

# Crustal structure and continent-ocean boundary along the Galicia continental margin (NW Iberia): insights from combined gravity and seismic interpretation

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## Key Points:

- The diverse rift-related structural domains in the margin have distinctive gravity anomaly imprints
- Mapping of the crustal and structural domains shows a strong segmentation of the NW Iberia margin with high inheritance control
- Rift architecture and segmentation control the development of the partial tectonic inversion of the margin

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## Abstract

The magma-poor rifted continental margin of Galicia has an extremely complex structure. Its formation involved several rifting episodes that occurred ultimately during the early Cretaceous near a ridge triple junction, which produced a change in the orientation of the main structures in its transition to the north Iberia margin. In addition, there is a superimposed partial tectonic inversion along its northwest and northern border which developed from the Late Cretaceous to at least Oligocene times. The present study integrates a large volume of new geophysical information (mainly marine gravity data and 2D seismic reflection profiles) to provide insights on the formation of this rift system and on the development of its later inversion. The combined interpretation and modeling of this data enable the presentation of a new crustal and structural domains map for the whole Galicia margin. This includes the rift domains related to the extreme thinning of the crust and the lithospheric mantle (stretched, necking, and hyperextension and mantle exhumation (HME) domains), as well as a domain of intense compressional deformation. New constraints arise on the origin, the deep structure, and the characterization of the along- and across-strike variation of the continent-ocean transition of the margin, where a progressive change from hyperextension to partial inversion is observed. The development of both rifting and later partial tectonic inversion is influenced by the existence of former first-order tectonic features. Most of the tectonic inversion is focused on the HME domain, which in some areas of the northwestern margin is completely overprinted by compressional deformation.

## 1 Introduction

The understanding of the development of magma-poor rifted margins and their behavior under a subsequent compressional tectonic regime has been in the spotlight during the last years. Recent models are based on the study of different domains, typically seaward arranged, that originated during the rifting. A good deal of these studies have been developed along the Iberia margins and its conjugates [e.g., *Lavier and Manatschal, 2006; Péron-Pinvidic et al., 2013; Tugend et al., 2014; Stanton et al., 2016*]. This rift domain mapping allows the interpretation of distinct stages of their formation, illustrating the margin evolution as well as the behavior of each domain during tectonic reactivation [e.g., *Tugend et al., 2014*]. New concepts and definitions have been developed, but there is still a relative variability about the nomenclature and delineation of boundaries between domains [*Péron-Pinvidic et al., 2013; Sutra et al., 2013; Tugend et al., 2015b; Chenin et al., 2017*].

Although some detailed rift-related domains mapping has been previously interpreted on the west Iberia margin [*Péron-Pinvidic et al., 2013*], and on the north Iberia margin [*Tugend et al., 2014*], there is a gap on this domain mapping on the assembly between both margins, i.e., on the (northwest) Galicia margin. The aim of this work is to analyze the structural framework of the whole Galicia margin, from west to north. A detailed analysis of gravity data and 2D reflection seismic profiles, accompanied by 2+3/4D gravity models serially arranged across the main features of the margin, provide information on its the deep crustal structure. The newly mapped rift domains and the observations made on the deep structure of the margin throw light on the way the transition from the west to the north Iberia margins takes place, as well as on the influence of the rift architecture on its later partial inversion.

### 1.1 Terminology

All magma-poor rifted margins have some common architectural features [e.g., *Reston , 2009*], with an outstanding succession of different domains that relate to specific deformation processes [e.g., *Sutra et al., 2013; Péron-Pinvidic et al., 2013; Tugend et al., 2014; Naliboff*

*et al.*, 2017]. The boundaries between these rift domains are sometimes diffuse, as deformation processes may overlap in time and space [e.g., *Tugend et al.*, 2015b]. There is a relatively wide range of rift domains definitions [e.g., *Sutra et al.*, 2013; *Péron-Pinvidic et al.*, 2013; *Tugend et al.*, 2014; *Chenin et al.*, 2015; *Stanton et al.*, 2016] and the criteria for the identification of their limits may differ. The terminology used in this study is adapted from those previously published [e.g., *Sutra et al.*, 2013; *Péron-Pinvidic et al.*, 2013; *Tugend et al.*, 2014; *Chenin et al.*, 2015], in order to account for the special characteristics of the Galicia margin, which are described further down. The terminology adopted here is:

#### *Stretched domain*

This domain corresponds to the *proximal domain* described by *Péron-Pinvidic et al.* [2013] and to the *stretched domain* defined by *Sutra et al.* [2013]. This region is moderately affected by crustal thinning [*Péron-Pinvidic and Manatschal*, 2009], and basement top and the Moho are approximately parallel. Deformation is decoupled, with brittle faulting at the upper part of the crust and a ductile behavior of the lower continental levels, separated by a brittle/ductile transition [decoupling level, *Sutra et al.*, 2013]. Due to the moderate thinning, little accommodation space is created [*Sutra et al.*, 2013].

#### *Necking domain*

In this region, the crust is wedge-shaped and records a thinning until reaching 10 km of crustal thickness [*Péron-Pinvidic and Manatschal*, 2009; *Mohn et al.*, 2010]. Moho inflects and rises, tending to converge with the dipping continental basement top, and much more accommodation space is created than in the stretched domain [*Péron-Pinvidic and Manatschal*, 2009; *Péron-Pinvidic et al.*, 2013]. Here, crustal deformation is still decoupled [*Sutra et al.*, 2013]. The distal border of this domain is located where the crustal thickness is less than 10 km thick, and an embrittlement of the whole crust occurs [*Pérez-Gussinyé and Reston*, 2001; *Pérez Gussinyé et al.*, 2003; *Péron-Pinvidic et al.*, 2013; *Sutra et al.*, 2013].

#### *Hyperextension and mantle exhumation (HME) domain*

Along this distal domain, crustal thickness is less than 10 km and the tectonic deformation is coupled. Extensional faulting penetrates and exhumes the continental lithospheric mantle under a simple shear regime [*Pérez-Gussinyé et al.*, 2003; *Doré and Lundin*, 2013; *Sutra et al.*, 2013], and serpentinization of the lithospheric mantle occurs [*Pérez-Gussinyé and Reston*, 2001]. As tectonic thinning progresses, the serpentinized mantle is exposed, and is in direct contact with the sedimentary cover or may crop out to the seafloor [e.g., *Boillot et al.*, 1988a, 1988b, *Sutra et al.*, 2013 and references therein].

## **2 Regional setting**

The Galicia margin, as the rest of the west and north Iberia margin, originated from the northwards propagation of the Atlantic Ocean opening from Late Jurassic to Early Cretaceous times [e.g., *Murillas et al.*, 1990]. This rift episode affecting Iberia culminated with the breakup of the conjugate margin pairs of Galicia – Flemish Cap (the northern sectors of the west Iberia – Newfoundland conjugate margins) [e.g. *Sutra et al.*, 2013 and references therein] and the opening of the Bay of Biscay, with the separation of the north Iberia – Armorica-Celtic conjugate margins [e.g., *Tugend et al.*, 2014 and references therein]. These conjugate margin pairs of west and north Iberia were connected by a triple ridge junction located to the northwest of Galicia. The kinematics of this triple junction initiation is strongly debated and different models are proposed for the Bay of Biscay opening, ranging from a transtensional model, to a scissor-type opening [e.g., *Sibuet and Collete*, 1991; *Olivet*, 1996; *Srivastava et al.*, 2000; *Sibuet et al.*, 2004b; *Jammes et al.*, 2009; *Mouthereau et al.*, 2014].

