portions. The Gulf of Lion margin is settled on the

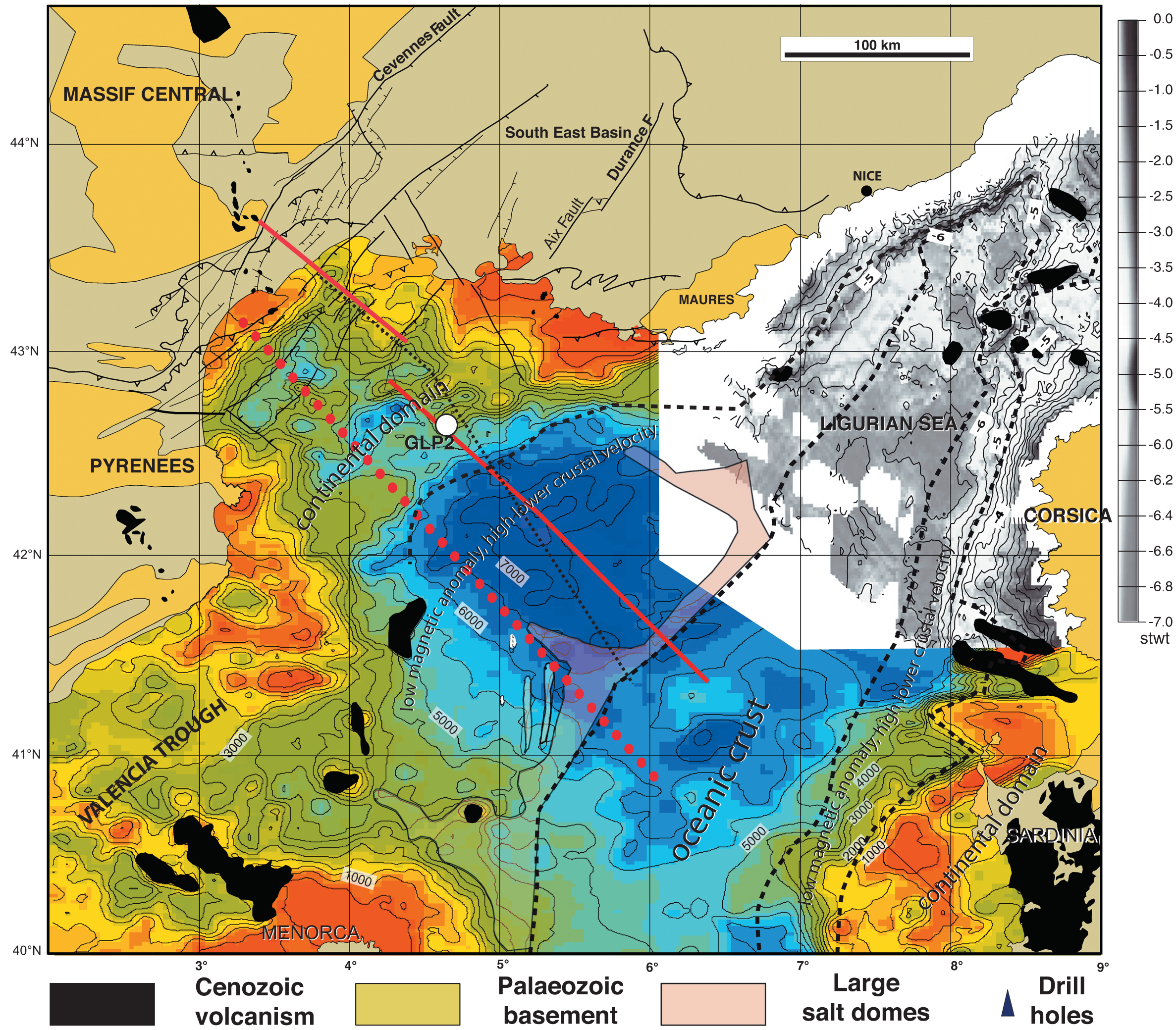

Figure 3. Possible map view reconstruction of the western Mediterranean at the Oligocene-Miocene transition adapted from *Lacombe and Jolivet* [2005]. The GoL metamorphic core complex belongs to a set of metamorphic core complexes cropping out around the western Mediterranean Sea. Most of them correspond to the thinning of a thick crust during the Oligocene-Miocene postorogenic stage. The GoL metamorphic core complex instead results from the thinning of a crust that had not been strongly thickened before, and the extraction of the lower crust by the basal drag of asthenospheric flow led to extreme thinning and thus strong postrift subsidence once the volcanic arc had retreated southeastward following the Apennine slab.

offshore extension of the Pyrenees and the Provence fold and thrust belt that predated the rifting episode. The continental crust, now extended, had thus been first thickened during the Late Cretaceous and the Eocene. Corsica has also recorded a crustal thickening event, as it was the southern extension of the Alps in the Eocene. The Gulf of Lion was thus in the vicinity of the triple junction between the Pyrenees and the Alps when extension started [*Gorini et al.*, 1994; *Vially and Tremolières*, 1996; *Chamot-Rooke et al.*, 1999; *Séranne*, 1999; *Lacombe and Jolivet*, 2005]. It is also on the southern extension of the West European Rift System. At the time of rifting, a calc-alkaline volcanic arc was active in Sardinia, southern Provence, and the Valencia Trough. It started some 30 Ma ago in Sardinia and Provence and slightly later in the Valencia trough.

Magnetic anomalies and seismic velocities show the presence of oceanic crust in the central part of the basin [*Le Douaran et al.*, 1984; *De Voogd et al.*, 1991; *Pascal et al.*, 1993]. The ocean-continent transition is characterized by an enigmatic crust with low-amplitude magnetic anomalies and abnormal velocities that suggest exhumed mantle or lower crustal material [*Gailler et al.*, 2009].

The margin itself has accumulated a thick sedimentary cover, mostly during the postrift stage [*Burrus*, 1984]. The synrift basins are indeed rather small, and early studies of the Gulf of Lion passive margin have emphasized a paradox between an apparent low stretching of the crust, as indicated by studies of faulting in

