Deep crustal structure across a young passive margin from wide-angle and reflection seismic data (The SARDINIA Experiment) - I. Gulf of Lion's margin

Structure crustale d'une jeune marge passive à partir des données sismiques grand-angle et de réflexion (mission SARDINIA) – I. Marge du golfe du Lion

Moulin Maryline ¹, Klingelhofer Frauke ¹, Afilhado Alexandra ^{2,3}, Aslanian Daniel ¹, Schnurle Philippe ¹, Nouze Herve ¹, Rabineau Marina ⁴, Beslier Marie-Odile ⁵, Feld Aurelie ¹

¹ IFREMER, Dept Marine Geosci, Plouzane, France.

² Univ Lisbon, Fac Ciencias, IDL, P-1749016 Lisbon, Portugal.

³ ISEL, Lisbon, Portugal.

⁴ CNRS, IUEM, Domaines Ocean, F-29280 Plouzane, France.

⁵ Geosci Azur, F-06235 Villefranche Sur Mer, France.

Abstract :

The conjugate margins system of the Gulf of Lion and West Sardinia (GLWS) represents a unique natural laboratory for addressing fundamental questions about rifting due to its landlocked situation, its youth, its thick sedimentary layers, including prominent palaeo-marker such as the MSC event, and the amount of available data and multidisciplinary studies. The main goals of the SARDINIA experiment, were to (i) investigate the deep structure of the entire system within the two conjugate margins: the Gulf of Lion and West Sardinia, (ii) characterize the nature of the crust, and (iii) define the geometry of the basin and provide important constraints on its genesis. This paper presents the results of P-wave velocity modelling on three coincident near-vertical reflection multi-channel seismic (MCS) and wide-angle seismic profiles acquired in the Gulf of Lion, to a depth of 35 km. A companion paper [part II Afilhado et al., 2015] addresses the results of two other SARDINIA profiles located on the oriental conjugate West Sardinian margin.

Forward wide-angle modelling of both data sets confirms that the margin is characterised by three distinct domains following the onshore unthinned, 33 km-thick continental crust domain: Domain I is bounded by two necking zones, where the crust thins respectively from 30 to 20 and from 20 to 7 km over a width of about 170 km; the outermost necking is imprinted by the well-known T-reflector at its crustal base; Domain II is characterised by a 7 km-thick crust with anomalous velocities ranging from 6 to 7.5 km/s; it represents the transition between the thinned continental crust (Domain I) and a very thin (only 4-5 km) "atypical" oceanic crust (Domain III). In Domain II, the hypothesis of the presence of exhumed mantle is falsified by our results: this domain may likely consist of a thin exhumed lower

continental crust overlying a heterogeneous, intruded lower layer. Moreover, despite the difference in their magnetic signatures, Domains II and III present the very similar seismic velocities profiles, and we discuss the possibility of a connection between these two different domains.

Keywords : Continental passive margin, Crustal structure, Wide-angle seismic, Gulf of Lion, Oceanic crust, Transitional domain

Résumé :

Le système de marges conjuguées jeunes du golfe du Lion et de l'Ouest de la Sardaigne (GLWS), avec son set de données géophysiques important et de toute résolution, son épaisse série sédimentaire (dont le paléo-jalon correspondant à l'événement messinien (MSC)), ses reconstructions cinématiques bien contraintes constituent un laboratoire naturel pour répondre aux questions fondamentales sur le rifting.

Les principaux objectifs de la mission SARDINIA, étaient (i) d'étudier la structure profonde du système entier avec les deux marges conjuguées : le golfe du Lion et la Sardaigne occidentale, (ii) de caractériser la nature de la croûte, et (iii) de définir la géométrie du bassin pour fournir des contraintes importantes sur sa genèse. Cet article présente les résultats obtenus à partir de l'interprétation et de la modélisation conjointes de trois profils de sismique grand-angle et de réflexion (MCS) acquis dans le golfe du lion, jusqu'à une profondeur de 35 kilomètres. Les résultats de deux autres profils, localisés sur la marge conjuguée ouest sarde sont présentés dans un papier compagnon [partie II – Afilhado et al., 2015].

La modélisation directe de l'ensemble des données (grand-angle et sismique multi-canal) confirme que la marge est caractérisée par trois domaines distincts depuis la croûte continentale non-amincie de 33 km d'épaisseur : Domaine I, qui est limité par 2 necking zones où la croûte s'amincit respectivement de ~30 à 20 et de 20 à 7 kilomètres sur une largeur d'environ 170 kilomètres; cette dernière zone d'amincissement est marquée par la présence du réflecteur T à la base de la croûte ; le Domaine II est caractérisé par une croûte de 7 kilomètres d'épaisseur avec des vitesses sismiques « anormales » variant de 6 à 7,5 km/s ; il représente la transition entre la croûte continentale amincie (domaine I) et une croûte océanique « atypique » très mince (seulement 4-5 kilomètres) (Domaine III). Dans le Domaine II, l'hypothèse de la présence du manteau exhumé est réfutée par nos résultats : ce domaine est vraisemblablement composé d'une mince croûte continentale inférieure exhumée recouvrant une couche inférieure hétérogène et intrudée. A noter qu'en dépit de la différence dans leurs signatures magnétiques, les domaines II et III présentent des profils de vitesses sismiques très similaires, et nous discutons la possibilité d'une connexion entre ces deux domaines

Mots-clés : Marge continentale passive, Structure crustale, Réfraction, Golfe du Lion, Croûte océanique, Domaine transitionnel

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1 Introduction

The Liguro–Provençal basin corresponds to a pair of rifted margins, formed by the rupture of the Corso–Sardinian micro-continent with respect to the Ibero–European plate since the end of the Eocene, in the context of collision between Africa and Europe (Olivet, 1996). Conjugate margins of the Gulf of Lion and West Sardinia represent a unique natural laboratory for addressing fundamental questions on the formation of continental margins (i.e. crustal thinning, nature of the continent–ocean transition zone, style of rifting and subsequent evolution, and connections between deep and surface processes): i) these margins are young; ii) an important amount of stratigraphic and tectonic data and studies are available, both onshore (e.g. Clauzon, 1978; Benedicto *et al.*, 1996; Oudet *et al.*, 2010) and offshore (e.g. Aloisi, 1986; Droz & Bellaiche, 1986; Gorini *et al.*, 1993; Lofi *et al.*, 2003; Baztan *et al.*, 2005; Rabineau *et al.*, 2005, 2006, 2014; Jouet *et al.*, 2006; Bassetti *et al.*, 2006; Droz *et al.*, 2006; Bache *et al.*, 2009, 2010 2012; Leroux *et al.*, 2014, Leroux *et al.*, *submitted*); iii) plate kinematic

reconstructions are well constrained (Smith, 1971; Auzende *et al.*, 1973; Dewey *et al.*, 1973; Olivet, 1996; Gueguen *et al.*, 1998; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007); iv) the presence of the Messinian reflector provides the palaeo-geometry of the basin 5.3 Ma ago (at the end of the Messinian event) (Guennoc, 2000; Bache *et al.*, 2009; Lofi *et al.*, 2011); and v) last but not least, this pair of margins are very close and accessible.

During the SARDINIA experiment on the French R/V L'Atalante, six coincident wide-angle and near-vertical reflection multi-channel seismic (MCS) profiles were acquired, together with high resolution sub-bottom (CHIRP) profiles and swath bathymetry. Three profiles are located on the Gulf of Lion margin and three on the conjugate margin offshore Sardinia (see Afilhado *et al.*, this issue). The main objectives of the SARDINIA seismic experiment, conducted in December 2006 in the Provencal Basin (Figure 1B), were:

- to image the deep crustal and upper mantle structure of this young paired margin along exactly conjugate profiles. Whilst expanding spread profiles (ESP) were acquired on the Gulf of Lion (Le Douaran *et al.*, 1984; Pascal *et al.*, 1993; Contrucci *et al.*, 2001), no deep information existed so far on the conjugate Sardinian margin;
- to characterize the nature of the crust in the Ocean-Continent Transition (OCT) zone.