Along-strike segmentation on the west Iberia margin has been related to the staged progression of the North Atlantic opening [e.g., *Malod and Mauffret*, 1990, *Srivastava et al.*, 1990; *Alves et al.*, 2009, *Brune et al.*, 2014]. This segmentation of the western margin is favored by the existence of NW-SE and NE-SW fractures, that have been considered either as inherited from the Late Hercynian fabric [e.g., *Sibuet et al.*, 1987; *Boillot and Malod*, 1988; *Murillas et al.*, 1990; *Martínez Catalán et al.*, 2009; *Manatschal et al.* 2015] or originated during the Late Triassic-Early Jurassic rift event [e.g., *Vegas et al.*, 2016]. On the Galicia sector, the succession of time-separated rifting pulses, lead also to a westward jump of the rift axis, due to lithospheric rheological changes [*Manatschal and Bernoulli*, 1999]. North of Iberia, recent studies point to a rift partition of the margin in different left-lateral transtensional rift systems, where NNE-SSW transfer faults (e.g., Pamplona transfer zone and Santander soft transfer zone), probably originated during Triassic times [e.g., *Ziegler*, 1982; *Sopeña et al.*, 1988; *De Vicente and Vegas*, 2008] play a major part during rifting [*Jammes et al.*, 2009; *Roca et al.*, 2011; *Tugend et al.*, 2015a]. The oceanic crust accretion continued in the Bay of Biscay until Santonian times [*Srivastava et al.*, 1990]. Afterwards, the roughly N-S compressional tectonic stress field taking place during the Alpine orogeny (since the Late Santonian to Oligocene-Miocene times) [e.g., *Thinon et al.*, 2001; *Capote et al.*, 2002; *Tugend et al.*, 2015a] lead to the reactivation and partial inversion of previous rift structures, and some new compressional structures were also generated, such as thrusts, reverse faults, folds and flexures [e.g., *Grimaud et al.*, 1982; *Murillas et al.*, 1990; *Vázquez et al.*, 2008; *Tugend et al.*, 2014].

As a result of this complex tectonic history, the present-day Galicia continental margin shows an intricate structure. Thereby, the upper continental slope imaged on the Galicia margin, comprises different geomorphologic provinces, which are (Fig. 1a): The marginal platforms region northwest of Galicia (MPR), the Galicia Interior Basin (GIB), the seamounts region (SR), and the deep Galicia margin (DGM). The lower continental slope is steep at the northwestern flank of the SR [*Ercilla et al.* 2008], as well as bordering the MPR and to the north Iberia margin. The GIB is located westwards of the Galician continental shelf, trending NNW-SSE, with an approximated width of 100 km. In this basin, great NNW-SSE normal faults stand out, delineating narrow and elongated continental basement blocks, mainly tilted during the Berriasian-Valanginian rifting stage [*Murillas et al.*, 1990] (see Fig. 1d), and offset by NE-SW and NW-SE faults. The sedimentary infill reaches a thickness of 6 km in the center of the basin, thinning towards the continental shelf and the SR [*Murillas et al.*, 1990; *Pérez-Gussinyé et al.*, 2003] (See Fig. 1d). Westwards of the GIB, the SR is constituted by a series of continental crust blocks, tilted during the Early Cretaceous rifting and raised later as a consequence of the Pyrenean (Campanian to Oligocene) compressional tectonics [e.g., *Boillot and Malod*, 1988]. This region limits towards the west on the DGM, where it has been identified a band up to few tens of kilometers width, constituted by extremely extended continental crust blocks with a thin sedimentary cover, and underlined by an ultramafic basement (see Fig. 1c); this region ends westwards in a serpentinized peridotite ridge [*Boillot et al.*, 1987]. This ultramafic basement is originated by sub-continental mantle exposure during the final rifting [*Krawczyk et al.*, 1996; *Kornprobst and Chazot*, 2016]. Oceanwards from the DGM, a transition from a thin oceanic crust, underlined by serpentinized mantle, to the normal oceanic crust of the Iberia abyssal plain (IAP) is observed [e.g., *Sibuet et al.*, 1995; *Davy et al.*, 2016]. To the northwest of Galicia, the MPR seems to be constituted by graben basins originated during the Mesozoic rifting (probably prior to Late Aptian times), as observed in other basins on the north Iberia margin and in the conjugate Armorican margin [*Thinon et al.*, 2003; *Tugend et al.*, 2015a], and tectonically inverted under the subsequent Santonian to Oligocene compressional regime [*Murillas et al.*, 1990; *Thinon et al.*, 2001; *Tugend et al.*, 2015a]. Oceanwards from the MPR and the north of Galicia, an abrupt

transition to the oceanic crust is noticed [Álvarez-Marrón *et al.*, 1997; Fernández-Viejo *et al.*, 1998, Tugend *et al.*, 2014].

### 3 Materials and methods

Data used in this work come from diverse sources. Offshore, gravity dataset was obtained during seven surveys, carried out between 2001 and 2009 northwest of Iberia, onboard the R/V Hésperides, within the frame of the Oceanographic and Hydrographic Research Project of the Spanish Economic Exclusive Zone. During these cruises, more than 400,000 ship gravity data points were obtained (Fig. 2). Where ship gravity information is not available, the satellite altimetry-derived Global Gravity Model data (version 23.1) have been used [Sandwell *et al.*, 2014]. On land, gravity data have been taken from the databases of the 'Instituto Geográfico Nacional' (IGN), the 'Empresa Nacional de Residuos Radiactivos' (ENRESA) and the 'Bureau Gravimétrique International'. The gravity-meter employed on the data acquisition during the sea surveys is a marine gravimeter Bell Aerospace BGM-3. These data were tied to the on land gravity network on the first order gravity bases of the IGN in Cartagena, Santander, Vigo and La Coruña, using a Lacoste & Romberg (model G) gravimeter. The reduction has been executed with the IGRS67 and the final Bouguer anomaly information includes sea-bottom correction [Kane, 1962; Nagy, 1966; Carbó *et al.*, 2003]. Bouguer correction is calculated for a reduction density of  $2.67 \text{ g/cm}^3$  and the topographic correction has been carried out up to 22 km, using a 2 km-gridded digital terrain model [Sandwell *et al.*, 2014]. Finally, ship gravity data, on land information and that derived from satellite-altimetry, were jointly processed and included in a geo-referenced database. This data set was interpolated to obtain a 2 minute regular grid for the Bouguer Anomaly values. In addition to the gravity data, several original multichannel reflection seismic profiles across northwestern flank of the SR were available. These were collected in the frame of the ERGAP Project, during two surveys carried out on 2007 on board the R/V L'Atalante (IFREMER). A deployment of five mini-GI airguns (Sercel®) and a three-section streamer (each one with 24 channels and 150 m-long) was used during the acquisition. The location of these 2D seismic profiles is shown on Fig. 3.

#### 3.1 Coherency analysis of gravity data

As the final Bouguer anomaly grid include both ship and satellite altimetry-derived gravity data, we have accomplished a spectral analysis of the coherence between both datasets. The aim of this analysis is to evaluate the precision and the presence of a cut-off frequency marking the accuracy lower limit of the dataset derived from satellite altimetry.

For this analysis, we sampled the satellite dataset along a line coincident with a long ship track, traversing the west Galicia margin (Fig. 4a. See location on Fig. 2). The power spectrum was calculated for both datasets profiles (Fig. 4c), following the procedure developed by Wessel and Smith [1995]. The power values are similar for both ship and satellite profiles (Fig. 4c). The coherence between both profiles has values close to 1 for the long wavelength, meanwhile for wavelengths of less than 20 km the coherence is much lower (Fig. 4c and d). According to these results, in areas where only satellite altimetry-derived data are available, we have avoided to model blocks related to anomalies of less than 20 km wavelength.

#### 3.2 Spectral analysis of gravity data

In order to accomplish a complete gravity study, a 2D spectral analysis of the Bouguer anomaly grid (Fig. 5a) has been made, with the aim to identify linear trends characterized by

different slopes that link to distinct wavelength groups [Karner and Watts, 1983], each one related to different kind of lithospheric sources (Fig. 6a). Afterwards, frequency filtering was applied to obtain the following grids: 1) long-wavelength Bouguer anomaly grid ( $\lambda > 150$  km), generally related to deep sources such as variations on the Moho topography (Fig. 6b); 2) mid-wavelength Bouguer anomaly grid ( $150 \text{ km} > \lambda > 33 \text{ km}$ ), that match with gravity anomaly sources located on lower and middle crust levels (Fig. 6c); 3) short-wavelength Bouguer Anomaly grid ( $33 \text{ km} > \lambda > 8 \text{ km}$ ), related to structures located on the upper crust, such as small graben and other relatively superficial anomalous bodies (Fig. 6d).