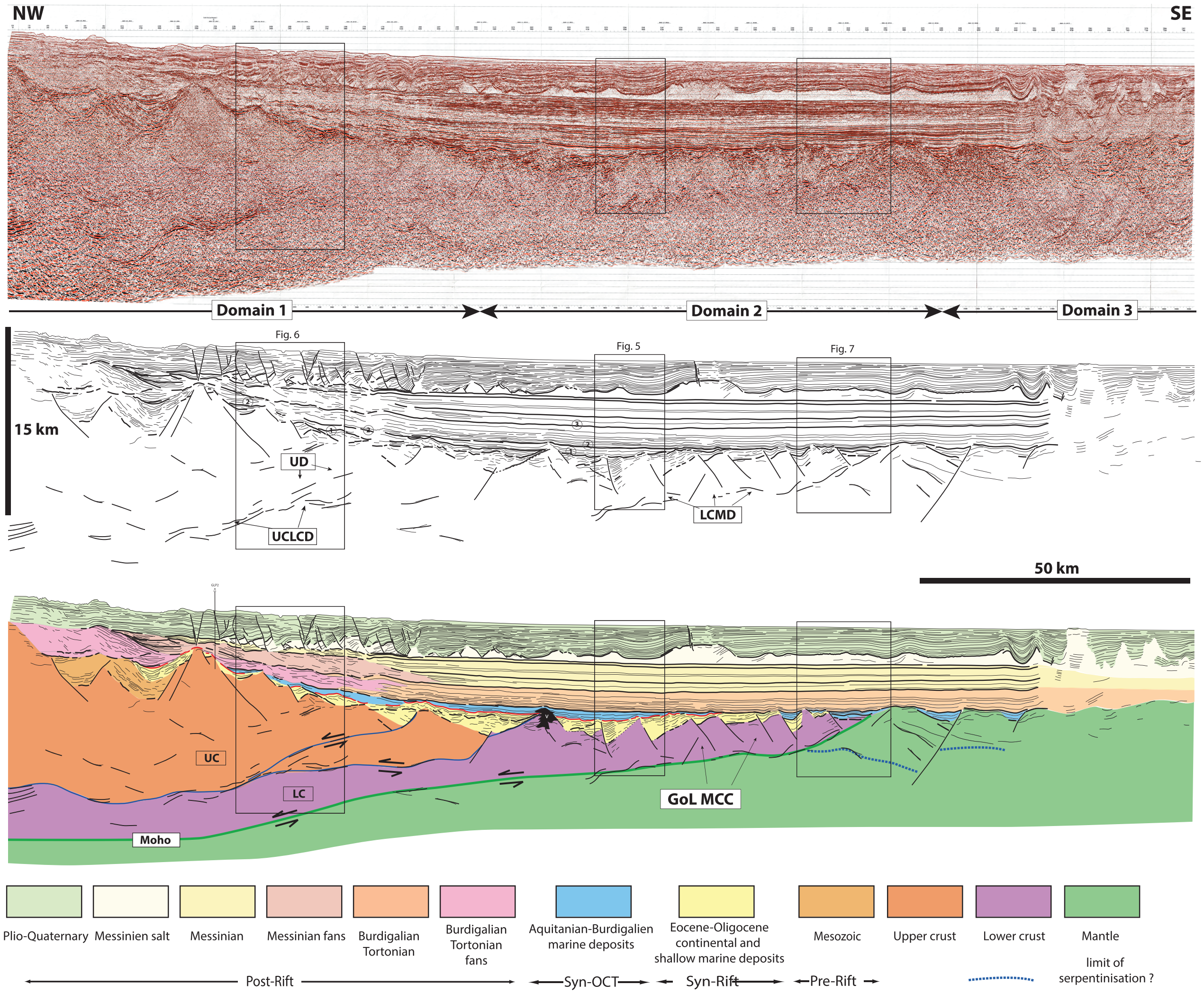

Figure 4. An interpretation of the TGS-NOPEC profile. (top) The seismic profile without interpretation. (middle) Line drawing. (bottom) Interpretation. UD: Upper Detachment, UCLCD: upper crust lower crust detachment, LCMD: lower crust-mantle detachment. Reflectors 1, 2, and 3: see text. A wide domain of lower crust has been exhumed within the continent-ocean transition zone below a series of shallow-dipping detachments forming the Gulf of Lion metamorphic core complex (GoL MCC). Once exhumed, it has then been cut by a series of steep normal faults dipping toward the continent. This geometry explains the contrasts, often noted, of a poorly extended upper crust and a strong finite stretching factor at crustal scale. The shape of the lower crustal body shows that it has been extracted from below the margin toward the ocean-continent transition. The presence of an (bottom) erosion surface at the base of the postrift sequence points to subaerial rather than submarine erosion during the rifting stage and shows that the entire subsidence occurred during the postrift episode as already described in detail by *Bache et al.* [2010].

the upper crust, and a fast thermal subsidence, suggesting instead a strong crustal thinning [*Burrus*, 1984; *Gorini*, 1993; *Séranne*, 1999]. All subsequent works have shown the existence of an 80–110 km wide zone of anomalous thin crust forming the transition between the little extended Provençal continental crust and the oceanic crust of the Liguro-Provençal Basin [*Bache et al.*, 2010]. One cornerstone study was the acquisition of the MCS Etude Continentale et Océanique par Reflexion et Refraction Sismique (ECORS) profile [*De Voogd et al.*, 1991; *Pascal et al.*, 1993; *Gorini et al.*, 1994] and associated refraction data that showed a thin crust with seismic velocities intermediate between continental and oceanic ones [*Pascal et al.*, 1993]. Besides several tilted blocks, oceanward dipping normal faults, and thin synrift sediments, the ECORS profile and its later reprocessing by prestack depth migration show a distinct reflector shallowly dipping toward the continent, interpreted as an antithetic detachment plane [*Mauffret et al.*, 1995; *Seranne et al.*, 1995]. Recent seismic refraction experiments (ocean bottom seismometers (OBS)) suggest that this anomalous crust may be made of lower continental crust [*Gailler et al.*, 2009], but its detailed geometry and exhumation mechanisms are still unknown because the lower part of the margin was so far poorly imaged by seismic profiles.

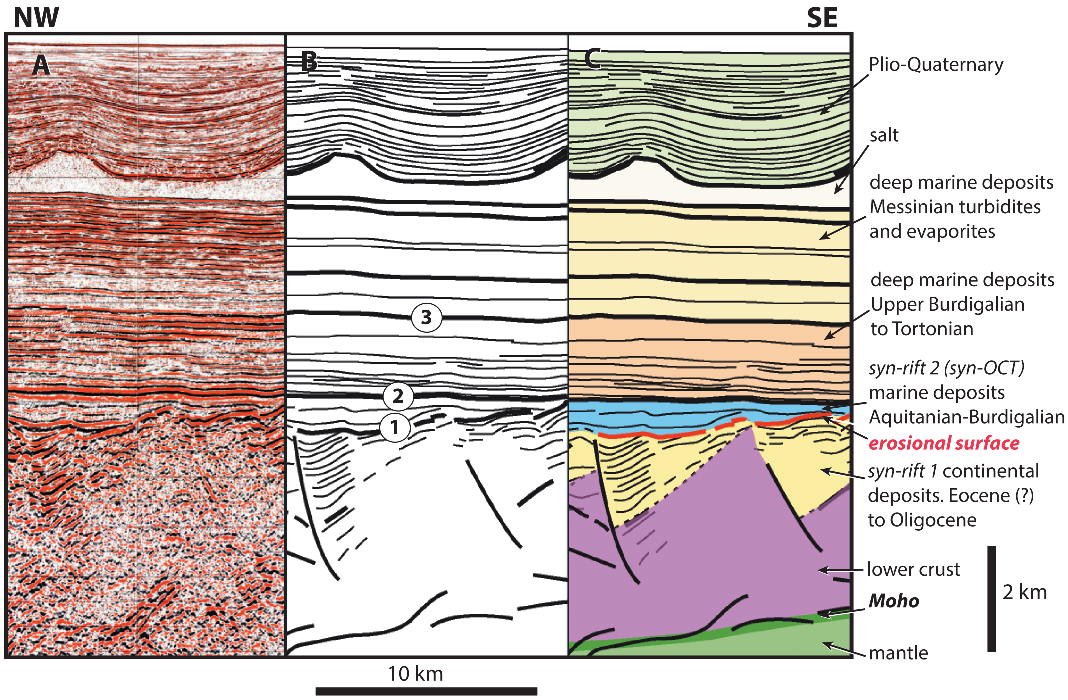

Figure 5. Details of a section of the distal margin located most up-slope and closest to well GLP2. Here the sediments rests directly on upper crustal basement (see details of more distal sections in Figures 6 and 7 and location in Figure 4). (left) The seismic profile without interpretation. (middle) Line drawing (1, 2, and 3 are the reflectors discussed in the text). (right) Interpretation.