In 2009, Gailler *et al.* published a first tomographic, unfocused image of the Gulf of Lion-Sardinia system with one Sardinia's profile from each side of the two conjugate margins. The present paper details the results of wide-angle seismic modeling, using all the seismic arrivals, of the three profiles acquired on the Gulf of Lion's margin. The results of wide-angle seismic modelling on the West Sardinia margin will be discussed in the companion paper of Afilhado *et al.* (this issue).

2 Geological setting

In a context of convergence between the African and European plates, local extension, surrounded by thrusting and mountain building, occurred in Western Europe during the Oligocene (Le Pichon *et al.*, 1971). The Liguro-Provencal basin (or North-Western

Mediterranean basin) is one of those areas. It is located at the eastern end of the Pyrenees and the southern end of the Eocene West European Rift system (Séranne, 1999; Bache *et al.*, 2010), and was formed by the anticlockwise rotation of the Corsica-Sardinia block (Figure 1A). The opening of the Liguro-Provençal Basin started at the end of the Eocene (Priabonian: 33.7 Ma ago) and the anticlockwise rotation of the Corsica-Sardinia block stopped between 23 and 19 Ma ago (Edel, 1980; Réhault *et al.*, 1984; Ferrandini *et al.*, 2003; Gattacceca *et al.*, 2007), in the general context of collision between Africa and Europe (Smith, 1971; Auzende *et al.*, 1973; Dewey *et al.*, 1973; Olivet, 1996; Gueguen *et al.*, 1998; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007). This opening was followed by the formation of the Tyrrhenian Sea, in the back-arc context of the south-eastward retreating Apennines-Maghrebides subduction zone (Auzende *et al.*, 1973; Boccaletti & Guazzone, 1974; Réhault *et al.*, 1984; Malinverno & Ryan, 1986; Jolivet and Faccenna, 2000; Jolivet *et al.*, 2006). Although rifting in back-arc basins might differ somewhat from cratonic rifting, the mechanics of the thinning of the continental crust, which is the first-order question, ought to be similar. Moreover, despite some differences on the subsidence rates (Séranne, 1999), the palinspastic reconstruction of the Liguro-Provençal basin shows the same initial configuration (Aslanian *et al.*, 2012; Rabineau *et al.*, 2014) as those of Central (Labails *et al.*, 2009) and South Atlantic margins (Aslanian *et al.*, 2009; Aslanian & Moulin, 2012) as well as intracontinental basins (Thybo & Nielsen, 2009).

This GLWS is bounded by the Provence Margin to the North, the Catalan margin to the West, the Ligurian Sea to the East, and the Valencia Trough to the South-West. The formation of the Valencia Trough is consecutive to the Liguro-Provençal basin genesis (Stoekinger, 1976a, 1976b; Garcia-Sineriz *et al.*, 1978; Soler *et al.*, 1983; Johns *et al.*, 1989; Clavell & Berastegui, 1991; Maillard *et al.*, 1992; Roca *et al.*, 1999). Whilst the neighboring Valencia Trough is underlain by thinned continental crust and represents an aborted rifted basin (Watts *et al.*, 1990; Torné *et al.*, 1992; Pascal *et al.*, 1993; Mauffret *et al.*, 1992; Maillard *et al.*, 1992; Collier *et al.*, 1994; Sabat *et al.*, 1995; Ayala *et al.*, 1996; Maillard & Mauffret, 1999), the central part of the Liguro-Provençal basin is assumed to be underlain by oceanic crust (Burrus, 1984; Le Douaran *et al.*, 1984; De Voogd *et al.*, 1991; Pascal *et al.*, 1993; Rollet *et al.*, 2002; Gailler *et al.*, 2009),

which would have initiated in the Late Aquitanian / Early Burdigalian (between 23 and 19 Ma), 9 Ma after the beginning of rifting (Gorini *et al.*, 1993; Mauffret *et al.*, 1995; Séranne, 1999; Guennoc *et al.*, 2000; Bache *et al.*, 2010).

The post-rift sedimentary evolution is well documented due to intensive onshore and offshore geophysical and geological studies conducted by the French program Action-Marges over the last few decades, as well as from stratigraphical and lithological information based on various wells. The post-rift sequence can be divided into three distinct depositional episodes: i) the Miocene sequence; ii) the Messinian sequence, related to major sea level fall in the Mediterranean after its isolation from Atlantic waters (Hsü *et al.*, 1973; Cita & Gartner, 1973; Clauzon, 1973; Ryan, 1976) and composed of detritics and massive evaporites (e.g. Ryan, 1973; Bache *et al.*, 2009), and iii) the Pliocene-quaternary sequence (Lofi *et al.*, 2003, Rabineau *et al.*, 2014).

Based on magnetic and gravity maps as well as seismic profiles (Bayer *et al.*, 1973; Galdéano & Rossignol, 1977; Rollet *et al.*, 2002; Bache *et al.*, 2010, Gailler *et al.*, 2009), five distinct domains (from I to V) (Figure 1A and blow-up of domains I to III on Figure 1B) can be distinguished in the Liguro-Provençal basin, with a clear symmetrical distribution (Gueguen, 1995; Gailler *et al.*, 2009) which refutes the hypothesis of simple shear mechanism along a lithospheric detachment fault as noticed by Gailler *et al.* (2009):

- Close to the coast, domains I & V, on the Gulf of Lion and West Sardinia margins, respectively, correspond to the « continental slope domain ». In the Gulf of Lion, Domain I is characterized by two contrasting zones: i) Domain Ia corresponds to a thinned and poorly stretched wide domain, characterized by a deficit of subsidence during the rifting phase, while ii) Domain Ib corresponds to a thinned and stretched narrow domain showing a tilted blocks zone (Bache *et al.*, 2010). The seaward limit of the Domain Ib, the end of the true continental domain, is highlighted by the presence of a strong, high amplitude, landward dipping seismic reflector, named the T-reflector, on the ECORS profiles (De Voogd *et al.*, 1991).

- In the center part of the basin, Domain III is characterized by a magnetic anomaly

pattern that sharply differs from the quiet adjacent domains (Bayer *et al.*, 1973; Galdéano & Rossignol, 1977), a concomitant low gravity anomaly, and coincides with the occurrence of tightly connected salt diapirs. The crust is assumed to be of oceanic nature (Burrus, 1984; Le Douaran *et al.*, 1984; De Voogd *et al.*, 1991; Pascal *et al.*, 1993; Rollet *et al.*, 2002; Gailler *et al.*, 2009).

• In between Domain III and Domains I & V, Domains II and IV correspond, respectively on the Gulf of Lion and West Sardinian margins, to the transitional domain, also called the Ocean-Continent Transition (OCT) as it was initially described on the Galicia margin by Boillot *et al.* (1980). The crustal nature of these domains is still a matter of debate. The study of 30 expanding spread profiles (ESP) shows a 20 – 40 km wide zone in the Liguro-Provençal Basin with high crustal velocities, that are neither from continental material nor upper mantle material (Le Douaran *et al.*, 1984; Pascal *et al.*, 1993; Contrucci *et al.*, 2001), as also observed in the adjacent North Ligurian Margin (Dessa *et al.*, 2011). Note that these high velocities are not observed on the ESP located in the « aborted » Valencia Basin (Watts *et al.*, 1990; Torné *et al.*, 1992; Pascal *et al.*, 1992; Mauffret *et al.*, 1992; Maillard *et al.*, 1992; Collier *et al.*, 1994; Maillard *et al.*, 1999).