### 3.3 Gravity modeling

Finally, eight serially arranged 2+3/4D density models were performed all along the Galicia margin (using Geosoft GM-SYS software), in order to analyze its across and along strike structure (Fig. 3). Three of the density models are at least 360 km long, crossing the whole margin: Two are W-E oriented (GM1 and GM2), crossing from the IAP to the western Galicia continental shelf; the third one (GM6) transects the margin N-S from the BAP, through the SR east of Galicia bank, to the IAP south of the SR and the GIB. The rest of the gravity models are shorter and serially arranged across the transition from the continental to the oceanic domain. The location of gravity models is usually conditioned by the existence of previous seismic profiles to constrain them. In our study area, most of the seismic information has been focused on the characterization of the west Iberia margin and its rift structure. As a result, the majority of the previous seismic profiles are E-W directed and located on the southern part of our interest area, and there is a lack of seismic profiling literature northwest of Galicia. Most of the gravity models constructed during this study, are supported by the original seismic sections available for this study, and also by previous reflection and refraction profiling carried out by others [Murillas *et al.*, 1990; Sibuet *et al.*, 1995; Whitmarsh *et al.*, 1996; Reston *et al.*, 1996; Fernández-Viejo *et al.*, 1998; González *et al.*, 1999; Pérez-Gussinyé *et al.*, 2003; Clark *et al.*, 2007; Ercilla *et al.*, 2008]. To set Moho depth, where no refraction seismic information is available, we have used the Moho depth model constructed by Díaz and Gallart [2009], and that derived from the GOCE satellite data (GEMMA Project) [Reguzzoni and Sampietro, 2015]. It must be taken into account that the GEMMA model is a low resolution global model that shows correlation errors in continental margins [Reguzzoni and Sampietro, 2015]. These misfits are due to a simplistic description of the oceanic seafloor used for the model and, locally, to a lack of seismic observations [Sampietro *et al.*, 2013]. Together with the approximate Moho depth [Reguzzoni and Sampietro, 2015; Díaz and Gallart, 2009], where no coincident seismic profiles are available, we use adjacent areas information by correlation of single gravity anomalies to construct models that make geological sense. This procedure has allowed modeling first order structures that imply an important lateral density contrast, such as those of the continent-ocean transition. Modeling the first order tectonic features along the whole Galicia continental margin, complement the gravity maps analysis. This allow moving forward on the comprehension of the margin structure and of the transition from the west Galicia rifted margin, to the north Iberia inverted one.

Calculations of the gravity model response are based on the classic methods and algorithms [Talwani *et al.*, 1959; Talwani and Heirtzler, 1964; Won and Bevis, 1987]. The correlation between densities and seismic waves velocities is made with the aid of the empiric relationship curves between these parameters [Ludwig *et al.*, 1970; Barton, 1986; Christensen and Mooney, 1995; Brocher, 2005]. Ten different density units are distinguished, from the refraction seismic velocities given by Córdoba *et al.* [1987], Sibuet *et al.* [1995], Whitmarsh *et al.* [1996] and Pérez-Gussinyé *et al.* [2003] (Table 1). The transverse extension of the

modeled blocks is prolonged perpendicularly 100 km towards each side in the case of the crustal blocks, and 50 km for the sedimentary cover blocks.

## 4 Results and interpretation

### 4.1 Bouguer anomaly maps

On the study area, Bouguer anomaly values vary from -105 mGal (onshore) to 385 mGal (offshore, IAP) (Fig. 5a). Offshore, in general terms, there is a progressive increase of the Bouguer anomaly values to the west and northwards, *i.e.*, to the oceanic crust regions. On the continental basement region, Bouguer anomaly values are usually below 220 mGal, while on the oceanic regions they are commonly over 300 mGal.

West of Galicia, the SR is characterized by a striking relative minimum (*m1*, Fig. 5a), roughly elongated NNW-SSE and shifted towards the southeast from the highest elevation area (Fig. 5c. See also Fig. 3). It is related to the greater continental crust thickness found in the SR [e.g., Pérez-Gussinyé *et al.*, 2003]. Between this relative minimum and the 100 mGal contour to the east, which approximately matches the continental shelf break, there is a relative maximum located on the GIB (Fig. 3). This maximum axis has three segments, elongated NNW-SSE (*M1* and *M2*, separated by the gradients *G7* and *G8*, on Fig. 5a, see profile c) and NE-SW (*M3*, Fig. 5a), according to the main structural directions. These maximums on the GIB can be attributed to the intense stretching of the continental crust during the first rifting stages [e.g., Murillas *et al.*, 1990; Manatschal and Bernoulli, 1999]. On the west margin, there are large Bouguer anomaly gradients on the transition from the continental domain (*i.e.*, the SR) to the oceanic domain (Fig. 5a): 1) West of the SR, there is an steep gradient of 3.12-3.6 mGal/km (*G1* and *G2* on Fig. 5a, see profile c) elongated N-S to NNW-SSE; 2) From the northwestern flank of the SR to the BAP, there is a nearly E-W directed gradient stripe, up to 2.8 mGal/km (*G3* on Fig. 5a, see profile b); 3) South of the SR, a 2.9 mGal/km NW-SE directed gradient band connects with the IAP (*G4* on Fig. 5a, see profile b). *M4* (up to 350 mGal) surrounds the SR by the north and northwest, and bifurcates in the proximity of the Finisterre bank (FB); it has to be emphasized that these Bouguer anomaly values are atypical for a topographically raised area.

The transition to the oceanic domain northwest of Galicia is staggered and complex, with a 0.7-1 mGal/km medium gradient (*G5*, Fig. 5a), and several relative minimum and maximum axis (e.g., *m2-m3* and *M7-M8* on Fig. 5a, see profile d). The *m2* and *m3* axis seem to converge on a relative minimum of less than 200 mGal at the foot of the marginal platforms, probably related to the depocenter mapped by Grimaud *et al.* [1982].

North of Galicia, the transition to the oceanic domain is shown as a regular Bouguer anomaly gradient of 1.2 mGal/km (*G6*, Fig. 5a).

#### 4.1.1 Long-wavelength Bouguer anomaly map

This map, related to deep anomaly sources such as variations in the Moho topography, shows a smoothed version of the great relative minimums, maximums and gradients observed on the previous map (Fig. 6b. See also Fig. 5a). Thus, the great minimum on the SR (*m1*) stands out, related to a deeper Moho, and slightly displaced to the southeast from the lower bathymetry area. The maximum axis *M1-M3* on the GIB approximately coincides with the morphological axis of the basin. The relative maximum axis *M4* bordering the foot of the slope north of the SR, and the relative minimum axis *m2* and *m3* on the BAP, draw attention as they imply a lithospheric scale deformation, according to observations by Muñoz-Martín *et al.*, [2010].

#### 4.1.2 Mid-wavelength Bouguer anomaly map

Here, anomalies mainly related to density variations at lower and mid-crustal levels are highlighted (Fig. 6c). Thereby, once removed the regional long-wavelength component (Fig. 6b), some important anomalies are highlighted.

Among the relative maximums, stand out: 1) A striking maximum axis (*M4*) bordering the SR to the NW and to the N. This anomaly stripe may be related to the presence of high density rocks, such as oceanic basement and serpentinized mantle, previously sampled on this area of the slope [Boillot *et al.*, 1988b; Malod *et al.*, 1993; Kornprobst and Chazot, 2016]; 2) West and southwestwards from the SR, some relative maximum are emphasized (*M9-M11*) aligned on a wide band that narrows northwards. These can be linked with the existence of high density rocks at crustal levels, such as exhumed mantle, or with an extremely thin crust, or with both effects [e.g., Sibuet *et al.*, 1995; Whitmarsh and Sawyer, 1996; Louden and Chian, 1999; Dean *et al.*, 2000]; 3) South of the SR, the relative maximum axis *M5* is elongated NW-SE over the IAP, looking similar to *M9-M11* and in continuation with them, so it may have the same origin; 4) On the GIB, the *M1-M3* maximum axis, together with the gradient bands *G1* and *G2* are highlighted. *M1* is slightly shifted from the morphological basin axis. These maximums are related to the continental crust thinning, and consequently the rising of the lithospheric mantle along a relative narrow region [e.g., González *et al.*, 1999; Pérez-Gussinyé *et al.*, 2003], that lead to a mid-wavelength Bouguer anomaly; 5) *M7-M8* are located at the foot of the marginal platforms, and can be considered as a prolongation of *M4*. They may be related again with the elevation of oceanic crust and the presence of a narrow stripe of serpentinized mantle, incorporated into the morphological continental slope; 6) On the BAP, *M13* and *M14* axis are elongated coherent with main structural directions and may be linked with a thicker oceanic crust, possibly also related to a slight compressional deformation [Medialdea *et al.*, 2009].