3. Interpretation of the TGS-NOPEC Seismic Profile

Located close to the ECORS profile, the TGS-NOPEC seismic profile gives a new picture of the distal margin. This profile was acquired on the M/V Zephir 1 by DMNG for TGS-NOPEC with a Bold Airgun source and a 480-channel streamer and is here presented as a depth-converted version. First, a Stolt F-K time migration: velocities clipped (4000 m/s) and smoothed (480 common depth point reflections (CDPs)), velocity analysis interactively picked every 1.5 km, a multivelocity stack ±10 % picked function, then a Kirchoff time migration, velocities clipped (6000 m/s), and smoothed (480 CDPs). For more details about depth conversion, please contact TGS as the original acquirers and processors of the processed migration of the seismic line. Existing refraction seismic data acquired in 1990s (expanding spread profiles (ESP)) [*Mauffret et al.*, 1995; *Contrucci et al.*, 2001] are used to obtain propagation velocities in different layers and to identify the nature of crustal units. We have interpreted the whole TGS-NOPEC profile from the recent deposits down to the Moho, but this paper is mainly concerned with the deep portions of the distal margin (Figures 4–7). Once interpreted, the profile has been integrated in a complete section of the margin from the onshore Provençal section to the deep basin (Figure 8), using earlier works [*Gorini et al.*, 1994; *Mauffret et al.*, 1995; *Seranne et al.*, 1995; *Séranne*, 1999].

3.1. Nature and Age of Sedimentary Cover of the Distal Margin

The basement is covered with some 8 km of Oligocene to Quaternary sediments. The structure of sedimentary deposits can be summarized as follows [*Gorini et al.*, 1993; *Bache et al.*, 2010]. The synrift (Oligocene-Aquitanian) sequences are covered by the postrift (from Burdigalian to Messinian), and the Messinian evaporites with their transparent facies are easily identified. The Pliocene-Quaternary sequence shows a uniform thickness in the deep margin with evidence for gravitational sliding above the salt layer, including synsedimentary normal faults and rollover structures upslope, as well as folds and diapirs downslope [*dos Reis et al.*, 2008].

Key observations to interpret the deep stratigraphy come from the Golfe du Lion Profond 2(GLP2) hole (location in Figures 1 and 2) located on a basement high, higher up on the margin [*Guennoc et al.*, 2000]: above a metamorphic basement, a first breccia cemented in continental conditions is overlain by Stampian (Rupelian) marine deposits followed by Burdigalian marines clays. Figure 5 shows the details of the deep

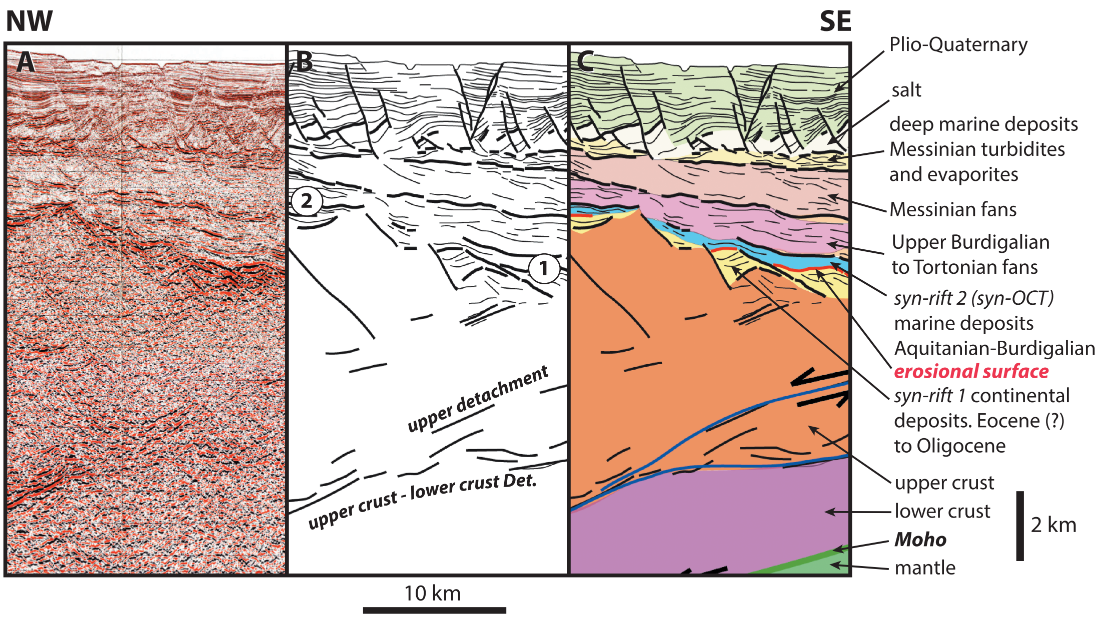

Figure 6. Details of a section of the distal margin where the sediments rest directly on exhumed lower crust (see details of more proximal and distal sections in Figures 5 and 7 and location in Figure 4). (left) The seismic profile without interpretation. (middle) Line drawing (1, 2, and 3 are the reflectors discussed in the text). (right) Interpretation.

margin section (see also Figures 6 and 7). Several reflectors separating domains with different seismic facies can be identified. The deepest sediments (yellow) show discontinuous high-amplitude, low-frequency reflections with a fan-shaped suggesting tilting of a crustal block. In the distal part of the profile, several tilted blocks are clearly imaged with synrift sediments in-between. The deepest sediments (yellow) are imaged by discontinuous high-amplitude, low-frequency reflections. These synrift deposits are fan-shaped and restricted to a series of basins, most of which are half-grabens above tilted basement blocks. The tilting of these basement blocks resulted in the apparent seaward onlaps of the synrift sequences. The exact base of the synrift sediments is unclear, and part of the yellow sequence could therefore correspond to reflections from the basement. The thickness of synrift sediments shown in Figures 4 and 5 should thus be taken as a

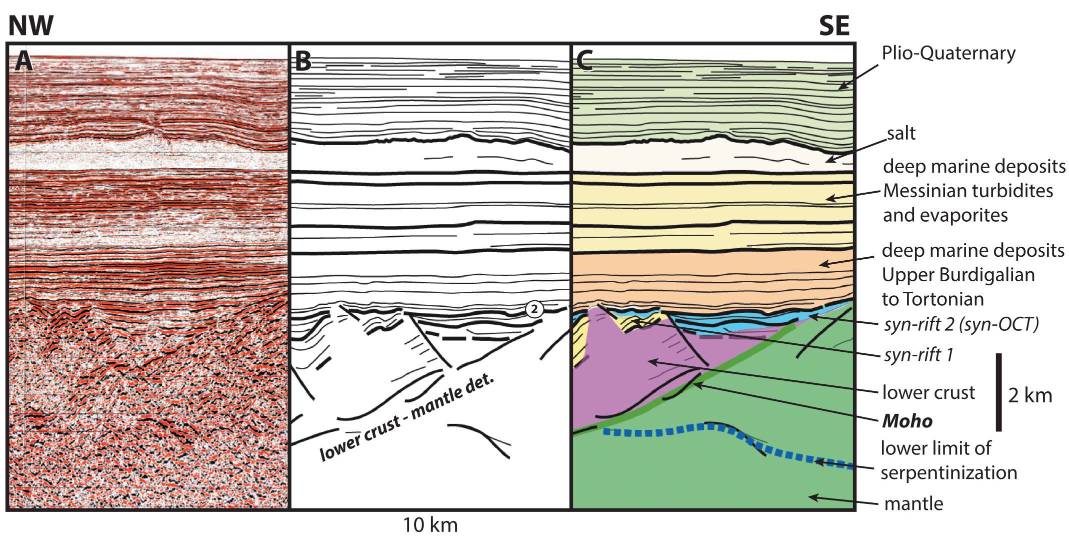

Figure 7. Details of a section of the distal margin where sediments rests directly on exhumed mantle (see details of more proximal sections in Figures 5 and 6 and location in Figure 4). (left) The seismic profile without interpretation. (middle) Line drawing (1, 2, and 3 are the reflectors discussed in the text). (right) Interpretation.