Several hypotheses were proposed to explain these anomalous velocity zones in the Liguro-Provençal Basin:

- 1) continental crust which was extremely thinned, possibly broken up, underplated, and intruded by partial melt (Pascal *et al.*, 1993) ;
- 2) exhumed mantle material, serpentinized by water through cracks/faults (Le Douaran *et al.*, 1984; De Voogd *et al.*, 1991; Pascal *et al.*, 1993; Guennoc *et al.*, 1994; Mauffret *et al.*, 1995; Chamot-Rooke *et al.*, 1999; Dessa *et al.*, 2011);
- 3) exhumed lower crustal material, possibly intruded by upper mantle material (Séranne, 1999; Gailler *et al.* 2009; Bache *et al.* 2010), as was proposed for the African/Brazilian Margins system (Aslanian *et al.*, 2009).

3. Data acquisition, quality and processing

During the SARDINIA experiment, three coincident MCS reflection and wide-angle

seismic profiles, sub-bottom high resolution (CHIRP) profiles, and bathymetry data were collected on the Gulf of Lion margin. The seismic reflection data were acquired using a 4.5 km, 360 trace digital streamer and a tuned airgun array of 8260 in³, towed at a depth of 18 – 28 m. A total of 57 ocean bottom seismometer/hydrophones (OBS/OBH) from Ifremer (Auffret *et al.*, 2004), University of Brest, and Geomar Kiel were deployed in the Gulf of Lion, spaced every 7 Nm (~13 km). The seismic profile perpendicular to the margin was extended on land by 9 land seismic stations (Figure 1B). Two OBH were moored on 900 m lines to record the far field signature of the airgun array, which consisted of 16 airguns ranging from 100 in³ G-guns to 16 L Bolt airguns, with main frequencies centered around 10 – 15 Hz. The airguns were tuned to the first bubble to enhance low frequencies and ensure a good penetration (Avedik *et al.*, 1993). The shot interval was 60 s at an average speed of 5 knots, which translates to a trace spacing of about 140 – 150 m. The sample rate was 8 ms for the micrOBS, 5 ms for the Geomar's OBS and 4 ms for the OldOBS. A total of 6160 shots (profile AB: 3032, profile CD: 1730, profile EF: 1398) were fired by the air gun array.

Profile AB (Figure 1B) crosses the Gulf of Lion's margin, parallel to the ECORS profile, from the continental shelf to Domain III. Twenty-five ocean bottom seismometers (OBS) were deployed on this profile, of 467 km in length. Twenty-two instruments were used in the wide-angle modelling, since three instruments (OBS 45, 49 and 52) did not properly record. The shots on this profile were additionally recorded by 9 land seismic stations (OSIRIS stations from the University of Brest and Geosciences Azur) that have extended the marine profile ~120 km onshore.

Profiles CD and EF (Figure 1B) are 250 and 210 km long transects parallel to the margin. Profile CD was acquired using 19 OBS of which 18 yielded usable data. Profile EF was acquired using 15 OBS of which 12 yielded usable data. These two profiles cross the profile AB, respectively at OBS 26 (Figures Annex-1 and Annex-2) and OBS 8 (Figures Annex-3 and Annex-4).

Instrument locations were corrected for drift from the deployment position during their descent to the seafloor using the direct water wave arrival. Instruments in water depths of less than 100 m were not corrected, as the expected drift is small. Furthermore, instruments that recorded no close shots were not corrected if no direct

water wave arrival could be picked as a first arrival. The drift of the instruments even in deep water never exceeded 200 m. Picking of the onset of first and later arrivals was performed without band-pass filtering or applying a zero-phase Butterworth band-pass filter, to avoid time delay. When necessary, a different frequency range was set for the band-pass filter. Arrivals from longer offsets are of lower frequency compared to short offset arrivals, so the band-pass frequencies were chosen appropriately.

Data quality from the seafloor instruments is equally good on the continental part and oceanic part of the margin's perpendicular profile (Profile AB) (Figure Annex-5). Useful arrivals could be picked up to a 100 – 160 km source-receiver offset, including arrivals reflected from the Moho (PmP) and refracted in the shallow mantle (Pn). Five of the nine land seismic stations positioned on the prolongation of Profile AB gave high quality data, with clear lower crustal and shallow mantle arrivals, however the other four stations (GDL02, GDL04, GDL06 and GDL08) gave no data at all (Figure 1). On Profiles CD and EF the data quality is comparable to that of Profile AB (Figure Annex-5). Profiles CD and EF, located on thin crust and undisturbed sediments, exhibit very good data quality with clear intra-crustal reflection (Figure Annex-6) and clear PmP and Pn arrivals on all stations (Figure Annex-7).

Processing of the multichannel seismic data was performed using the Geovecteur processing package. The processing sequence consisted of: data geometry and binning, noise editing, band-pass filtering (2-8-48-64 Hz), spherical divergence amplitude correction, FK multiple attenuation, external and internal mute, dynamic amplitude correction, and stacking. Velocity analysis was performed every 200 CDP for the final stack. Finally, a post-stack Kirchhoff time migration using stacking velocities was applied.

4 Wide-angle Modelling

The OBS data were modelled using the inversion and ray-tracing algorithm of Zelt and Smith (1992). Modelling was performed using a layer-stripping approach, starting from the top of the structure (sea-bottom) towards the bottom (Moho). The layer boundaries were also constrained by their TWT, picked from the seismic reflection record section, and were iteratively adjusted until an acceptable fit of the data was obtained. However,

only reflectors correlated to reflected phases identified in the OBS data were included to avoid over-parameterization of the model. The velocity gradients set in the velocity model were further constrained by amplitude modelling, using the asymptotic approach of ray theory to compute a synthetic seismogram (Zelt & Ellis, 1988).

The seafloor bathymetry was taken from the multibeam data where possible and otherwise from the echo-sounder data corrected for the water measurements. On land, topography was included from the altimetry measurements at each land seismic station. The two-way travel times of the main sedimentary layers, the top of the basement, and a few segments of clear reflections from the Moho were picked from the reflection seismic data. An iterative approach was followed to include these constraints in the modelling in order to guarantee the consistency of the velocity model with the OBS and MCS data sets. The picking interval in the sedimentary layer depends on the observed topography of the interfaces imaged on the seismic reflection record section. The model depth node interval in the deeper part was chosen between 10 and 20 km, as the OBS spacing will not allow a finer resolution on deep interfaces. All models represent a minimum structure modelling approach, therefore we did not include lateral velocity changes or layer topography changes where not required by the data.

Two of the three models (CD and EF) are parameterized by nine different layers, while the model AB is parameterized by eight different layers: the water layer, five (model CD, where it was possible to differentiate two layers in the Plio-Quaternary sediments) and four (models AB and EF) sedimentary layers, two (models AB and CD) and three (model EF) crustal layers, and the lithospheric mantle layer (blue lines on [Figure 2](#)). Each model layer is defined by depth and velocity nodes. Water velocities, set when correcting OBS location for drift from the deployment position, are 1.52 km/s on profiles AB and CD and 1.53 km/s on profile EF. Depth node spacing of the seafloor and sedimentary layers are set from picking of sediments of mega-sequences boundaries from the seismic reflection and vary between 1 and 10 km. For the crustal layers the depth node distance varies between 10 and 20 km depending on the depth of the interface.

The sedimentary layers have top and bottom seismic velocities of 2.05 – 3.30 km/s (Post-Messinian), 4.15 – 4.50 km/s (Messinian), 4.55 – 4.9 km/s, and 5.0 – 5.4 km/s (both Pre-Messinian) on Profile AB ([Figure 2](#)). Between 30 and 125 km the velocity of

the salt layer (Messinian) decreases landward to 3.40 – 3.50 km/s. This zone of velocity decrease matches the erosional (with smooth Messinian erosion surface) and intermediate province defined by Bache *et al.* (2009), where no salt was yet deposited during the Messinian. Furthermore, in the same area (Domain Ib), the velocities of the pre-salt layer remain lower (4.55 to 4.65 versus 4.65 to 4.75 km/s), as already observed on the ESP data and on the well GLP2 (Leroux, 2012). The model of profile AB comprises three basement layers: upper crust, lower crust, and lithospheric mantle (Figure 2). The upper crust shows velocities between 5.5 – 6.5 km/s on the proximal part (between 140 and 130 km) and 6.0 – 6.4 km/s on the distal part of the profile, which are well resolved, at least up to ~280 km (see discussion ahead, point 5). The lower crust is characterized by velocities between 6.4 and ~6.9 – 7.0 km/s in the proximal part (~140 – 130 km), ~6.6 – 6.7 and 7.5 km/s in the central part (130 – 300 km), and finally 6.5 and ~6.7 – 6.8 km/s in the most distal part of the profile. Lithospheric mantle velocities are constrained to 7.9 – 8.2 km/s along the entire model profile.