From the relative minimums observed in this mid-wavelength map, we note: 1) the great minimum *m1* located over Galicia bank (the most prominent high on the SR) and other surrounding structural highs. Southwards, this relative minimum area continues with *m4*, centered on other structural highs. These minimums may be related to the relative thicker continental crust, and an important sediment cover [e.g., Montadert *et al.*, 1974; Wilson *et al.*, 1989]; 2) Facing the northwestern shore of Galicia, *m5* axis, NE-SW directed, is located on the MPR. These marginal platforms are former graben basins, with important sediment thickness (up to 3 s TWTT), tectonically inverted and uplifted during the Pyrenean orogeny [Murillas *et al.*, 1990]; 3) The relative minimum axis *m6*, elongated NNW-SSE, is found over the Porto basin on the continental shelf [Murillas *et al.*, 1990]; 4) NW and N of Galicia, *m7* and *m8* are located at the foot of the compressional belt. Their location and relation with surrounding maximums indicate that they could be linked with the existence of important sediment depocenters ahead of the compressional front, as previously shown by Grimaud *et al.* [1982]; 5) South of the Coruña seamounts (CS, Fig. 1), within the IAP, another large minimum is found (*m9*), also noticeable on the Bouguer and long-wavelength Bouguer anomaly maps although less striking there. This last minimum is probably related to another sedimentary depocenter, which would be the distal part of the Biscay turbidite systems [Jané *et al.*, 2011].

#### 4.1.3 Short-wavelength Bouguer anomaly map

On this map, we notice the maximum axis bordering the continental margin, close to the foot of the slope, to the NW and N of the SR (*M4*), following to the N of the GIB and at the foot of the MPR (*M7*), and to the northern margin (*M8*). Together with, and parallel to these maximum axis, three separate minimum axes are also observed to the north (*m2*, *m7* and *m10*). This minimum-maximum axis belt seems to be related to the compressional front and

the elevation of oceanic crust there (relative maximums), with probably associated basins in the foreland (relative minimums). The continuity of this anomaly belt is interrupted by means of NNE-SSW trending gradients *G9* and *G10* (See Supporting information Fig. S1). The orientation of these gradients is coherent with the main lineament orientations observed by *Maestro et al.* [2017] on this region north and northwest of Galicia, from a detailed bathymetry analysis. The NNE-SSW structural direction is observed also on land (e.g., Guimaraes-Orense-Eo, Regua-Verín, and Manteigas-Vilariça-Bragança strike-slip fault systems. See Fig. 1) [e.g., *De Vicente*, 2004; *De Vicente et al.*, 2008, 2009; *Rockwell et al.*, 2009; *Tavani*, 2012], and elsewhere offshore in the north Iberia margin for first-order strike-slip and transfer faults [e.g., *Roca et al.* 2011] connecting different segments of the Late Jurassic-Early Cretaceous rift system [*Tugend et al.* 2014, and references therein].

#### 4.2 Seismic profiling and gravity modeling: combined interpretation

Here we present the selection of the gravity models and, where available, coincident or nearby original seismic profiles interpretation. The different sections are described following a clockwise direction, covering the whole northwest Iberia margin from west to north Galicia.

As shown below, for the sake of geological consistency with adjacent interpretations and in the absence of sampling or seismic evidence, in models GM5-GM7 we have opted to model a serpentinized mantle unit, somewhat underthrust in the transition from the continental to the oceanic domains. This is possible because the densities of the serpentinized bodies are variable (from 2.4 g/cm<sup>3</sup> to 3.3 g/cm<sup>3</sup>) and in the range of the continental and oceanic basements. A more conservative model construction, without considering the existence of this unit, can be also achieved without outstanding differences in the overall shape of the model. In fact, on the gravity model GM8 (Fig. 11), coincident with the IAM-12 seismic line [*Fernández-Viejo et al.*, 1998; *Tugend et al.*, 2014], a serpentinized mantle unit can be included at the base of the continental crust, according to the interpretations by *Tugend et al.* [2014], although *Fernández-Viejo et al.* [1998] didn't make this distinction on their seismic and gravity models.

##### *GM1 model*

The gravity model GM1 crosses the main Early Cretaceous rift structures of west Galicia margin, south of the SR, in a region without remarkable tectonic inversion features (Fig. 7). It is supported by refraction seismic data on its western side [*Sibuet et al.*, 1995; *Whitmarsh et al.*, 1996; *González et al.*, 1999], and by reflection seismic profiles interpreted by *Murillas et al.* [1990], *Reston et al.* [1996] and *Ercilla et al.* [2008]. The proposed model shows a great stretching in the GIB, where a thick sediment sequence can be found (>5 km). To adjust the relative maximum at the GIB (see *M1-M2*, Fig. 5a), the Moho is modeled to be located at a minimum depth of 16 km, rising symmetrically. At the SR, the Moho is modeled at a depth up to 22 km, soaring westwards on the DGM zone. Here, the continental crust is extremely stretched, and its lower unit is delaminated until it retreated at the distance of 118 km. From this point to the 68 km distance, a high density unit of serpentinized mantle is modeled, rising westwards as a ridge, and with some upper continental crustal blocks resting on it. The serpentinized mantle unit prolongs ~20 km westwards from the distal ridge, underlining an anomalously thin oceanic basement [*Sibuet et al.*, 1995; *Davy et al.*, 2016]. Along the DGM zone, Bouguer anomaly values are higher and stable, in the range of those typical of an oceanic crust, so the previous seismic information is essential to constrain the modeling. It should be noted that there is a striking difference between the global Moho depth model [*Reguzzoni and Sampietro*, 2015] and the refraction seismic constraints along the DGM and the IAP.

### *GM2 model*

The gravity model GM2 is approximately parallel to the previous one, traced northwards, and crosses the southwest edge of the compressional front (Fig. 7). It is supported by wide-angle and reflection seismic data from *Pérez-Gussinyé et al.* [2003] and reflection seismic profiles from *Murillas et al.* [1990] and *Ercilla et al.* [2008]. To adjust the relative maximum  $M1$  on the Bouguer anomaly curve (Fig. 7, see also Fig. 5a and 6b,c), it is necessary to model a Moho rising under the stretched continental crust of the GIB. The Moho is modeled there at a minimum depth of 12 km, according to the observations of *Pérez-Gussinyé et al.* [2003]. Also on the GIB, the section shows depocenters of up to 5 km thickness [*Pérez-Gussinyé et al.*, 2003]. This Moho rising is narrower than that observed on de model GM1, asymmetric, and slightly shifted from the morphological basin axis. At the SR, the Moho is at a maximum 21 km depth (related to  $m1$ ), and eastward offset from the bathymetric minimum (Galicia bank). To the west, continental crust stretching and Moho soaring occur in a narrower band, from 85 km to 138 km distance, related to a steeper Bouguer anomaly gradient ( $G2$ , Figs. 5 and 7). As in GM1, the lower continental crust is delaminated up to disappear, and a serpentinized mantle unit is modeled at the DGM, underlining upper continental crustal blocks (85 km to 112 km distance). The serpentinized mantle unit forms a distal ridge, outcropping at the seafloor, and prolongs oceanwards ~15 km below the first thin oceanic basement. This ridge is modeled overthrusting the oceanic crust basement, according to the compressional structures observed previously in this area [e.g., *Groupe Galice*, 1979; *Grimaud et al.*, 1982; *Boillot et al.*, 1995; *Vázquez et al.*, 2008], and in agreement with the seismic interpretation and modeling along profile GM3 (see below), which crosses the compressional front slightly northwards of this model (Fig. 8). To adjust the model with the Bouguer anomaly curve, it is necessary to model a convex morphology of the serpentinized mantle unit east of the ridge; this morphology could be also related to the compressional deformation.

### *GM3 model*

The seismic and gravity section GM3 (Fig. 8) is representative for the structure of the transition to the oceanic domain in the DGM. This profile, NW-SE directed, crosses the southwest end of the compressional front from the IAP to the DGM, southeast of the Galicia bank. Thus, this section traverses the main compressional features, but is oblique to the main rifting structures. On the seismic section, three different types of acoustic basements can be identified (continental, oceanic and serpentinized mantle). To the southeast, the  $S$  reflector is interpreted as a group of bright reflectors, with relative continuity, underlining the continental crust basement [*Reston et al.*, 1996]. This continental basement seems to be extremely thinned, with faulted and tilted blocks, locally isolated. Below these tilted continental basement blocks, and with the  $S$  reflector at the top, the serpentinized mantle basement is interpreted. Northwestwards, this unit rises with a ridge shape, which outcrops at the foot of the slope [*Boillot et al.*, 1988b]. To the northwest, the sedimentary cover is intensely deformed over the oceanic crust basement close to the serpentinized crust ridge. The contact between this serpentinized mantle ridge and the oceanic crust to the northwest is interpreted as a reverse fault, with a 0.8 s (TWTT) dip-slip throw. On the continental domain, a syn-rift sedimentary sequence is identified, locally overlaying the serpentinized mantle unit, and the post-rift sedimentary sequence is only slightly deformed. On the coincident gravity model, we use the GEMMA Moho depth model [*Reguzzoni and Sampietro*, 2015] as an initial approximation, although on the final forward modeling Moho depth moves away from this global reference. This difference is also observed at the nearby GM1 model in the DGM. In order to adjust the architecture of the basement top interpreted from the coincident reflection seismic profile, it is necessary to model a wide high density unit, which corresponds to the

serpentinized mantle body, sampled nearby by *Boillot et al.* [1988b]. This unit has a distal ridge outcropping at the foot of the slope, related to the Bouguer anomaly maximum observed on this profile (at 13 km distance). Again, we have modeled a compressional feature at the contact between this ridge and the adjacent oceanic crust, and a small bulge of the serpentinized mantle unit (at 28 km distance), probably related also to a thrust. The serpentinized mantle unit prolongs more than 10 km oceanwards from the distal ridge, below a thin oceanic crust similar to that observed by *Davy et al.* [2016] further south.