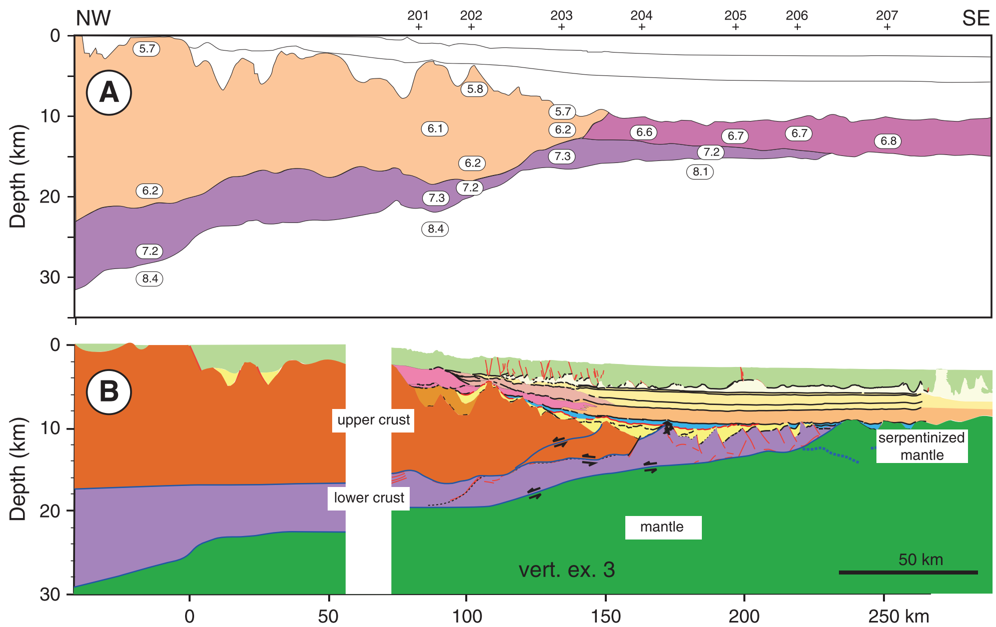

Figure 8. A cross section of the Gulf of Lion margin including the TGS-NOPEC profile (B), and comparison with the seismic velocities along the ECORS profile (A) [*Gorini et al.*, 1994]. The numbers in the profile are *P* wave seismic velocities are shown in km/s. Difference in interpretation stems from the spacing of ESP and the lesser quality of the ECORS profile compared to the new profile studied here.

maximum estimate. They are overlain by horizontal sediments (blue) with discontinuous low-amplitude reflections, interlayered with more transparent sediments. The sediments are laterally continuous and do not show any tilting. The basal contact with the synrift breccia (reflector 1) can be followed across a large portion of the profile all the way to the GLP2 drill hole. Clear truncations of the lower series by the upper series across reflector 1 indicate a phase of erosion, probably in subaerial conditions. This erosional surface can also be deduced from truncations higher up on the margin (Figure 6, right side). This phase of subaerial erosion has already been recognized higher up on the margin by *Bache et al.* [2010]. The blue sequence is then covered by a sequence of more continuous reflections (violet), with high amplitude near the base (reflector 2), present all over the lower part of the profile, grading into more transparent sediments toward the northwest when approaching GLP2, in which these sediments correspond to the Burdigalian marine clays that rest directly on top of the drilled breccia. The blue interval sandwiched between reflectors 1 and 2 is not present in GLP2, and the Burdigalian sediments rest directly on top of the erosion surface and the Stampian marine deposits. One can thus attribute the lower sedimentary sequences seen on the seismic profile to the following intervals, from base to top.

The lower sequence, below the erosion surface (reflector 1), corresponds to synrift continental deposits, similar to the continental breccia drilled at GLP2. After a phase of erosion, a first postrift marine flooding has deposited shallow water marine deposits (blue interval), contemporaneous with the Stampian (Rupelian) marine deposits of GLP2. This sequence lasts until the Lower Burdigalian; it corresponds to the initiation of the deep basin. Then, a second flooding event deposited well-stratified sediments, probably starting with shallow water limestones seen as the prominent reflections of reflector 2 and passing progressively upward into deeper marine deposits, regularly stratified, from the Burdigalian upward. The "blue" sequence is deposited on the basement all the way to the southernmost part of the profile, while the "yellow" synrift sequence is not seen in the southernmost half-grabens (Figure 7).

The blue interval (Aquitanian to Burdigalian) corresponds to a transition from the synrift to the postrift (*syn-ocean-continent transition, syn-OCT sediments* hereafter). A similar transition can be seen onland with the same timing on the Provençal margin [*Oudet et al.*, 2010]. On the Sardinian side of the rift, Oligocene (Rupelian-Chattian) subaerial clastics (Ussana formation), contemporaneous with the activity of the volcanic arc, are covered with transgressive early Aquitanian marine deposits [*Casula et al.*, 2001], coeval with the blue interval on the deep Gulf of Lion margin.

The seismic basement seen on the distal Gulf of Lion margin was thus covered with synextension continental deposits filling half-grabens before a marine transgression at the end of Oligocene or earliest Miocene. A phase of subaerial erosion is also registered in the distal margin. These observations are in favor of a crustal nature for the basement of the distal margin. Various interpretations have been discussed so far in the literature.

3.2. Deep Reflections and Nature of Basement

Two prominent series of reflectors can be seen in the deep part of the profile below the sediments. The first series of discontinuous and high-amplitude reflections is located below the half-grabens and synrift deposits of the distal margin (lower crust-mantle detachment; Figures 4 and 7). This reflector shows a shallow dip toward the continent and ramps up to the base of the syn-OCT sediments (blue color in Figures 4–7). Farther south, two half-grabens that are filled with the same marine transgressive sequence have an opposite polarity. The point where the lower crust-mantle detachment reflector is unconformably covered with the marine deposits corresponds approximately with the end of a basement block of abnormally high seismic velocities [*Pascal et al.*, 1993] (Figure 8) in the deep part of the distal margin (see discussion below on these abnormal velocities). Typical mantle velocities are reported, and the velocity contrast matches the lower crust-mantle detachment high-amplitude reflections, suggesting that the lower crust-mantle detachment is the local Moho, thus implying that the tilted blocks further to the south are made of upper mantle lithologies.

The second series of reflections (Figures 4–7) is observed farther north with a similar low northward dip. Several distinct shallow-dipping discontinuous reflectors with a ramp and flat geometry are similar to a continuous reflector seen on the ECORS 1 and 4 profiles [*De Voogd et al.*, 1991]. This series of reflection is well known as the T reflector, commonly observed on passive margins, sometimes also named the S reflector and interpreted either as the contact between the upper and lower crusts or as an interface within the lower crust [*Le Pichon and Barbier*, 1987; *Gorini et al.*, 1994; *Mauffret et al.*, 1995; *Gernigon et al.*, 2004]. In the case of the Gulf of Lion, the T reflector is interpreted as the top of an abnormally high-velocity layer that is diversely interpreted as lower crust or a mixture of lower crustal material and serpentinized mantle [*Gailler et al.*, 2009; *Bache et al.*, 2010; *Aslanian et al.*, 2012].