On the margin parallel profiles CD and EF, sedimentary and crustal layers show less lateral variation than on profile AB that is perpendicular to the margin. On the more proximal profile CD (Figure 2), water depth is around 2.5 km and the sedimentary thickness is around 7 km. Sedimentary velocities are very similar to the ones modeled in profile AB. The most recent sediments were split in two layers, with velocities of 2.05 – 2.15 km/s and 2.75 – 3.70 km/s (Post-Messinian), the older sediments were still modeled in three layers with velocities of 4.25 – 4.55 km/s (Messinian), 4.25 – 4.90 km/s, and 5.3 – 5.5 km/s (both Pre-Messinian). The upper crust in the center of the profile is characterized by velocities between 6.0 – 6.1 km/s, whereas on the extremities of the profile, velocities vary in the range of 5.8 – 6.1 km/s. Lower crust velocities change from 6.6 to 7.2 km/s in the center of the profile (80 – 240 km) to 6.5 – 6.9 km/s on the extremities. Finally, the lithospheric mantle velocity, constrained by Pn arrivals, is 8.0 km/s.

On the most distal profile EF (Figure 2), water depth is also around 2.5 km, with a bathymetric bulge in the center of the profile, corresponding to the Rhone delta, with a sedimentary thickness around 7 km. Sedimentary velocities are 1.85 – 3.70 km/s (Post-Messinian), 4.32 – 4.45 km/s (Messinian), 4.8 – 4.9 km/s, and 5.3 – 5.6 km/s (both

Pre-Messinian). The upper crust is modelled with velocities between 6.0 – 6.05 km/s, underneath a thin layer of 5.6 km/s on its top. Although having similar apparent velocity near the crossing AB-EF, we observe a systematic delay of the Pg arrival times along profile EF relative to the Pg arrival times along profile AB. To fit the delay of Pg arrivals, we inserted a 5.6 km/s thin layer in the top of the upper crust, that also produces an adequate fit of amplitudes (Figure Annex-4). Since no turning waves from this interface were identified, a lower velocity in the top of basement is also admissible, due to a possible local anisotropic shallow basement around EF. Lower crust velocities range between 6.8 and 7.5 km/s, changing to 6.8 – 7.2 km/s at the eastern end of the profiles. The lithospheric mantle velocity was constrained to be a constant 8.1 km/s by Pn arrivals along the entire model profile.

5 Error Analysis

The fit between predicted arrivals times and travel-time picks provides information about the quality of the model (Figures 3 – 5). The χ^2 is defined as the root-mean-square travel-time misfit between observed and calculated arrivals, normalized to the picking uncertainty. Estimated picking uncertainties were 75 ms for all arrivals. The number of picks, RMS travel-time residuals, and the χ^2 -error for all phases are listed in Tables 1-3 (for models AB, CD and EF respectively).

Two-point ray-tracing between source and receiver (Figures 3 – 6) shows well imaged and unconstrained areas for all models. On profile AB, ray coverage is very good, except for the top of the basement in the ocean-ward tip of the profile. Only near-offset reflected phases from the top of the basement, at model distances larger than 250 km are identified. In the distal part of the model (between 280 and 400 km model distance), the coverage is poor due to the presence of salt diapirs. All rays arriving at the land seismic stations emerge at very steep angles and therefore do not allow us to model the shallow sedimentary structures on land. On profile CD, ray coverage is very good, except in the western part of the model (between 20 and 90 km model distance) where the sedimentary and shallow crustal layers are not well covered due to the complex morphology of the basement. However, the lower crustal layer and lithospheric mantle are well covered along the entire profile. For profile EF, the coverage is excellent for

sedimentary and crustal layers along the entire profile.

Additional information about the quality of the velocity model can be gained from the model resolution (Figure 7) (Zelt & Smith, 1992). Resolution indicates how well a model parameter is constrained by the data. A high-resolution value for a particular model velocity or depth implies good confidence in this velocity or depth value. The resolution of one particular velocity node depends on the relative number and angular distribution of rays constraining that node. Layers with a low number of nodes can be well constrained by relatively few rays passing through the layer, while layers with many velocity nodes need a large number of rays passing through to achieve a similar resolution level. Velocity nodes having resolution greater than 0.5, corresponding to white and yellow areas in the model, are considered well resolved (Figure 7). Only a few regions along the profiles show a resolution of less than 0.5, which is considered poorly resolved, and are underlined in red. They correspond to the extremities of the profiles and to the upper crust between 290 and 400 km, where velocities are not constrained due to the absence of turning or reflected rays from the upper crust in this region (Figure 3).

We also performed a direct evaluation on the uncertainty of the Moho depth by increasing and decreasing its value until it was no longer possible to fit the observed arrival times within its uncertainty. A variation of ± 700 m implies an increase of 50 % of the RMS for Moho depth, which is considered the maximum acceptable arrival time misfit.

6 Comparison to reflection seismic data

Figure 8 presents the coherence between the depth to two-way travel-time conversion of the velocity models and the MCS seismic profiles.

Although the depth and velocities of the crustal layers and the lithospheric mantle were modelled exclusively from the OBS data, we find a good agreement of the velocity model interfaces with the deep reflectors identified on the seismic reflection section. This is particularly obvious in the ocean-continent transition zone (Domain II) due to the excellent quality of the seismic reflection data. On the dip AB profile, the basement

in Domain II is characterized by a strong and chaotic reflector, underlying a more transparent crustal layer, until 260 km where the salt screens the signal. On profiles EF and CD, the basement is less marked by this strong amplitude reflector but corresponds to the bottom of the highly stratified Miocene unit. At about ~ 8 s TWT depth, a strong energetic reflector marks the Moho on all the profiles; at the transition between Domains Ib and II, this high energetic reflector pinches out landwards at 145 km. This pinch-out coincides with the landward-dipping so called T reflector, well marked between 110 and 150 km, as already noticed by Pascal *et al.*, (1993), it also coincides with the top of the high velocities zone (thick white line on [Figure 9](#)).

7 Anisotropy of seismic velocity

In order to test a possible anisotropy of seismic velocity (Diaz *et al.*, 2013), within the crustal layer due to the movement of Sardinia-Corsica block, we tested two options: i) Model D, with a strong crustal gradient (6.6 to 7.2 km/s), in agreement with profile AB, and ii) model E, with a low crustal velocity gradient (6.8 to 7.0 km/s) which represents a considerable anisotropy with respect to profile AB. For the two models, the calculated and observed travel-times fit both visually and statistically, within the error bounds ([Figure 10](#)). In order to distinguish between these two hypotheses, the synthetic seismograms of the two options were calculated and compared with the data of OBS26, located at the intersection between profiles AB and CD ([Figure 10A](#)). The analysis of the synthetic seismograms reveals a much better fit for model (D), where the computed PmP phase stops at -50 km and 48 km offset the model ([Figure 11C](#)), as observed on data, whilst the cut-off of the PmP phase is outside the record section range in model (E) ([Figure 11B](#)). This result shows that no anisotropy is required by the data in the Gulf of Lion and demonstrates the importance of the amplitude analysis, especially concerning the PmP phase.