#### *GM4 model*

The next section to the northwest corresponds to gravity model GM4 and the nearby seismic profiles GM4a and GM4b (Fig. 9). Seismic line GM4a is NW-SE directed, crossing the FB, and the lower continental slope, transecting the compressional front to the northwest of the SR (Fig. 9). According to the observations done over the Bouguer anomaly map (Fig. 5), most of the basement along the seismic profile is interpreted to be oceanic, faulted and tilted to the northwest, with FB acting as a sediment trap. In the compressional front, an oceanic basement thrust with a dip-slip throw of about 1.2 s (TWTT) is identified, making the oceanic basement outcrop at the foot of the continental slope. Southwest of this oceanic basement thrust, we interpret a thrust at the base of the the serpentinized mantle ridge, making it outcrop at the continental slope (according to nearby dredge samples by *Boillot et al.* [1988b], see also Fig. 1), with a large morphologic escarpment. The sedimentary cover on the oceanic crust is thick (about 1.7 s, TWTT), showing a moderate-to-intense compressional deformation. Seismic line GM4b (Fig. 9) is also NW-SE directed. It crosses from the proximal part of the northwestern flank of the SR, to the top of Galicia bank. Here, the whole acoustic basement is interpreted as continental in origin (see Bouguer anomaly map, Fig. 5), and is faulted and tilted to the northwest. The sedimentary cover thickness is variable, reaching at least 2 s (TWTT) northwest of the García structural high, where some compressional deformation can be interpreted. The nearby GM4 gravity model transects part of the oceanic fabric reliefs (FB) and the compressional front, finishing at Galicia bank (SR) (see location on Figs. 3 and 9). On its northwestern side, it coincides with the GM4a seismic profile, and the proximal half of the model is nearly parallel to GM4b seismic profile (8 km to the southwest). To construct this gravity model, we use the GEMMA Moho depth model [*Reguzzoni and Sampietro, 2015*] as a first approximation for this surface, although some discrepancies arise between the global model and the final GM4 model. The basement type and the overall basement top architecture interpreted on the nearby seismic profiles GM4a and GM4b are extrapolated on GM4. In this case, the transition from the continental Bouguer anomaly values to those regular values of the oceanic domain is narrower, with the exhumed mantle unit outcropping at the foot of the slope (sampled by *Boillot et al.*, [1988b]). Differing from the previous models, the HME domain is narrower. Here, the upper continental crust seems to be moderately thickened close to the transition to the oceanic domain (consistent with the compressional deformation observed on the seismic section GM4b), and is modeled with a relatively constant thickness along the whole model. To fit with the Bouguer anomaly values, a relatively thin lower continental crust density unit is modeled next to the serpentinized mantle body, and the upper continental crust unit is modeled with a relatively constant thickness of 6 km. A 3 km depocenter is modeled close to the serpentinized mantle ridge to adjust the relative Bouguer anomaly inflexion (relative minimum at 87 km distance) in the gradient, similar to the depocenter observed on the seismic profile GM4b. A reverse structure, uplifting the oceanic basement unit, is modeled to accommodate the small relative maximum at 63 km distance, consistent with the oceanic basement thrust observed on the GM4a seismic profile.

### *GM5 model*

Northwards, the next section crosses from the BAP to the north of the SR, spanning the compressional lower slope and the Ordoño structural high, following on the SR parallel to the eastern slope of Galicia bank (7 km away, see Fig. 10). Seismic lines GM5a and GM5b are interpreted together, as they are continuation from each other, and close to them, the gravity section GM5 is modeled (Fig. 10). Basaltic oceanic seafloor dredge samples were recovered at the foot of the slope in the vicinity of this profile [Boillot *et al.*, 1988b]. As there is no additional information to distinguish a serpentinized mantle basement, on the seismic profile, only oceanic and continental basement types are interpreted. The continental basement is interpreted on most of the composed seismic section, affected by normal faulting, although the basement top is not easily identified along the southern half of the profile. No first order tectonic features are observed on the southern half of the section because it is parallel to the main structural direction in this region. To the north, the continental basement outcrops on the Ordoño structural high, limited by normal faults, and also at the lower slope. The contact between the continental and the oceanic basements is made by means of a reverse fault that places a continental basement block over the oceanic basement, resulting on a morphological escarpment at the continental slope. A proximal oceanic basement block overthrust is interpreted to the north on the compressional front (vertical displacement of about 0.5 s, TWTT), outcropping at the foot of the slope. The sedimentary cover is slightly deformed close to the compressional front, and no compressional deformation is observed on the sedimentary cover of the continental crust south of the Ordoño high. To construct the gravity model GM5, the GEMMA Moho depth model [Reguzzoni and Sampietro, 2015] is used, and the nearby composed seismic profile is used to approximately constrain the basement top morphology. A serpentinized mantle unit can be postulated in the transition between the oceanic and the continental basement units, as an underthrust body at the base of the compressional belt (similar to that interpreted north of Galicia by Tugend *et al.*, [2014]). The modeled continental crust thickness is fairly constant (about 14-18 km), from 60 km distance to the SSE edge of the model. From 60 km to 25 km distance, the continental crustal thinning is modeled as a thinning of the lower crust unit, underlined by the serpentinized mantle unit. The transition from the continental to the oceanic domain consists of a first order thrust. To adjust the Bouguer anomaly maximum at 16 km distance (*M4* on Figs. 5a and 10), an oceanic crust thickening is modeled, by means of another thrust that makes the oceanic basement outcrop at the foot of the slope, according to the GM5a seismic profile interpretation (Fig. 10).

### *GM6 model*

The GM6 gravity model (Fig. 11) starts on the BAP, transects the SR east of Galicia bank, and finish on the IAP. From the southern part of the SR to the south edge, this model is nearly coincident with the reflection and refraction seismic profile ISE-9 [Clark *et al.*, 2007]. Apart from the mentioned refraction model, the GEMMA Moho depth model [Reguzzoni and Sampietro, 2015] is used as an approximation in the northern half of the section. The modeled continental crust shows significant differences north and south of 240 km distance. Northwards, the thickness of the upper and lower levels is similar. Southwards, to keep the Moho depth given by Clark *et al.* [2007], it is necessary to model a relatively thicker lower continental crust unit to adjust the observed Bouguer anomaly. On the southern half of the model, the steeped gravity gradient is related to thickenings of the lower continental crust unit, and upper continental crust blocks, probably faulted and tilted to the south. Southwards from 340 km distance, lower continental crust is thinned until it disappears (at 370 km distance, *G4*. See also Fig. 5a). The upper continental crust is also extremely thinned, and the whole crustal thickness is less than 10 km. Southwards, on the IAP, it is necessary to model

an anomalous, high density basement, with its base at a depth of 14 km. This basement could be similar to that modeled at the DGM (models GM1, GM2 and GM3, Figs. 7 and 8), probably composed of serpentized mantle, and is consistent with the observations made on the traverse seismic reflection and refraction profile [Pickup *et al.*, 1996; Dean *et al.*, 2000] that is crossed by this model on its southern edge (Fig. 3). On the northern half of the model, the regular Bouguer anomaly gradient is related to a thinning structure similar to that modeled at the GM5 model: thinning of the lower continental crust unit, and a thrust of the continental crust over the oceanic one, with an underthrust serpentized mantle unit. The relative Bouguer anomaly maximum at 87 km distance (*M4*), close to the continent-ocean transition, is related to oceanic basement overthrust (see Fig. 5a and 6c, d). This oceanic basement thrust is modeled involving a flexure of the footwall, with an associated depocenter, that accommodates the relative mid-wavelength minimum on the Bouguer anomaly curve (*m2* on Figs. 5, 6 and 11).