This contact merges toward the northwest with a series of more or less continuous reflectors (upper detachment (UD)) connected to a major shallow-dipping normal fault controlling a syn-ocean-continent transition half-graben. Upslope, the margin is cut by several normal faults bounding upper crustal blocks and synrift sediments (Figures 4 and 7).

Previously acquired data on the *P* wave seismic velocity structure give indications on the nature of the basement in the distal part of the margin, in the transitional domain between the continental margin (domain I) and the characteristic oceanic crust (domain III) (Figures 4). A comparison (Figure 8) between the velocity structure of *Pascal et al.* [1993] and our interpretation of the Petroceltic International PLC seismic profile shows that the domain of lower continental crust corresponds to a portion of the crust, in which the upper part has quite high *P* wave velocities (6.6 km s^{-1}), higher than expected for the upper crust, while the lower part has velocities similar to those of the lower crust imaged farther inland. The nature of the crust in this transitional domain between continental and oceanic crusts has been more recently studied and discussed in details by *Gailler et al.* [2009] (domain II). Data from the expanding spread profiles [*Contrucci et al.*, 2001] and the more recent *Sardinia* refraction cruise [*Gailler et al.*, 2009] show both abnormal high *P* wave velocities in this portion of the crust, suggesting a lower continental crust origin or exhumed mantle. *P* wave velocities vary between 6.6 and 7.2 km s^{-1} and show a high vertical gradient and no specific double reflection as underplated bodies would show [*Watremez et al.*, 2011], which may be compatible with exhumed lower continental crust [*Christensen and Mooney*, 1995; *Gailler et al.*, 2009]. It is furthermore characterized by unusually high velocities in the deepest parts, up to 7.9 km/s. An equivalent domain is found on the Sardinian side but twice narrower than on the Gulf of Lion side. *Gailler et al.* [2009] have discussed various hypotheses for the nature of this anomalous crust. They analyzed five different scenarios to explain the exceptionally high velocities in the transitional domain (1) magmatic underplating, (2) thin continental crust overlying serpentinized mantle, (3) thin oceanic crust overlying a serpentinized mantle, (4) upper mantle material exhumed during initial opening of the basin, and (5) lower crustal material or mixed lower crust continental material and serpentinized mantle. The unusually thin oceanic crust in this region suggests that the asthenosphere was not exceptionally hot during rifting, which does not favor magmatic underplating. This conclusion is further confirmed by the characteristics of the OBS sections: the characteristics of *PmP* arrivals(crust-mantle boundary) are not simply compatible with scenarios 2, 3, and 4, and *Gailler et al.* [2009] conclude that scenario 5 (lower crustal material or mixed continental lower crust and oceanic mantle) best fits their data, but they also warn that the resolution of their tomographic model is not sufficient to distinguish between the two possibilities.

Several additional arguments in favor of an exhumed lower crustal body can be put forward. First of all, on the seismic profile, this domain seems continuous with the lower crust present below the more proximal portion of the margin. Then, the distinct series of reflectors at its base are aligned with the Moho discontinuity below the margin, and they are likely to be reflections on the crust/mantle boundary (lower crust-mantle detachment) as discussed above. In the case of the Galicia margin, where exhumed mantle peridotites have been found in the transition zone, the basement is instead poorly reflective instead and there is no clear Moho reflection [*Reston*, 1996]. One more important point is the weak magnetization of this domain. According to *Sibuet et al.* [2006], serpentinization leads to a strong magnetization although other studies suggest instead that magnetic anomalies related to serpentinites are weak [*Oufi et al.*, 2002; *Leroy et al.*, 2010]. Finally, the presence of normal faults and tilted blocks with fan-shaped sedimentation seen on the seismic profile is also in favor of an extended continental crust. We thus prefer the solution where this anomalous body is entirely made of continental lower crust, which fits the observation of synrift Oligocene continental deposits on top of this piece of basement. Some amount of magmatic intrusion, difficult to appreciate, is still, however, possible as this domain was in the vicinity of the magmatic arc during its exhumation. It would give a peculiar seismic character to this domain with respect to the lower crust, still present below the margin further inland.

In summary, following our interpretation, the profile shows the succession of a domain with progressively thinner upper continental crust oceanward and strongly attenuated lower crust (domain I; Figure 4), a domain of exhumed lower crust (domain II) and a domain of exhumed mantle, probably serpentinized, before the true oceanic crust. If this interpretation holds, the exhumation of lower crust and mantle requires extensional shear zones to accommodate the differential southeastward movement. The upper/lower crusts and lower crust/mantle transitions on the profile are associated with prominent reflections that could represent these hypothetic shear zones.

The exhumed lower continental crust is replaced oceanward by exhumed subcontinental upper mantle. *P* wave seismic velocities, however, suggest that this mantle material should be partly serpentinized [*Horen et al.*, 1996]. A distinct reflector located at ~12 km, which is a depth compatible with the seismic Moho discontinuity, may correspond to the transition between the serpentinized and fresh mantle (Figures 4, 7, and 8). The Moho has then probably been reactivated as a low-angle normal shear zone to exhume the mantle. We refer to this shear zone as the lower crust-mantle detachment (Figures 4 and 7).

If the high velocity domain corresponds to exhumed lower crust, then the reflections seen above (T reflector) may correspond to the low-angle normal shear zone responsible for the exhumation (upper-lower crust detachment, hereafter). The Moho depth below domain I, taken from the ECORS profile [*Pascal et al.*, 1993], shows a crustal stretching factor higher than what can be deduced from the upper crustal normal faults alone [*Bache et al.*, 2010], implying that part of the lower crust has been removed and displaced seaward during extension.

4. Discussion

4.1. A Synthetic Cross Section of the Margin and its Reconstruction

The synthetic section of the margin (Figures 8 and 9 for the same section without vertical exaggeration) from the coastline to the truly oceanic domain shows the structure of the continent-ocean transition. As mentioned by earlier workers [*Gorini et al.*, 1994; *Seranne et al.*, 1995; *Bache et al.*, 2010], thinning of the upper crust is very limited on most of the offshore domain until the northern extremity of the Petroceltic International PLC profile. Farther south, the upper crust is strongly thinned within a distance of approximately 150 km, and it disappears totally 100 km before the first oceanic crust is encountered. In an opposite way, the lower crust is strongly thinned on most of the section and most particularly below the toe of the upper crustal wedge, where it is reduced to less than 3 km. The lower crust slightly thickens again farther to the southeast where it is exhumed, forming a 80 km wide complex bounded upward by the shallow-dipping