8 Discussion

The velocity models of the three SARDINIA-Gulf of Lion profiles ([Figure 11](#)) allow us

to make several observations on the sedimentary section, deep crustal structures, and the crust-mantle boundaries of the Gulf of Lion area.

Sedimentary layers. The thickness of sedimentary layers increases from 1.4 km (at OBS 35) near the shoreline up to 7.6 km around km-185. Beyond that point, it progressively thins southwards down to 6.3 km. Beyond km-330, the sedimentary thickness is constant. The sedimentary section is comprised of three main layers (*cf.* wide-angle modelling chapter 4). The basal layer: the lower Miocene sediments, above the basement, displays high velocities (5.0 – 5.4 km/s), which confirms the observations made by Pascal *et al.* (1993). This layer starts near the shoreline (OBS36) and disappears between OBS55 and OBS56, in Domain III. Its maximum thickness is located in Domain II (at km-185), reaching 2 km. This sedimentary sequence, first interpreted as pre-rift sediments by Le Douaran *et al.* (1984), has then been interpreted to be syn-rift or post-rift sediments (Pascal *et al.*, 1993). For Bache *et al.* (2010) this layer represents post-rift Aquitanian to Tortonian deltaic deposits, characterized by sedimentation on a wide prograding shelf, with downlaps and progradations towards the basin. Such high velocity sedimentary layers are observed in various geodynamical contexts (rift, graben, margin) as in the Angola Margin (Contrucci *et al.*, 2004), Orphan Basin (Chian *et al.*, 2001), Newfoundland Grabens (Lau *et al.*, 2006), and the Baikal Rift (Thybo & Nielsen, 2009).

Crustal structure. Four main regions can be distinguished in the crustal structure of the Gulf of Lion margin (Figures 1, 2 and 11) as previously stated by Gailler *et al.* (2009) and Bache *et al.* (2010):

- A) The onshore domain of relatively unthinned continental crust (>30 km) from km -140 to km -40. This crust is divided into two layers (upper and lower crusts) of roughly similar thickness (16 and 17 km) with velocities of 5.5 – 6.5 and 6.4 and 6.9 – 7.0 km/s respectively (Figure 12A). Such velocities and crustal thickness are typical of continental crust (Christensen & Mooney, 1995) and are in good agreement with previous studies in this region (Sapin & Hirn, 1974).
- B) The “continental slope crust” (km -40 to km 130), which can be divided into three sub-domains (Figure 11), as proposed by Bache *et al.*, 2010: a necking zone, a thinned continental domain with constant thickness and a second

necking zone. i) The first 40 km-wide necking zone where the crustal thickness decreases from 29 km to 21 km, hereafter called the hinge zone. The two crustal layers are still present, but the main thinning occurs in the lower crust, where thickness decreases from 15 to 7 km. This domain did not subside. The seaward limit of this hinge zone corresponds to the onset of increasingly thick sedimentary deposits toward the basin. ii) a 90 km-wide region of thinned continental crust (about 20 km) with some tilted blocks. In that domain the lower crust has a constant thickness of 7 km, whilst the upper crust thins from 14.5 to 7 km. On the other hand, the lower crustal velocities increase oceanward, from 6.3 – 6.5 to 6.5 – 7 km/s, while the upper crustal velocities remain constant. This domain did not subside much during rifting and displays significant thinning of the crust, which cannot be explained by the stretching of the upper crust. Recently, Huisman & Beaumont (2014) showed that depth-dependant models of type II margins predict limited syn-rift subsidence because the crust decouples and stretches well into the continent. iii) A second necking zone where the upper and lower crust thins simultaneously, respectively from 7 to 5 km and from 7 to 3 km, in a narrow region of less than 40 km. The offshore limit of Domain Ib is marked by a prominent T-reflector (De Voogd *et al.*, 1991), a major fault near the surface identified on the seismic profiles (Séranne *et al.*, 1995; Bache *et al.*, 2010), and the beginning of the anomalous velocities zone in the next domain. The zonation of thinning areas in the Hinge zone, Domains Ia and Ib, suggests a thinning process in two different stages, compatible with a depth-dependent model of thinning and/or maybe also the influence from Pyrenean inheritance (Bache *et al.*, 2010).

- C) A 130-km-wide domain with a very thin (~5 km) crust of undetermined nature (Domain II: 140 – 270 km). This domain presents a very smooth basement with a high amplitude reflector, a well-marked deep reflector at around 8 s TWT, and a relatively high crustal velocity (varying between 6.0 and 6.4 km/s for the upper layer and 6.6 and 7.5 km/s for the lower layer on profile AB, [Figures 8, 9 and 11](#)). Several hypotheses can be considered for the nature of its basement: i) extremely thinned/stretched continental crust (Purdy, 1975; McKenzie, 1978; Pascal *et al.*, 1993), ii) thin oceanic crust formed at very slow

spreading rate (Sawyer *et al.*, 1994; Whitmarsh & Sawyer, 1996; Srivastava *et al.*, 2000), iii) exhumed upper mantle material, partially serpentinized by seawater intruding into the crust through extensive faults (Boillot *et al.*, 1980, 1987, 1989; Beslier *et al.*, 1993), or iv) exhumed lower continental crust (Wernicke, 1985; Séranne *et al.*, 1995; Séranne, 1999; Moulin *et al.* 2005 ; Aslanian *et al.*, 2009, Huisman & Beaumont, 2011).

- D) Domain III differs from the previous transitional Domain II by a decrease of the velocity and velocity gradient in the lower crust (Figure 12C) and the loss of deep seismic reflectors for the basement and the Moho. Although the velocities of the upper crust are not well constrained (see Figure 7), this transition occurs at around 270 – 280 km (Figures 8 and 11), 50 km southeastwards to the limit previously proposed by Gailler *et al.* (2009). This domain, with a relatively thin crust (4 – 5 km), is supposed to correspond to an “atypical” oceanic crust (Le Douaran *et al.*, 1984; Pascal *et al.*, 1993; Mauffret *et al.*, 1995; Chamot-Rooke *et al.*, 1999; Rollet *et al.*, 2002, Gailler *et al.*, 2009). This domain corresponds to a change in the magnetic and gravity pattern (Galdeano & Rossignol, 1977; Sandwell & Smith, 1997; Leroux, 2012). It is worth to note that there is a clear coincidence between the limit of Domain III and the change in the morphology of the salt layer from a tabular province to a connected domes province (Dos Reis *et al.*, 2005).

Specific features: In the south-western extremity of the Gulf of Lion, the profile CD crosses an E-W elongated prominent structure (Figure 1), which was interpreted as a volcanic structure (Maillard & Mauffret, 1999; García *et al.*, 2011). Nevertheless, the profile CD demonstrates two E-W parallel structures of more than 17 km thick crust with seismic velocities typical of continental crust, separated by a 4 km deep, 20 km-wide basin, which seems to preclude the volcanic hypothesis. Situated in-between the Liguro-Provençal Basin and the Valencia Basin, in a buffer zone at the Northwestern end of the North Balearic Transfer Zone (Figure 1), and oblique to all geodynamic movement, these structures may be a consequence of the different histories of these two basins.

509 1D-VZ profiles and crustal nature

510 Domains of Unthinned Continental Crust and Continental Slope

511 The nature of the crust is analyzed through 1D-Velocity/Depth profiles on each
 512 domain every 10 km (Figure 12) in order to be compared to the compiled worldwide
 513 envelope available for continental crust (Christensen & Mooney, 1995) and normal
 514 oceanic crust (White *et al.*, 1992).

515 The 1D-VZ profiles on images of unthinned continental crust underneath land seismic
 516 stations (purple line on Figure 12A) show similarities with the worldwide compilation
 517 of unthinned continental crust profiles, both in velocities and gradients. A very small
 518 velocity step marks the transition between the upper and lower crustal layer (around 16
 519 km depth).