#### *GM7 model*

The gravity model GM4 is NNW-SSE directed. It starts on the BAP, crosses the MPR on their northeast sector, and finishes on the continental shelf (Fig. 11). As an approximation to Moho depth modeling, we have used both the global GEMMA model [Reguzzoni and Sampietro, 2015], mainly uses to constrain the continental and transition areas, and a projection of the Moho modeled from the nearby IAM-12 refraction profile [Fernández-Viejo *et al.*, 1998], particularly to constrain Moho depth on the oceanic domain. The smooth Bouguer anomaly gradient over the BAP is adjusted in the model as a progressive deepening of the oceanic crust basement. The relative Bouguer anomaly minimum-maximum association (at 100 km and 115 km distance, respectively) is solved, as previously on the GM6 model, by means of an oceanic basement overthrust (*M8* on Figs. 5 and 7) uplifting the oceanic basement in the continental lower slope, and a small foredeep-type basin (see Supporting information Fig. S2). The oceanic basement overthrust would have a morphological imprint as a relief at the foot of the slope. The transition to the continental domain is noticeable on the Bouguer anomaly profile as a clear inflexion of the curve, with a steeper gradient on the continental domain. This transition is modeled as a reverse feature, with an underthrust serpentized mantle unit, similar to the one interpreted on the nearby IAM-12 seismic profile [Tugend *et al.*, 2014]. The upper continental crust unit has a nearly constant thickness, meanwhile the lower continental crust thins oceanwards. The short wavelength relative minimums on the continental domain are modeled as local depocenters originated during the extensional stage on this margin sector, and partially inverted under the compressional regime [Murillas *et al.*, 1990].

#### *GM8 model*

The GM8 gravity model is N-S oriented, starting at the BAP, crossing the compressional front north of Galicia, and finishing on the continental shelf at the same point as the GM7 model (Fig. 11). It is coincident with the seismic model IAM-12 (see Fig. 3) [Fernández-Viejo *et al.*, 1998]. Moho depth is modeled from the IAM-12 results [Fernández-Viejo *et al.*, 1998], which differs from the GEMMA Moho depth model [Reguzzoni and Sampietro, 2015] (Fig. 11). The main Bouguer anomaly gradients are fitted in the model by the Moho southwards deepening. At the continent-ocean transition, the relative maximum amplitude (at 157 km distance) is lower than that observed on the previous model profiles. Thereby, and according to the previous works [Álvarez-Marrón *et al.*, 1997; Fernández-Viejo *et al.*, 1998], it is not necessary to model an oceanic basement thickening, and the continental domain reaches the foot of the slope. To adjust the mid-wavelength minimum located at 144 km distance, it is necessary to model a 2.3 g/cm<sup>3</sup> density block at the foot of the slope, which can

be related to the accretionary sedimentary wedge [Tugend *et al.*, 2014], that is coherent with the seismic velocity observed there by Fernández-Viejo *et al.* [1998]. To fit the mid-wavelength relative maximum at 160 km distance, a high density block is modeled at the base of the continental crust, which could correspond to an underthrust serpentinized mantle relict [Tugend *et al.*, 2014].

### 4.3 Structural domains mapping

As explained before (sub-section 1.1), geometrical and structural-based domain partitioning of rifted margins has been suggested [Péron-Pinvidic *et al.*, 2013; Sutra *et al.*, 2013; Tugend *et al.*, 2015a]. An approximation from both gravity and magnetic signatures has been recently made by Stanton *et al.* [2016] for the Iberia-Newfoundland conjugate margins at a regional scale.

The detailed gravity analysis that we have achieved, allows an initial approach to the continental and oceanic domains mapping along the margin. Distinct morphostructural features observed along the margin, are associated to different deformation processes. Moreover, these deformation processes may leave an imprint on the Bouguer anomaly signal. Except for the intense compressional deformation domain, most of the structural domains described here are rift-related (described above on section 1.1). The boundaries between these domains are usually diffuse, and the intense compressional deformation domain overlaps the others.

Here, we propose additional criteria for a detailed characterization of rift domains along the Galicia margin (Fig. 12), based on gravity data analysis and modeling. To accomplish this continuous mapping, we first identify each domain on the gravity (seismic-supported) models. Subsequently, gravity maps allow an extrapolation to complete the domain mapping gaps between models.

#### 4.3.1 Rift domains

##### *Stretched domain*

This domain is mapped bordering the continental shelf to the west margin and northwest of Galicia (MPR), and on the SR (Fig. 12). In this domain, crustal thinning is moderate, and faulting of the upper crust generates small graben and half-graben basins. Bouguer anomaly values are intermediate, from 90 mGal to about 200 mGal (Fig. 5a). The SR is mapped as *stretched domain* because, although crustal thickness here is less than 30 km, basement top and Moho surfaces are approximately parallel, and the characteristic wedge shape of the crust of a necking domain is not observed [Péron-Pinvidic *et al.*, 2013]. The Moho is found at a depth between >22 km in the continental shelf region, and 18-21 km in the SR, where the staged rift evolution leaved this continental ribbon [Naliboff *et al.*, 2017] (see Figs. 7, 9-11).

##### *Necking domain*

On the study area, crustal thickness in the necking domain usually decrease from 22 km or less to 10 km, then there is a Moho inflection and it rises, tending to converge with a deepening continental basement top. Unlike the domain mapping made by others [e.g., Tugend *et al.*, 2014; Chenin *et al.*, 2015], considering a crustal thickness between ~30 and 10 km criterion, the primary use of the specific wedge shape of the crust approach, with converging Moho and basement top surfaces [Péron-Pinvidic *et al.*, 2013], allow us to distinguish two main necking regions on the west Galicia margin. These are:

- a) The necking domain of the GIB, where intense thinning can be associated to the first rifting episode (Berriasian-Valanginian) [Murillas *et al.*, 1990]. Here Moho rises from 22-

18 km to 12 km depth [Pérez-Gussinyé *et al.*, 2003] (Fig. 7), and this ascent is associated to a positive mid-wavelength Bouguer anomaly (Fig. 6c). Locally, crustal thickness is slightly below 10 km thick (GM2, Fig. 7) and it is possible that embrittlement of the whole crust could have been reached. As no fault penetration is observed in the mantle at that point [Pérez-Gussinyé *et al.*, 2003], the westward jump of the rift axis should have occurred just before the onset of hyperextension in the GIB.

- b) The necking domain westwards of the SR, where the intense crustal thinning is related to the final rifting stage (Hauterivian-Aptian) [Murillas *et al.*, 1990]. Here, there is a correlation between this domain and the Bouguer anomaly gradient (see *G1*, *G2* and *G4* on Fig. 5a), similar to that observed by Stanton *et al.* [2016].

North of the SR and northwest of Galicia, the compressional deformation overprint avoid an accurate identification of the distal boundary of this domain, and we are not able to identify the points of 10 km crustal thickness in the former rifted margin. However, along this sector, we interpret a probable remnant of underthrust serpentized mantle (Sub-section 4.2, Figs. 10-11), similar to that interpreted by Tugend *et al.* [2014]. This fact lead us to map a minimum (overthickened) extent of the necking domain along this sector (Fig. 12), although it is not possible to delineate a net coupling line marking the transition from the necking to the HME domain (dashed purple line on Fig. 12).

#### *Hyperextension and mantle exhumation (HME) domain*

This domain is related to a positive mid-wavelength Bouguer anomaly (Fig. 6c), partly due to the high density serpentized mantle basement but also due to the rising of the unaltered mantle below it (Fig. 7). West of the SR, this domain narrows northwards, and north of the 43°N latitude, it is only plainly noticeable by the relict serpentized mantle ridge, up to 43°20'N [Boillot *et al.*, 1988b]. From this this latitude on the west margin to north of Galicia, the partial tectonic inversion overprints the former HME domain structure, although a serpentized mantle unit can be modeled as an underthrust body in the compressional front (Figs. 10 and 11). The tectonic inversion hampers an accurate identification and delineation of the HME domain (as those accomplished by Sutra *et al.* [2013] or Tugend *et al.* [2014]) so, in the absence of further information to undertake a finer mapping, only a blurry boundary between the necking and the HME domain can be established (Fig. 12). North of Galicia, and east of 7°30'W, a positive mid-wavelength Bouguer anomaly is noticeable (*M10*, Fig. 6c). This anomaly can be related to the western edge of the mantle exhumation domain mapped by Tugend *et al.* [2014, 2015a] on the BAP.