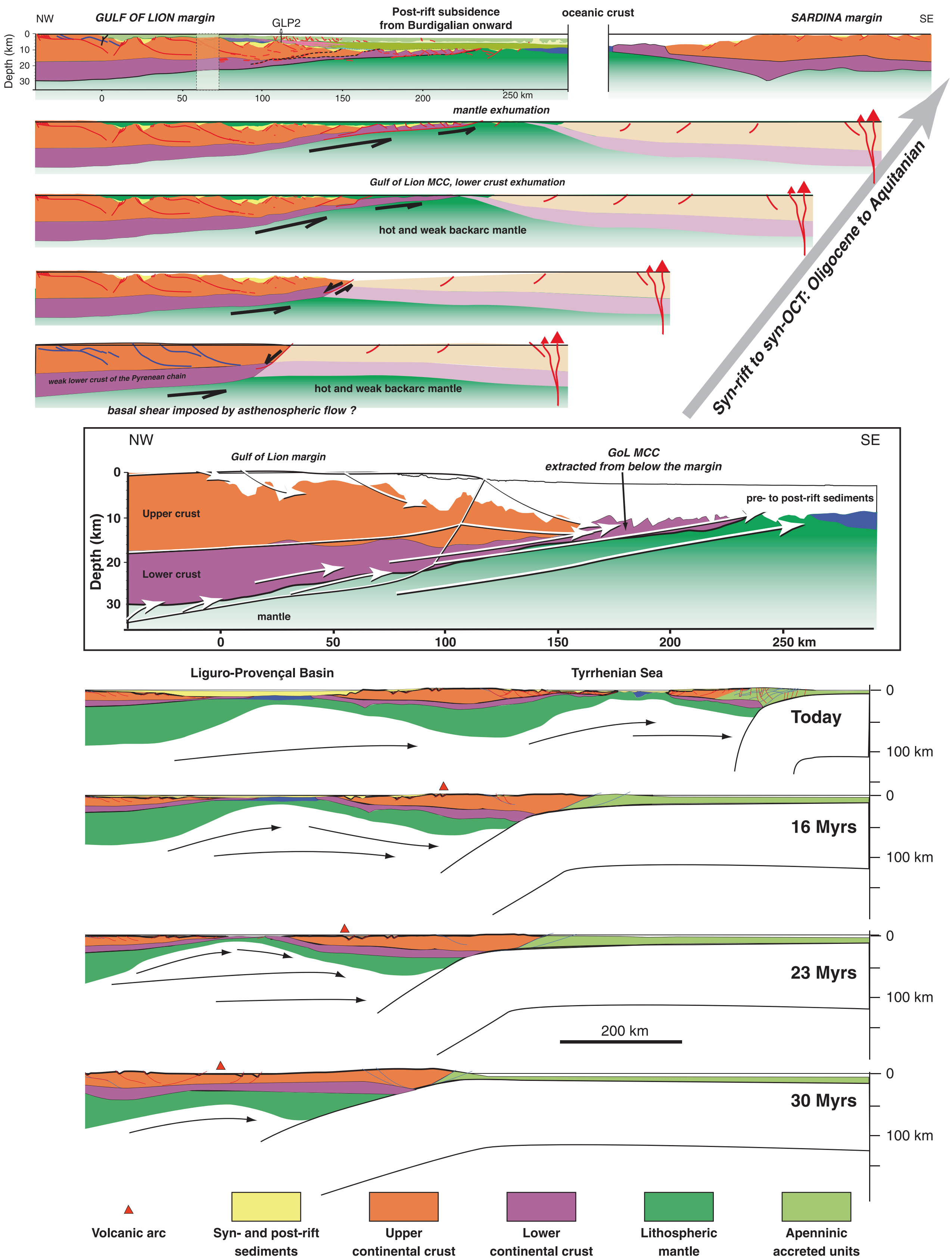
NW
GULF OF LION margin
GLP2
Post-rift subsidence
from Burdigalian onward
oceanic crust
SARDINA margin
SE
Depth (km)
0
10
20
30
0
50
100
150
200
250 km
mantle exhumation
Gulf of Lion MCC, lower crust exhumation
hot and weak backarc mantle
weak lower crust of the Pyrenean chain
hot and weak backarc mantle
basal shear imposed by asthenospheric flow ?
Syn-rift to syn-OCT: Oligocene to Aquitanian
NW
SE
Gulf of Lion margin
GoL MCC
extracted from below the margin
pre- to post-rift sediments
Upper crust
Lower crust
mantle
Depth (km)
0
10
20
30
0
50
100
150
200
250 km
Liguro-Provençal Basin
Tyrrhenian Sea
Today
16 Myrs
23 Myrs
30 Myrs
0
100 km
200 km
Volcanic arc
Syn- and post-rift sediments
Upper continental crust
Lower continental crust
Lithospheric mantle
Apenninic accreted units

Figure 9

detachment (upper-lower crust detachment). It can be compared to a metamorphic core complex with the exhumed lower crustal body as the core of the metamorphic core complex and the upper crustal material, the upper plate above the upper-lower crust detachment. This detachment has brought into close contact crustal materials with different maximum metamorphic pressure and temperature with a pressure gap across. The reconstructed cross section of the margin shows that the lower crust has been much more intensely thinned than the upper crust, and some lower crust is missing below the upper crust. The piece of lower crust observed in the GoL metamorphic core complex may correspond to that missing farther inland and may thus have been dragged from below the margin. A second core complex made of exhumed mantle is found before the true oceanic crust, it is probably made of serpentinized mantle, and it is topped by the lower crust-mantle detachment.

The GoL metamorphic core complex was cut after its exhumation by steep normal faults dipping toward the ocean, and synrift sediments were deposited in half-grabens. The actual thickness of these synrift sediments is sometimes difficult to ascertain in the deepest part of the margin, and our interpretation is more a maximum estimate than an accurate figure because some of the reflections could also correspond to some lower crustal fabric. The exhumed mantle is also cut by late steep faults that controlled the deposition of syn-ocean-continent transition sediments, but these faults dip toward the continent. This late brittle evolution of metamorphic core complexes evolving into a horst and graben structure is observed in several regions of postorogenic extension such as the Aegean [*Tirel et al.*, 2004]. The Aegean Sea was formed from the Oligocene onward after the Eocene formation of the Hellenic nappe stack. Although crustal thickening was likely much less significant in the case of the Pyrénées, extension was set at the same period in the Liguro-Provençal Basin right after the Pyrenean orogeny [*Jolivet and Faccenna*, 2000; *Lacombe and Jolivet*, 2005]. The Gulf of Lion and the Aegean Sea can thus be usefully compared. The erosional unconformity recognized at the top of the synrift sediments by *Bache et al.* [2010] is observed also at the top of the synrift sediments deposited on the GoL metamorphic core complex, implying that the exhumation of the core complex occurred in subaerial conditions before the fast postrift subsidence. A balanced reconstruction of the evolution of the margin (Figure 9) shows the progressive thinning of the crust and exhumation of the GoL metamorphic core complex during the rifting of the Liguro-Provençal Basin. The restoration shows that the lower crust has been more extended than the upper crust. As already mentioned by *Gailler et al.* [2009] (see also *Carminati et al.* [2004]), a pronounced asymmetry is observed between the GoL and Sardinian margins, the latter being much narrower, with a half as wide deep domain that is characterized by abnormally high seismic velocities.

4.2. Extraction of Lower Crust and Mantle During Rifting: The Gulf of Lion Metamorphic Core Complex

The analysis of the MCS profile in the continent-ocean transition thus shows deeper portions of the lithosphere toward the southeast with a maximum of ductile thinning accommodated by the lower crust, with prominent reflectors dipping toward the continent along the contact that we interpret as detachments. Furthermore, the portion of lower crust exhumed in the GoL metamorphic core complex is thicker than the portion still lying underneath the toe of upper crust, and the GoL metamorphic core complex is juxtaposed with exhumed serpentinized mantle farther to the southeast. Part of the lower crust has thus been transferred from below the thinned upper crust, where it is now partly missing, extracted by a distributed shear, and some localized deformation along the detachments (Figure 9).