520 The 1D-VZ profiles on the continental slope (Figure 12A) show similar shape but a
 521 variable Moho depth. From land to deep sea, we can describe three areas of thinning: i)
 522 in the Hinge Zone (brown profiles) where the thinning occurred in the lower part; ii) in
 523 Domain Ia (blue profiles), where the thinning occurred mainly in the upper part; and
 524 iii) in Domain Ib (green profiles), where the thinning occurred in both parts of the
 525 crust. In each domain, the decreasing of the thickness is connected with a slight
 526 increase of velocities in both layers.

527

528 Domain II

529 The 1D-VZ profiles in Domain II (Figure 12B) show a sharp change from the previous
 530 domain, with a fairly homogeneous shape and a strong velocity gradient in the lower
 531 layer. These profiles do not fit a typical oceanic crust profile, nor an unthinned
 532 continental crust profile. The upper layer has a high velocity (>6 km/s) and the Moho is
 533 well marked by a step of about 0.5 km/s.

534 In order to test the hypothesis of exhumed mantle in Domain II, Figure 13 presents the
 535 comparison of the 1-D velocity structure of Domain II and of two areas where the
 536 presence of exhumed mantle is inferred, (Newfoundland Grand Banks margins:
 537 Screech line 2 and Southern Iberia Abyssal Plain: IAM9 and CAM) and the Tagus
 538 Abyssal Plain. In the southern Iberian Abyssal Plain (IAM-9 profiles, Figure 13A), a
 539 zone inferred to be exhumed upper mantle has a thickness of 5 km divided into two
 540 layers (Dean *et al.*, 2000). An upper, high-velocity gradient layer, 2 – 4 km thick,

interpreted to be upper mantle serpentinitised peridotite (with a mean bulk serpentinitisation from 25% to 100%) along faults, which may have been subsequently locally intruded by the products of decompression melting in the mantle, and a lower layer up to 4 km thick with a velocity of 7.3 – 7.9 km/s, representing mantle peridotite with a mean bulk serpentinitisation of < 25%, possibly concentrated along fewer, but steep cutting faults. On the IAM9 profile reflections from the Moho, representing the serpentinitisation front, are weak in the wide-angle seismic data and absent in the normal incident data. Comparison with vz-profiles in Domain II of Gulf of Lion highlights a completely different velocity structure: Domain II shows much higher velocities in the upper layer, much lower velocities in the lower layer, very different gradients and the presence of a strong velocity step at the Moho.

On the eastern Grand Banks of Newfoundland, the wide-angle seismic profile (SCREECH 2) shows a ~25 km-wide zone with a smooth basement, characterized by velocities higher than 6.3 km/s, increasing to 7.7 km/s at a depth of 5 km (Figure 13B). This region is interpreted to be exhumed upper mantle material (Van Avendonk *et al.*, 2006). Similar zones are much wider (60 – 160 km) on SCREECH line 3 (Lau *et al.*, 2006) and on the Iberia Abyssal Plain (Dean *et al.*, 2000). Comparison of vz-profiles shows Domain II having a much thicker crust and the presence of a strong velocity step that is not seen on the Screech profiles.

South of the Galician margin, where the serpentinitised upper mantle material was drilled during Legs 173 and 149 of the Ocean Drilling Program (ODP Leg 173 scientific party 1998), the crustal structure was imaged by wide-angle profile (Chian *et al.*, 1999; Minshull *et al.*, 1998). Figure 13C shows the differences between the vz-profiles of the South Galician margin and of domain II: this latter appears with much higher velocities in the upper layer and a strong velocity step at the Moho.

These three areas therefore present huge differences from the velocity structure observed in Domain II: absence of a strong velocity step at the top of the mantle; absence of strong reflectors on the MCS profile at the base of the lower layers; much lower velocities in upper layers.

These evidences clearly demonstrate that domain II cannot be made by exhumed mantle.

At the Tagus Abyssal Plain the ocean-continent transition zone is marked by two

distinct layers with velocities ranging from roughly 6.0 – 6.3 km/s and 6.4 – 7.2 km/s (Figure 13D; Afilhado *et al.*, 2008). Modeling the data from this region required two velocity steps, one at the intra-crustal layer boundary and one at the Moho. Both correspond to changes in the reflectivity observed in the MCS data. Based on these findings the authors argue for a fourth hypothesis, « a transitional domain basement composed of thinned continental crust overlying a heterogeneous layer of intruded gabbros or intruded serpentinised peridotites, that vanished below the undoubtedly continental crust in the continental slope. » (Afilhado *et al.*, 2008). This hypothesis differs completely from the one proposed less than 100 km to the North on the Iberian Margin (Dean *et al.*, 2000). In the comparison of vz-profiles shown in Figure 13D the velocity-depth profiles match very well with the vz-profiles in Domain II of Gulf of Lion, except for differences in thickness of the layers (both layers have a thickness of 2 – 3 km in the Tagus Abyssal Plain while the upper layer in Domain II is significantly thinner than the lower layer). The occurrence of the T-reflector (De Voogd *et al.*, 1991), which seems to separate the top of the high velocity zone of Domain II (Pascal *et al.*, 1993) from the pinch-out of the lower crust of Domain I, and the lack of evidence for crustal extension in Domain II favour exhumation of the lower crust along a landward deep detachment as described previously by Séranne *et al.* (1995) and Séranne (1999). Domain II may likely consist of a thin exhumed lower continental crust overlying a heterogeneous, intruded lower layer.

Domain III

Comparison between the 1D velocity structure of Domain III and the typical oceanic crust (White *et al.* 1992) is shown in Figure 12C. The crust in Domain III is thinner than “normal” oceanic crust (5 km). Thin oceanic crusts can be found in distinct contexts (Edwards *et al.*, 1997): i) at mid-ocean ridges spreading at ultra-slow rates (< 15 mm yr⁻¹ full rate according to Bown & White, 1994); ii) adjacent to non-volcanic rifted margins (Whitmarsh *et al.*, 1990; 1993; Pinheiro *et al.*, 1991; Horsefield *et al.*, 1993; Afilhado *et al.*, 2008); iii) at oceanic transform faults/fracture zones (Sage, 1994; Mascle *et al.*, 1995; Edwards *et al.*, 1997; ...) and some back-arc basins (Louden, 1980, Crawford *et al.*, 2003; Nishizawa *et al.*, 2011). Nevertheless, the upper layer velocity of Domain III, though not well-resolved, is higher than those in layer 2 of “normal”

oceanic crust formed by pillow lavas with typical velocities around 3.5 – 4 km/s. This result contradicts to the results obtained by Gailler *et al.*, 2009 with the tomographic, unfocused analysis. As a matter of fact, the 1D velocity profiles of Domain III seem to be shifted in depth by 2 km, just as if layer 2 was missing. Figure 14 compares these profiles to the one acquired on the Gabbroic Atlantis Platform in the South-west Indian Ocean (Muller *et al.*, 1997).

On the other hand, the comparison between 1D-VZ profiles in Domains II and III (Figure 12C) exhibits very similar velocities and gradients to the upper layer and the same depth for the Moho discontinuity, which may suggest a relationship for the genesis of these two domains: the only difference concerns the lower layer, which has a nearly constant velocity in Domain III (about 6.5 – 6.7 km/s) while Domain II shows a strong gradient (6.5 – 7.5 km/s). In term of magnetic signature, in the Gulf of Lion, the coincident presence of a magnetic anomalies domain suggests a more typical oceanic nature in domain III. Nevertheless, the existence of wide marginal magnetic anomalies, like the East Coast Magnetic Anomaly (Vogt *et al.*, 1973) or the West Coast Magnetic Anomaly (Sahabi *et al.*, 2004), in the Central Atlantic Ocean on the transitional domain before any true oceanic crust, may complicate this view. Following Bott (1971) and Aslanian *et al.*, 2009, Sibuet & Tucholke (2012) have suggested that in the case of the Brazilian margins, the lower crust may have flown and fed a first atypical oceanic crust.