#### 4.3.2 Intense compressional deformation (ICD) domain

Here are included regions where first order reverse tectonic features are interpreted and/or modeled, such as the great thrusts identified on the seismic profiles (Figs. 8-10) and modeled to fit the gravity curves (Figs. 7-11). In addition, this domain encloses also those areas where great magnitude tectonic inversion evidences can be deduced. These last evidences include:

- a) The greater thickness of the previously, extremely stretched continental crust during the rift episode: In addition to the tectonic inversion evidences in the MPR [Murillas *et al.*, 1990], gravity modeling and previous researches [Álvarez-Marrón *et al.*, 1996; Fernández-Viejo *et al.*, 1998] show that, in the lower continental slope, the continental crust is more than 10 km thick. As margin hyperextension occurred during the Bay of Biscay opening [Tugend *et al.*, 2014], here crustal thickness must have been less than 10 km [Sutra *et al.*, 2013]. The fact that the tectonic inversion has thickened the continental crust in the HME domain is then an evidence of intense compressional deformation. From gravity modeling, we can calculate a thickening of the continental crust of at least of 2 km

north of the SR (in the same order of magnitude as calculated by *Boillot et al.* [1979]) and of ~5 km northwest of Galicia (MPR).

- b) The overprinted structure of the HME domain: This domain is clearly identified in the west Galicia and north Iberia margins [e.g., *Péron-Pinvidic et al.*, 2013; *Tugend et al.*, 2014], as well as in the conjugate margin [*Thinon et al.*, 2003]. But, along the northwest Galicia margin, this domain appears shortened and thickened (from the north of the SR to the north of Galicia), and there is an apparent absence of a mantle exhumation region. In this work, we have modeled a (postulated) serpentized mantle unit north of the SR and northwest of the MPR, appearing as a relict wedge underthrust at the base of the continental crust [*Tugend et al.*, 2014], although its geophysical characteristics and the scarcity of information make difficult to identify it [*Fernández-Viejo et al.*, 2012].

Beyond the ICD domain boundaries, some compressional tectonic features, indicative of a less intense deformation, can be identified. Among others, stand out small sub-vertical faults related to slight reactivation of continental crust blocks, and soft folding and long-wavelength positive bulges, are identified in the seamounts region [*Vázquez et al.*, 2008].

## 5 Discussion

### 5.1 Continent-ocean boundary delineation

The location of the continent-ocean boundary (COB) in magma-poor rifted margin is a matter of controversy. Usually, the main criteria for the determination of this limit are:

- 1) The localization of a distal serpentized mantle ridge [e.g., *Boillot et al.*, 1988a]: The use of the distal serpentized mantle ridge to delimit the COB may be ambiguous, as some times there are two parallel ridges, as in the IAP south of the SR [*Whitmarsh et al.*, 1996], and its along-strike continuation is difficult to map in the absence of seismic and sampling criteria. Moreover, the use of the correlation of unconformity between the syn- and post-rift sedimentary sequences with the beginning of the continental drift is under discussion [*Péron-Pinvidic et al.*, 2007; *Sibuet et al.*, 2007; *Tucholke et al.*, 2007].
- 2) The presence of isochrones, related to oceanic accretion [*Sibuet et al.*, 2004a; *Müller et al.*, 2008]: The recognition of the first isochrones of the new accreted oceanic crust is ambiguous in magma-poor rifted margins, as some magnetic anomalies can be identified on the exhumed lithospheric mantle [*Sayers et al.*, 2001; *Sibuet et al.*, 2007; *Direen et al.*, 2007; *Bronner et al.*, 2011; *Stanton et al.*, 2016; *Kornprobst and Chazot*, 2016; *Nirrengarten et al.*, 2016]. Moreover, on the west Iberia margin, the beginning of the continental drift occurs during the Cretaceous magnetic quiet lapse.
- 3) The absence of continental crust [*Boillot and Frotzheim*, 2001]: This proposal of the distal edge of the continental crust as the boundary between the oceanic and the continental domains is approachable for discrete seismic profiles, but it would result in a complex limit for a regional mapping, as many isolated continental crust blocks are left on the hyperextended region, and much seismic coverage would be needed.

Here we consider the COB as a simplified line that marks the landward edge of the oceanic crust [e.g. *Autin et al.*, 2010; *Ball et al.*, 2013; *Sutra et al.*, 2013], even if the oceanic crust is anomalously thin, like west of Galicia [*Sibuet et al.*, 1995; *Davy et al.*, 2016], probably being part of a complex proto-oceanic domain [*Davy et al.*, 2016; *Gillard et al.*, 2015]. The available information for the present study does not allow differentiating this probable proto-oceanic domain from the steady state oceanic crust. The detailed gravity analysis and modeling addressed here, together with other geophysical, geological and sampling information, has allowed delineating a continuous and consistent COB (as defined here)

along the whole Galicia margin. The only observation of the Bouguer anomaly map is not sufficient to map this border, as values vary from 90 mGal to 350 mGal due to the complex geometry of the margin and of the transition to the oceanic domain (Fig. 5a). It was necessary to perform a detailed geophysical data processing and analysis to delineate this limit. Thus, frequency filtering and 2+3/4 D modeling from the Bouguer anomaly grid highlighted the effects of the compressional tectonics over the igneous oceanic basement, which is found uplifted and overthrusting at the foot of the slope of the MPR and north and northwest of the SR (Figs. 6, 9-11).

## 5.2 Role of the compositional and structural inheritance in the rift segmentation

The studied area was tentatively divided into different along-strike segments which are, from south to north: south Iberia abyssal plain segment (SIAP), west Galicia segment (WG), northwest Galicia segment (NWG) and north Iberia segment (NI). Starting from the south, the SIAP segment of the Iberia margin shows relatively narrow stretched and necking domains, and a wide HME domain (Fig. 12). The transition to the WG segment is made by means of a NW-SE structure (1 at Fig. 12), characterized by the Bouguer anomaly gradient  $G4$  (Fig. 5) that connects the SR relative minimums with the  $M5$  relative maximum over the south IAP (Figs. 6b and 6c). On the WG segment, the main features are NNW-SSE directed (Fig. 1a,  $M1$ ,  $M2$  and gradient band  $G1$ - $G2$  on Figs. 5 and 6), cross-cut by NW-SE and NE-SW features ( $G7$  and  $G8$ , Figs. 5 and 6) that probably acted as transfer zones during rifting [e.g., *Murillas et al.*, 1990]. Northwards, there is a complex transition to the north Iberia margin, with an interference of both the main structural directions related to the west and the north Iberia rifts. Thus, the NWG segment is marked by various directional changes on the main margin structures (from NE-SW northwest of the SR, to E-W north of the SR, to NE-SW northwest of Galicia, see Figs. 5 and 6), also noticeable on detailed quantitative analysis of bathymetric lineaments [*Maestro et al.*, 2017]. These orientation changes seem to be favored by NNE-SSW features (2-5 at Fig. 12), identified as Bouguer anomaly gradients and morphotectonic lineaments ( $G9$  and  $G10$  at Fig. 6d; see Supporting information Fig. S1). These NNE-SSW structures may have acted as transfer zones in this margin segment, influenced by the north Iberia margin tectonics, where rift occurs oblique to the main divergence directions in a transtensional stress regime context [e.g., *Jammes et al.*, 2009, and references therein], as is the case of the Gulf of Aden [*Bellahsen et al.*, 2013, and references therein]. The structural sketch of this sector of the margin is also similar to those obtained by analog modeling for an oblique rift [*Bellahsen et al.*, 2013, and references therein], probably favored by the preexisting NE-SW Late Triassic-Early Jurassic structures [e.g., *Vegas et al.*, 2016]. The NI segment is found east of the last NNE-SSW transfer zone (5, Fig. 12). At this sector, the main margin features are approximately E-W directed [*Maestro et al.*, 2017], continuing eastwards on the Bay of Biscay [e.g., *Roca et al.*, 2011; *Tugend et al.*, 2014, and references therein].