The sequence of exhumation appears clearly on the seismic profile. The second synrift sequence (blue interval on Figures 4–7) rests directly on the mantle in the southern part of the profile (Figure 7), showing that the most distal part has been exhumed last. Moreover, the first synrift sequence is thinner in the south than in

Figure 9. Proposed evolution for the Gulf of Lion from Oligocene to Aquitanian (no vertical exaggeration). (top) Progressive thinning of the crust and exhumation of the GoL metamorphic core complex during the rifting of the Liguro-Provençal Basin. Note the asymmetry between the GoL and Sardinian margins, the latter being much narrower. The upper section (present time) corresponds to the interpretation of the margin structure, drawn here with no vertical exaggeration. The reconstruction shows that the crust was not so thick when rifting started despite the Pyrenean shortening episode. This observation confirms earlier works, suggesting that in Gulf of Lion, the Pyrenean orogeny was set on a thin crust (see *Lacombe and Jolivet* [2005] and references therein). (middle) Kinematics of exhumation of the lower crust and upper mantle that are more extended than the upper crust. (bottom) Lithospheric-scale cross sections showing the progressive retreat of the trench due to slab rollback and the opening of the Liguro-Provençal Basin and Tyrrhenian Sea. The arrows indicate the asthenospheric flow toward the retreating slab that entrains the weakened mantle and lower crust and exhumes the GoL metamorphic core complex.

the north, suggesting a progressive exhumation of the lower crust, followed by the mantle. The velocity of exhumation is more difficult to constrain as the age of the first synrift deposits is not precisely constrained. If one assumes that it was exhumed between ~35 and 23 Ma, 80 km of lower crust MCC were exhumed at an average velocity of ~0.6 cm/yr, which is a reasonable value for intracontinental rifts. The purely oceanic domain between Provence and Sardinia is about 250 km wide along a small circle running from the Gulf of Lion and Sardinia and was opened between ~24 and 15 Ma [*Gattacceca et al.*, 2007], which gives an average velocity of ~2.5 cm/yr. The timing of rotation of Sardinia implies still higher velocities in the southern part of the basin, and the Tyrrhenian Sea was opened also at much higher velocities over 10 cm/yr [*Nicolosi et al.*, 2006]. The Oligocene rifting was thus much slower than the true oceanic opening.

The subaerial conditions during the exhumation of the GoL metamorphic core complex and the very high finite stretching imply that the Moho was strongly uplifted during the extension. In classical metamorphic core complexes, the Moho remains rather flat and horizontal during the extension [*Wernicke*, 1992; *Tirel et al.*, 2004] and the weak lower crust is sucked into the dome between the boudins of the resisting upper crust. Here the lower crust is highly stretched in a ductile manner, but the crust is much thinner below the metamorphic core complex than farther toward the continent. An additional difference with classical metamorphic core complexes is the exhumation of serpentinized mantle at the end of the rifting process.

The Gulf of Lion margin has in common with Atlantic-type nonvolcanic passive margins, a highly attenuated continental crust in the ocean-continent transition zone corresponding to a wide and thick basin that started its evolution in subaerial conditions during rifting. The main difference is that in the Gulf of Lion, the attenuated domain is probably made of exhumed lower continental crust. The continent-ocean transition is made of exhumed lower continental crust and not of upper crust resting directly on top of the mantle like on the Angola margin [*Moulin et al.*, 2005], for instance.

4.3. Prerift Crustal Thickness and Thermal Regime

The absence of subsidence during rifting and the strong Moho uplift require a peculiar setting. A comparison can be made with the Woodlark basin (east of Papua New Guinea) and the exhumation of metamorphic core complexes [*Taylor and Huchon*, 2002]. In this region, an oceanic rift propagates westward in the back arc of the Trobriand trough and the continental crust is severely extended ahead of the tip of the propagator. This extension is superimposed onto a previous crustal thickening episode, and the conjunction of a thick crust and a hot environment produces metamorphic core complexes. As shown by *Abers et al.* [2002], at variance with ordinary metamorphic core complexes, the Moho is strongly uplifted below the extending zone, which is in isostatic equilibrium, implying that the extension process involves both the crust and the mantle. In the Gulf of Lion, we face a very similar situation with the Pyrenean crustal thickening episode during the Eocene [*Seranne et al.*, 1995] just before the beginning of back-arc rifting in the Oligocene, combined with the presence of an active volcanic arc nearby. The whole lithosphere has thus been thinned, and a wide metamorphic core complex formed by extraction of the lower crust from below the margin with little initial subsidence. It must be noted here that the exhumation of lower crust is ascertained only in the case of the Gulf of Lion margin. Farther east, the margin is much narrower and abrupt and no such structure is expected. As discussed by *Seranne et al.* [1995] or *Lacombe and Jolivet* [2005], the Gulf of Lion margin is located on the offshore extension of the Pyrénées. It also faced the well-developed Sardinia volcanic arc in the Oligocene (Figure 3) that was much smaller between Corsica and the Southern Alps. The conjunction of a thick crust and the high heat flow associated with the volcanic arc may have induced a weaker crust and a more distributed deformation in the Gulf of Lion than farther east (see further discussion below).

4.4. Slab Retreat and Asthenospheric Flow

The study of outcropping Mediterranean metamorphic core complexes in the Aegean, Tyrrhenian, and Alboran Seas reveals simple patterns of deformation in map view with consistent stretching directions over large areas [*Jolivet et al.*, 2008, 2009]. This simple pattern of deformation shows that the kinematics of stretching and the shear sense in the middle and lower crusts were constant over the entire back-arc domain. These directions are moreover systematically similar to those of the fast direction of *SKS* waves in the mantle [*Jolivet et al.*, 2009] (Figure 1). This suggests (1) that in the back-arc regions, the fast direction of seismic anisotropy is indeed a stretching direction in the asthenospheric mantle and (2) that the stretching directions in the crust and the mantle are the same. They are furthermore perpendicular to the trench and parallel to the expected direction of slab retreat, suggesting that the deformation in the crust is coupled with the flux of mantle induced by slab retreat [*Jolivet et al.*, 2009]. This coupling can be envisaged in two scenarios: (1) the mantle and the crust react to the same extensional stresses induced by slab retreat or (2) the extensional stresses are partly transmitted to the crust from the flowing asthenosphere as a response to slab retreat by a component of basal drag as proposed in *Jolivet et al.* [2009].

The Gulf of Lion is affected by the mantle flow due to slab retreat over its full width (Figure 1). The GoL metamorphic core complex was thus taken in the same stretching context as the other outcropping Mediterranean metamorphic core complexes. We thus now discuss these two scenarios, whether or not the mantle flow toward the trench during slab retreat may have provided the engine capable to extract the lower crust and serpentinized mantle from below the margin (scenario 2).

As discussed by *Hoïnk et al.* [2011], basal drag of the lithosphere by asthenospheric flow can work only with viscous coupling between asthenosphere and lithosphere and if the velocity of the flow underneath is faster than the motion of the upper plate. In the case of the Gulf of Lion, the second condition is most likely met. The Gulf of Lion margin is carried by the European plate, which moves slowly northeastward with respect to the mantle, and the asthenospheric flow is due to the retreat of the subducting slab below the Tyrrhenian Sea nowadays and below the nascent Liguro-Provençal Rift in the Oligocene. This retreat has always been much faster than the motion of Eurasia and in a different direction.

For shear stresses being transmitted from the flowing asthenosphere to the lower crust through viscous coupling with the lower crust, the resistant lithospheric mantle should be very thin or even absent, which for the Gulf of Lion is consistent with the inflow of hot asthenosphere from the NW due to the slab retreat. Without lithospheric mantle, the lower crust in direct contact with the asthenosphere would be at high temperature and thus as weak or even weaker than the flowing mantle underneath and could thus be entrained southeastward together with the flow.