Lastly, the boundary between these domains II and III coincides with a clear change in the morphology of the salt layer together with the magnetic signature. The coincidence between atypical oceanic crust domain and the area where the salt presents a connected diapirs morphology is striking. This relationship between salt morphology segmentation and crustal domains with different temperature gradient was already observed on the Angolan margin (Moulin *et al.*, 2005; Aslanian *et al.*, 2009) and can be observed on the Algerian Margin (Medaouri *et al.*, 2014) and on the Gulf of Mexico where two different crusts are present. This connexion between salt behaviour and temperature (which will be different according to the heat flow) is also shown in laboratory experiments (R. Ziane, 2014).

9 Conclusions

Modelling of combined wide-angle and near-vertical reflection seismic (MCS) data from three profiles along the Liguro-Provençal basin provides images into the deep structure of the margin, from unthinned continental crust to the deep basin. On the basis of our study, we can divide the margin into four domains: i) the unthinned crust; ii) the « continental slope domain »; iii) the transitional domain and iv) the atypical magnetic oceanic crust. The image obtained in the continental slope domain confirms the results of Bache *et al.*, 2010.

In the transitional region (Domain II, 140 – 270 km), we refine the results of Pascal *et al.*, 1993, the presence of a velocity step at the top of the mantle, strong reflectors on the MCS profile at the base of the lower layer, and the general shape of the 1-D VZ profiles preclude the hypothesis of exhumed mantle. The co-occurrence of the T-reflector, which seems to separate the top of the anomalous velocity zone of Domain II from the pinch-out of the lower crust of Domain I, and the lack of evidence of crustal distention in Domain II play in favor of a thin layer of lower continental crust, exhumed along a landward deep detachment and overlying a heterogeneous, intruded layer (gabbros or granulites).

Our results demonstrate also that the region of “atypical”, 4 – 5 km thick, magnetic oceanic crust (Domain III), exhibits very similar velocities and gradients than Domain II, which may suggest a relationship for the genesis of these two domains. This domain coincides with a clear change in the morphology of the salt layer and the magnetic signature.

Finally, the refraction data confirm and precise the overall geometry obtained by tomography inversion: a clear symmetrical distribution of the velocity structures in the two conjugate margins (Gailler *et al.*, 2009), but with some differences (see Afilhado *et al.*, this issue). This symmetrical distribution seems to exclude the hypothesis of simple shear mechanism along a lithospheric detachment fault.

Last, wide-angle modeling give also a supplementary information about the E-W elongated prominent structures on the south–western part of the Gulf of Lion, which exhibits very typical continental seismic velocities, which seems to preclude a pure volcanic origin.

668

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690 **Author contributions**

691 The Sardinia Project was imagined by J.-L. Olivet and led by D. Aslanian from Ifremer.
 692 Modelling of the Sardinia-GOL profiles was done by M. Moulin, F. Klingelhoefer, A.
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 695 Seismic interpretations of the sedimentary sequences were done in collaboration with
 696 M. Rabineau.

697 **Figures**

698 Figure 1 - A/ Location of the study area of the Gulf of Lion's margin marked by the
 699 black outlined rectangle, in the general context of Western Mediterranean Basin.
 700 Isocontours are taken from the bathymetric map of the Provencal Basin and Tyrrhenian
 701 Sea (in meters) (modified from Berné *et al.*, 2002) contoured every 200 meters. The
 702 Provencal Basin was created by counter-clockwise rotation of the Corsica-Sardinia
 703 micro-plate during the Miocene (see flowlines). The crustal Domains I to V are labeled;
 704 the « atypical » oceanic crust lying in the center of the basin (Domain III) is hatched in
 705 thin parallel black lines. The red lines indicate the location of all SARDINIA seismic
 706 profiles. GL-Gulf of Lion, PM-Provence Margin, CM-Catalan Margin, WS-West
 707 Sardinia, LS-Ligurian Sea, VT-Valencia Trough, LPB-Liguro Provencal Basin.

708 B/ Topographic and bathymetric map of the study area (in meters) (modified
 709 from Berné *et al.*, 2002 & Loubrieu *et al.*, 2008) with isocontours every 100 m and the
 710 location of the seismic profiles (black lines). Ocean bottom seismometer locations are
 711 represented by red triangles, land seismic stations by red circles. The yellow ones
 712 correspond to instruments with problems, unusable for processing and modelling.
 713 Three kinds of OBS were used during the SARDINIA experiment, they are identified
 714 by a simple red triangle (micrObs from Ifremer pool), a red triangle with thick black
 715 lines (OldOBS from Ifremer pool) and a clear/transparent red triangle (OBS/H from
 716 Geomar pool). Thin black lines represent the multi-channel seismic profile of the
 717 ECORS Experiment (De Voogd *et al.*, 1991). Blue stars mark the position of the ESP
 718 (Expanding Spread profiles) of the CROC-2 cruise (Le Douaran *et al.*, 1984; Pascal *et*
 719 *al.*, 1993; Contrucci *et al.*, 2001).

720 Thick grey lines delimit the main crustal domains of the margin, based on seismic and

magnetic analysis (Bayer *et al.*, 1973; Galdéano & Rossignol, 1977; Rollet *et al.*, 2002; Bache *et al.*, 2010). The position of the North Balearic fracture zone is based on Maillard & Mauffret (1999).

Figure 2 – Velocity models for all three SARDINIA-Gulf of Lion profiles including the model boundaries used during inversion (blue solid lines) and isovelocity contours every 0.20 km/s. OBS locations are indicated by black triangles and land seismic stations by grey triangles. The various domains are separated by thin black lines. V.E. = ~4. A/ Profile AB, B/ Profile CD, and C/ Profile EF.

Figure 3 – A/ Upper panel: Ray coverage of the sedimentary layers of profile AB with every tenth ray from two-point ray-tracing plotting. Lower Panel: Observed travel-time picks and calculated travel-times (lines) of the sedimentary layers for all receivers along the model. B/ Same as A/ but for the crustal layers, C/ Same as A/ but for the Moho and upper mantle layers.

Figure 4 – A/ Upper panel: Ray coverage of the sedimentary layers of profile CD with every tenth ray from two-point ray-tracing plotting. Lower Panel: Observed travel-time picks and calculated travel-times (line) of the sedimentary layers for all receivers along the model. B/ Same as A/ but for the crustal layers, C/ Same as A/ but for the Moho and upper mantle layers.

Figure 5 – A/ Upper panel: Ray coverage of the sedimentary layers of profile EF with every tenth ray from two-point ray-tracing plotting. Lower Panel: Observed travel-time picks and calculated travel-times (line) of the sedimentary layers for all receivers along the model. B/ Same as A/ but for the crustal layers, C/ Same as A/ but for the Moho and upper mantle layers.

Figure 6 – A/ Ray Coverage of land seismic station GDL05 on profile AB; B/ Observed travel-time picks and calculated travel times of land seismic station GDL05 on profile AB; C/ Ray Coverage of OBS43 on profile AB; D/ Observed travel-time picks and calculated travel times of OBS43 on profile AB; E/ Ray Coverage of OBS08 on profile

AB; F/ Observed travel-time picks and calculated travel times of OBS08 on profile AB; G/ Ray Coverage of OBS56 on profile AB; H/ Observed travel-time picks and calculated travel times of OBS56 on profile AB; I/ Ray Coverage of OBS20 on profile CD; J/ Observed travel-time picks and calculated travel times of OBS20 on profile CD; K/ Ray Coverage of OBS29 on profile CD; L/ Observed travel-time picks and calculated travel times of OBS29 on profile CD; M/ Ray Coverage of OBS10 on profile EF; N/ Observed travel-time picks and calculated travel times of OBS-10 on profile EF; O/ Ray Coverage of OBS05 on profile EF; P/ Observed travel-time picks and calculated travel times of OBS05 on profile EF.

Figure 7 - Resolution of all velocity nodes of the final velocity models. White and yellow areas can be considered well resolved. OBS positions are indicated by black triangles, land seismic stations by grey triangles. A/ Profile AB, B/ Profile CD, and C/ Profile EF.