As shown in Figure 3, there is no doubt that the shortened crust once continued eastward where the Gulf of Lion is now observed [*Seranne et al.*, 1995; *Chamot-Rooke et al.*, 1999]. The Pyrenees result from the Late Cretaceous to Eocene closure of a partly oceanic rift and the underthrusting of Iberia below the southern margin of Europe. Part of the shortening was taken up by the Iberian plate with south vergent thrusts and partly by the European crust with north vergent back thrusts with decreasing shortening from east to west [*Choukroune*, 1989; *Roure et al.*, 1989, 1996; *Vergés et al.*, 2002; *Chevrot et al.*, 2014; *Mouthereau et al.*, 2014]. This shortening event is also recorded in Provence at the same period [*Arthaud and Séguret*, 1981; *Lacombe et al.*, 1992; *Seranne et al.*, 1995] with a smaller amount of shortening than in the Pyrenees, but one has to add up the shortening involving the Variscan basement of Western Corsica. The presence of the shortened Pyrenean crust below the Gulf of Lion is thus ascertained, but reconstruction suggests that this crust was not very thick because of preshortening thinning episodes, as discussed by previous workers (see discussion in *Lacombe and Jolivet* [2005]). According to *Seranne et al.* [1995], a maximum relief of 1 km was present before the extension started. This part of the orogenic crust was, however, located in the back-arc region of the Oligocene subduction where arc volcanism provided further heating, like in Papua New Guinea at present [*Abers et al.*, 2002]. The deep parts of the orogenic continental crust would be weak and prone to easy flowing in such a warm environment.

The second scenario with basal drag would impose a noncoaxial component that should reflect is some asymmetry of the deformation at large scale. One of the characteristics of the Liguro-Provençal Basin transect is indeed its asymmetry with a narrower margin on the Sardinian side (Figure 9). A similar asymmetry can be observed in the Tyrrhenian Sea (Figure 9) with a much wider margin supported by a thinner crust on the Sardinian side than on the Campanian or Sicilian side [*Carminati et al.*, 2004; *Sartori et al.*, 2004]. During the two stages of extension of the Liguro-Provençal Basin, then of the Tyrrhenian Sea, the same asymmetry is repeated. It suggests that the driving mechanism involves a certain asymmetry too. Scenario 1 supposes that extensional stresses are transmitted from the retreating slab to the entire lithosphere without any special decoupling, while scenario 2 instead involves a component of basal drag that may be an explanation for the observed asymmetry. The observed asymmetry thus favors scenario 2 and suggests that the component of basal drag provided by the flow of mantle toward the trench is able to extract the lower crust and the lithospheric mantle from below the margin, leading to a wider continent-ocean transition zone on the Provence side (Figure 9).

The question of the age of the mantle fabric shown by seismic anisotropy can also be discussed. The main point is to know whether it corresponds to a still active flow in the asthenosphere or only to a frozen feature dating back to the rifting stage. The analyses of the orientation of the fast direction of anisotropy below the northern Tyrrhenian Sea led *Jolivet et al.* [2009] to conclude that the mantle fabric can be reset within a short time window, bracketed between 5 and 8 Ma, which is rather short compared to the protracted history of rifting in the western Mediterranean. A second observation is that after rifting and during southeastward slab migration, the previously thinned lithosphere has progressively reequilibrated at the expense of the asthenosphere and may thus have frozen the asthenospheric fabric. In the case of the western Mediterranean, there has not been any significant change in the direction of extension through time, from 35 Ma to the present. So there is no way to clearly separate active and frozen mantle fabric. It is possible that part of the seismic anisotropy signal corresponds to active asthenospheric flow below the entire region and partly to frozen fabric in the lithospheric rigid lid.

After the Late Cretaceous-Eocene period of crustal thickening that led to the formation of mountain belts form the Betic-Rif in the west to the Hellenides-Taurides in the east, postorogenic extension and metamorphic core complex formation is a common process in the Mediterranean realm, but they show significant differences with the Gulf of Lion. In the Aegean Sea and the Northern Tyrrhenian Sea, metamorphic core complexes formed in a more classical way as the thick crust of the Hellenides or the Corsica-Apennines chains run parallel to the subduction zone and they are now spread over the whole back-arc domain. The Moho is flat at 25 km below most of the Aegean Sea, especially below the Cycladic core complex [*Tirel et al.*, 2004]. In the case of the Gulf of Lion, the metamorphic core complex formed only where the Pyrenean chain met the back-arc domain and received additional heat from asthenospheric upwelling and arc volcanism. Elsewhere, back-arc rifting was set onto a normal crust, leading to a classical, more localized rifting. Following up on the comparison with the Aegean or the Basin and Range metamorphic core complexes, one can notice an evolution from a metamorphic core complex mode of extension toward a simpler horst and graben pattern once the ductile crust has reached the upper crustal levels and has cooled down, crossing the ductile to brittle transition during its exhumation. The same is seen in the GoL metamorphic core complex, and the 80 km wide core complex is dissected by a series of SE dipping normal faults and half-grabens.

5. Conclusion

The analysis of the TGS-NOPEC MCS profile thus shows that the continent-ocean transition of the Gulf of Lion is made of exhumed lower crust and upper mantle. The profile shows that the deep margin is divided into three domains: (1) a zone of crustal thickness gradient where the upper crust thins out and totally disappears, followed by (2) the margin toe where lower crust has been exhumed from below the margin by shallow north dipping detachments and then dissected by steep normal faults dipping southward, and finally, (3) through a transition zone where the subcontinental mantle is exhumed below a north dipping detachment that reactivates the Moho, the true oceanic crust. The exhumation of the lower crust dates back to the first stage of rifting from the Eocene to the Oligocene, while the exhumation of the mantle is more recent, dating back to the Aquitanian and Burdigalian. During rifting, the thinning crust was standing high as suggested by a subaerial erosional surface between the two stages of rifting. The exhumed lower crust is similar to a metamorphic core complex that we name the Gulf of Lion metamorphic core complex. During rifting, the average horizontal velocity of exhumation can be estimated around 0.6 cm/yr, thus much slower than during oceanic crust emplacement afterward. Two scenarios are then discussed for rifting mechanism, one where extensional stresses are transmitted through the whole lithospheric column from the retreating Apenninic slab and an alternative model where the asthenospheric flow induced by slab retreat extracts the lower crust and the mantle from below the margin. The general asymmetry of the Liguro-Provençal Basin and Tyrrhenian Sea favors the second scenario. The warm regime of the back-arc environment and the prerift orogenic episode are likely to make the lithosphere thin and weak and the lower crust highly ductile, which are favorable conditions for an efficient basal drag. The upper crust, more resistant, has been left behind and was only moderately thinned, thus proving an explanation to a long-debated observation. The efficiency of such a basal drag is not ascertained and merits further testing. However, the coupling between asthenospheric and lower crustal deformation seems quite strong in the Mediterranean back-arc regions as indicated by the comparison of stretching directions in metamorphic core complexes and the seismic anisotropy of *SKS* waves, suggesting that shear stresses due to asthenospheric flow toward retreating subduction zones can be transmitted up to the lower crust.

Acknowledgments
The authors wish to thank François Bache for stimulating discussions. They are also indebted to Leonardo Seeber and Michel Séranne, who provided insightful reviews that helped us prepare a better manuscript. This article is a contribution of the ERC Advanced Research grant 290864 (RHEOLITH). Special thanks are also due Petroceltic International PLC, which approved our use and publication of the seismic line, and TGS as the original acquirers and processors of the processed migration of the seismic line. Data are available at Petroceltic International PLC.

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