Figure 8 - Reflection seismic sections for the all three profiles. Model boundaries from wide-angle modelling converted to two-way travel-time are overlain. OBS positions are indicated by black triangles. A/ Profile AB; B/ Profile CD and C/ Profile EF.

Figure 9 - Pre-stack depth migration of the profile AB, with the wide-angle reflection/refraction velocity model. Model boundaries and crustal isovelocities (at 6, 6.75, 7, and 7.25 km/s) are overlain, as well as the interpretative sketch of this profile, based on our seismic velocity model and previous seismic interpretations.

Figure 10 - A/ Data from OBS 26 (OBS at the intersection from profile AB and CD) on profile CD, with the same gain, filter and scaling applied as in Figure 2. B/ and C/ Corresponding synthetic seismograms for the two different CD models: D/ Final model used in this study with the lower crustal layer presenting a higher gradient (6.6 - 7.2 km/s); E/ First model obtained with very lower gradient in the lower crust (6.8 - 7 km/s).

Figure 11 - Velocity models for all three SARDINIA-GOL profiles including the model

boundaries used during inversion (blue solid lines) and isovelocity contours every 0.20 km/s, OBS locations are indicated by black triangles and land seismic stations by inverted grey triangles. Areas unconstrained by ray-tracing modelling are uncolored. The various domains are separated by thin black lines. V.E.= ~4. A/ Profile AB, B/ Profile CD, and C/ Profile EF.

Figure 12 - P-wave Velocity depth profiles below the basement in the three distinct domains defined in this study on the profile AB. Each domain is compared to the compilations made for continental crust (Christensen & Mooney, 1995) (grey lines) and a « normal oceanic crust » (White et al., 1992) (blue shaded envelope). Velocity-depth profiles are extracted each 10 km (except in the HZ, each 5km) in A) the unthinned (in purple) and the continental slope domains (in brown (HZ), blue (Ia) and green (Ib); B), the transitional domain (in orange) and C) the atypical oceanic crust domain (in red). D) The exact position of the extracted Vz-profiles is illustrated on profile AB with the same colors.

Figure 13 - Comparison of 1D velocity versus depth below basement (vz) profiles in Domain II of Sardinia-AB (130 - 270 km) between the bounds of vz-profiles (blue shaded areas) from wide-angle seismic models where ocean-continent transition zones was interpreted to be exhumed and/or serpentinised upper mantle: (A) IAM-9 in Iberia Abyssal Plain (Dean et al., 2000); (B) SCREECH line 2 in the Grand Banks margin offshore Newfoundland (Van Avendonk et al., 2006); (C) CAM in the Southern Galician margin (Chian et al., 1999); and (D) IAM-5 wide-angle seismic model acquired at the Tagus Abyssal Plain (Afilhado et al., 2008). Velocity-depth profiles are extracted every 10 km on all profiles.

Figure 14 - Comparison of 1D velocity versus depth below basement (vz) profiles in Domain III and vz-profiles acquired on the gabbroic massif of the Atlantis Bank (SW Indian Ridge). A/ Bathymetric map of the Atlantis Bank, on the SW Indian Ridge where a massif of Gabbro in situ of an oceanic ridge was discovered Depths greater than 5500 m and less than 1300 m are shaded, the latter correspond to the area referred to as the Atlantis Platform where the site ODP 735B was drilled. The circles shows locations of ROV and submersible dives, dredge hauls, existing ODP and IODP

drill sites and their compositions (in green are indicated the lithology relative to gabbro). Red triangles mark ocean bottom hydrophones acquired on line CAM 101, one of the two refraction profiles acquired in this area (Muller *et al.*, 1997; White, com. Pers.). B/ Vertical profiles through the Pwave seismic velocity model of CAM101 (Muller *et al.*, 1997) at positions A, B, and C. Grey area: compilation of velocity profiles for the Atlantic “normal” 0 – 7 Ma aged oceanic crust (White *et al.*, 1992). Profile B (in green) is located on crustal material, consisting of a 4.50 m thick, massive olivine gabbro layer, topped by a small unit of gabbronorite and intruded by troctolite and troctolite gabbro at its base. The red profiles correspond to 1D velocity versus depth below basement (v_z) profiles in Domain III of Sardinia-AB (270 – 400 km) for comparison. Velocity-depth profiles are extracted every 10 km on all profiles. C/ P-wave seismic velocity model on lines CAM101 (Muller *et al.*, 1997) The velocity contour interval is 0.3 km/s. Numbered OBH positions are shown on the seafloor by red triangles. The position of ODP hole 735B has been projected (thick black line). The Moho is indicated as a thicker line where its depth is constrained by wide-angle reflections. A, B and C show the locations of the v_z -profiles presented in part B (see above).

Figure Annex-1 – A/ Bandpass filtered (5–20 Hz) seismic record section data from OBS 26 on profile AB. The seismic data are gain-adjusted and time reduced to 6 km/s according to the offset. B/ Synthetic seismic record section calculated from the model for the same station using the asymptotic approach of ray theory from the Zelt code (Zelt & Ellis, 1988). The synthetic seismograms are calculated every 500 m with a sampling rate of 120 Hz, obtained by convolution of the impulse response with a 29-point low passed Ricker wavelet. Time and amplitude corrections according to offset are the same as in A/.

Figure Annex-2 – A/ Data from OBS 26 on profile CD, with the same gain, filter, and scaling applied as in Fig. 2. B/ Corresponding synthetic seismograms.

Figure Annex-3 – A/ Data from OBS 8 on profile AB, with the same gain, filter, and scaling applied as in Fig. 2. B/ Corresponding synthetic seismograms.

850

851 Figure Annex-4 - A/ Data from OBS 8 on profile EF, with the same gain, filter, and
852 scaling applied as in Fig. 2. B/ Corresponding synthetic seismograms.

853

854 Figure Annex-5 - A/ Data from OBS 39 (located on continental domain) on profile
855 AB, with the same gain, filter, and scaling applied as in Fig. 2. B/ Corresponding
856 synthetic seismograms. C/ Data from OBS56 (located on oceanic domain) on profile
857 AB, with the same gain, filter, and scaling applied as in Fig. 2. D/ Corresponding
858 synthetic seismograms.

859

860 Figure Annex-6 - A/ Data from OBS 28 on profile CD, with the same gain, filter, and
861 scaling applied as in Fig. 2, but time-reduced to a velocity of 8 km/s, showing two
862 distinct reflective phases in the transitional crust. B/ Corresponding synthetic
863 seismograms.

864

865 Figure Annex-7 - A/ Data from OBS 25 on profile CD, with the same gain, filter, and
866 scaling applied as in Fig. 2, but time-reduced to a velocity of 8 km/s, showing a well-
867 defined Pn. B/ Corresponding synthetic seismograms.

868

869 Table 1 - Travel-time residuals and chi-squared error for all phases and the complete
870 model of profile AB.

871

872 Table 2 - Travel-time residuals and chi-squared error for all phases and the complete
873 model of profile CD.

874

875 Table 3 - Travel-time residuals and chi-squared error for all phases and the complete
876 model of profile EF.

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Figure 1 - Sardinia GDL

Figure 2 - Sardinia GDL

Figure 3 - Sardinia GDL

Figure 4 - Sardinia GDL

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Figure 7 - Sardinia GDL

Figure 9 - Sardinia GDL

Figure 10 - Sardinia GDL

Figure 11 - Sardinia GDL

Figure 12 - Sardinia GDL

Figure 13 - Sardinia GDL

Figure Annex-1 - Sardinia GDL

Figure Annex-2 - Sardinia GDL

Figure Annex-3 - Sardinia GDL

Figure Annex-4 - Sardinia GDL

Figure Annex-5 - Sardinia GDL

Figure Annex-6 - Sardinia GDL

Figure Annex-7 - Sardinia GDL