It seems clear that rift segmentation along the Galicia margin is highly influenced by preexistent rheological discontinuities and tectonic fabric. Many studies have related the influence of first order rheological heterogeneities on the strain localization during continental rifting and breakup [e.g., *Dunbar and Sawyer*, 1989; *Muntener and Manatschal*, 2006; *Sutra and Manatschal*, 2012; *Ball et al.*, 2013, *Manatschal et al.*, 2015]. In the west Galicia margin, both compositional and structural inheritances appear to condition rift development [*Chenin et al.*, 2015; *Manatschal et al.*, 2015]. Regarding compositional inheritance, seismic velocity modeling demonstrates that the GIB, where initial Berriasian-Valanginian rift axis was centered [*Murillas et al.*, 1990], originated at the border between two different Paleozoic terrains of the Hercynian belt [*Pérez-Gussinyé et al.*, 2003], which

are the Central Iberian and the Ossa-Morena zones (Fig. 1b) [Capdevila and Mougénot, 1988; Mamet et al., 1991; Martínez Catalán et al., 2009 and Manatschal et al. 2015]. Moreover, the separation between the WG sector and the SIAP sector is made by means of the NW-SE transfer zone mentioned above (1 on Fig. 12). This transfer zone, evidenced by bathymetry (as it coincides with the southern border of the SR), and gravity anomaly analysis, is approximately located on the Hercynian deformation front (Fig. 1b), separating two terrains (Avalonia and Gondwana) with distinct rheological behavior [Martínez Catalán et al., 2009; Mohn et al., 2015; Manatschal et al., 2015]. With respect to the structural inheritance, the whole Iberia rift system seems to surround the Hercynian core [Lefort and Haworth, 1979; Capdevila and Mougénot, 1998; Bowling and Harry, 2001; Chenin et al., 2015]. Rifting is apparently favored by first order “weak” structures corresponding to Hercynian thrust fronts and strike-slip faults (Fig. 1b), but also possibly conditioned by pre-rift (Triassic and Late Triassic-Early Jurassic) strike-slip faulting [Ziegler, 1982; Sopeña et al., 1988; Vegas et al., 2016]. This pre-existent fabric seems to control the main structural directions under extension (evidenced by the different gravity anomaly lineaments described above), at both the northern and western segments, as these coincide approximately with the main thrusts and strike-slip faults directions observed onshore (Fig. 1a and 1b) [Ramírez et al., 2006]. The influence of the Triassic tectonic event [e.g., Sopeña et al., 1988] on the margin segmentation is evident in the NNE-SSW trending structures (G9 and G10 on Fig. 6d, and 2-5 on Fig. 12). These structures probably acted as transform faults during the oblique rifting north of the SR and northwest of Galicia, allowing the changes on the main structural directions as rifting progressed towards the north Iberia margin. The NNE-SSW structural direction manifests also on land [e.g., Rockwell et al., 2009], as left-lateral strike slip faults under the compressional regime [De Vicente, 2004; De Vicente et al., 2008, 2009]. The NE-SW and NW-SE Late Triassic-Early Jurassic rift structures [Vegas et al., 2016] act as strike-slip fractures in the west Iberia margin, meanwhile NE-SW structures behaves as normal faults in the NWG segment during the Cretaceous rifting.

### **5.3 Influence of rift architecture on the partial tectonic inversion of the margin**

Rift-related architecture has been proven to influence the margin tectonic inversion and in subsequent subduction and collision processes. In this context, the HME domain plays a key role, as the serpentized mantle unit acts as a detachment surface, facilitating the inversion processes [e.g., Tugend et al., 2014; Chenin et al., 2017, and references therein].

Our observations of the deep structure of the Galicia continental margin allow us to propose a simplified conceptual model for the initiation of the tectonic inversion of the magma-poor rifted margin north and northwest of Galicia (Fig. 13). In this model, the initial HME domain is the first region to be inverted and which accommodates most of the compressional deformation, as mantle partial serpentization leads to weakening, making this domain prone to compressive deformation [Lundin and Doré, 2011]. The location of the compressional deformation in this region is favored by the reactivation of the former detachment at the base of the continental crust, which originated during the rifting episode (Fig. 13a). At the outer flank of the HME domain, the serpentized peridotite ridge acts as a weakness zone that favors the nucleation of a thrust under the compressional regime (Fig. 13b) [Lundin and Doré, 2011; Fernández-Viejo et al., 2013; Tugend et al., 2014], such as that observed west and north of the SR (Figs. 7-9). Some deformation can develop also towards the continent in the serpentized mantle unit, as those convex morphologies modeled at the GM2-GM4 that could be related to thrusts (Figs. 7-9). With the compressional deformation progress, the tectonic inversion propagates along the detachment surface towards the continent in the HME domain, and some normal faults are also reactivated in the continental basement. The

proximal oceanic crust is anomalously thin and may be also underlined by serpentinized peridotite, entailing a weakness zone under the compressional regime (Figs. 7-9) [Sibuet *et al.*, 1995; Whitmarsh *et al.*, 1996; Thinon *et al.*, 2003; Davy *et al.*, 2016]. As compression advances, new thrust features can develop in this thin and weak oceanic crust, with a foreland shifting of the deformation front (Fig. 13c). This result on the formation of oceanic basement in sequence thrusts as those observed north of the SR and north and northwest of Galicia (Figs. 9-11). If tectonic inversion make progresses, the reactivation of previous normal faults on the continental basement would advance towards the necking and stretching domains, leading to a deformation of the post-rift sedimentary cover, as previously noticed northwest of Galicia [Murillas *et al.*, 1990] and in the SR [Vázquez *et al.*, 2008].

The compressional belt bordering the SR and the northwestern margin of Galicia can be considered as an intermediate stage between a hyperextended margin and a “continental margin type” ophiolitic belt [Dilek and Furnes, 2011]. If compression had progressed to a collisional orogeny here, the remnants of the serpentinized mantle ridge, together with the overthrust oceanic basement blocks, could have been obducted in the accretionary prism, and exposed afterwards as an ophiolitic belt such as those observed at the Pyrenees [e.g., Jammes *et al.*, 2009; Lagabrielle *et al.*, 2010; Mouthereau *et al.*, 2014] and the Alps [Whitmarsh *et al.*, 1993; Manatschal, 2004; Mohn *et al.*, 2014; Beltrando *et al.*, 2014]. Conversely, the underthrust remnants of the serpentinized mantle body would become a component of the subducted part of the orogeny [Chenin *et al.*, 2017].

## 6 Conclusions

The use of gravity analysis, in combination with seismic and sampling data is an adequate methodology for performing structural domain mapping. On the northwestern Galicia margin, where no good coverage of seismic information is available, detailed gravity analysis has allowed an extrapolation of the seismic and sampling constrained interpretations, and a first approach to this mapping is achieved. This methodology is also useful for determining the across and along-strike structure of the margin.

The transition from continental to oceanic domain west of Galicia and south of 42°20'N latitude is very progressive. The necking and the HME domains were mapped on this sector of the margin. The HME domain narrows northwards with a wedge shape from approximately 50 km wide at 42°N latitude until it disappears north of 43°N. The serpentinized mantle ridge prolongs until 43°20'N, with no surficial evidence of its existence further north of the SR, where a rotation of the main tectonic features is observed. From south to north, the serpentinized mantle ridge displays morphological variations, from being symmetrical in the south to show a steeper distal flank northwards, where its contact with the oceanic crust occurs through overthrusting. North of the SR and the GIB, and at the foot of the MPR, the continent-ocean transition is more abrupt and is modeled as a major continental basement thrust over the oceanic one, with an underthrust serpentinized mantle unit at the base of the continental crust. Here, the boundary between the necking and the HME domain is hardly established, as it is overprinted by the tectonic inversion of the margin.

The detailed gravity analysis has allowed a continuous mapping of the crustal and structural domains along the margin, and strengthens the utility of the gravity method for such mapping. As a result, an accurate delineation of the COB has been mapped, as well as the coupling line that marks the limit between the necking and the HME domains. Because a superimposed ICD domain also has been mapped, a compressional deformation front has been delineated.

The studied area has been divided in four different segments, according to their compositional and structural characteristics: SIAP, WG, NWG and NI. Different sectors seem to develop in distinct Hercynian belt zones, with the segmentation favored by preexisting weak zones that act as first order transfer zones during rifting. In turn, rift architecture strongly conditions its evolution under the subsequent compressional regime, with the HME domain focusing most of the compressive deformation. The observations made on the deep structure of the Galicia continental margin have led us to propose a simplified model for the initiation of the tectonic inversion of a hyperextended rift margin. In accordance with previous works, the HME domain accommodates most of the compressional deformation. As tectonic inversion progresses, the first oceanic basement, anomalously thin and underlined by serpentinized mantle, is faulted and overthrusts, remaining uplifted at the base of the continental slope.

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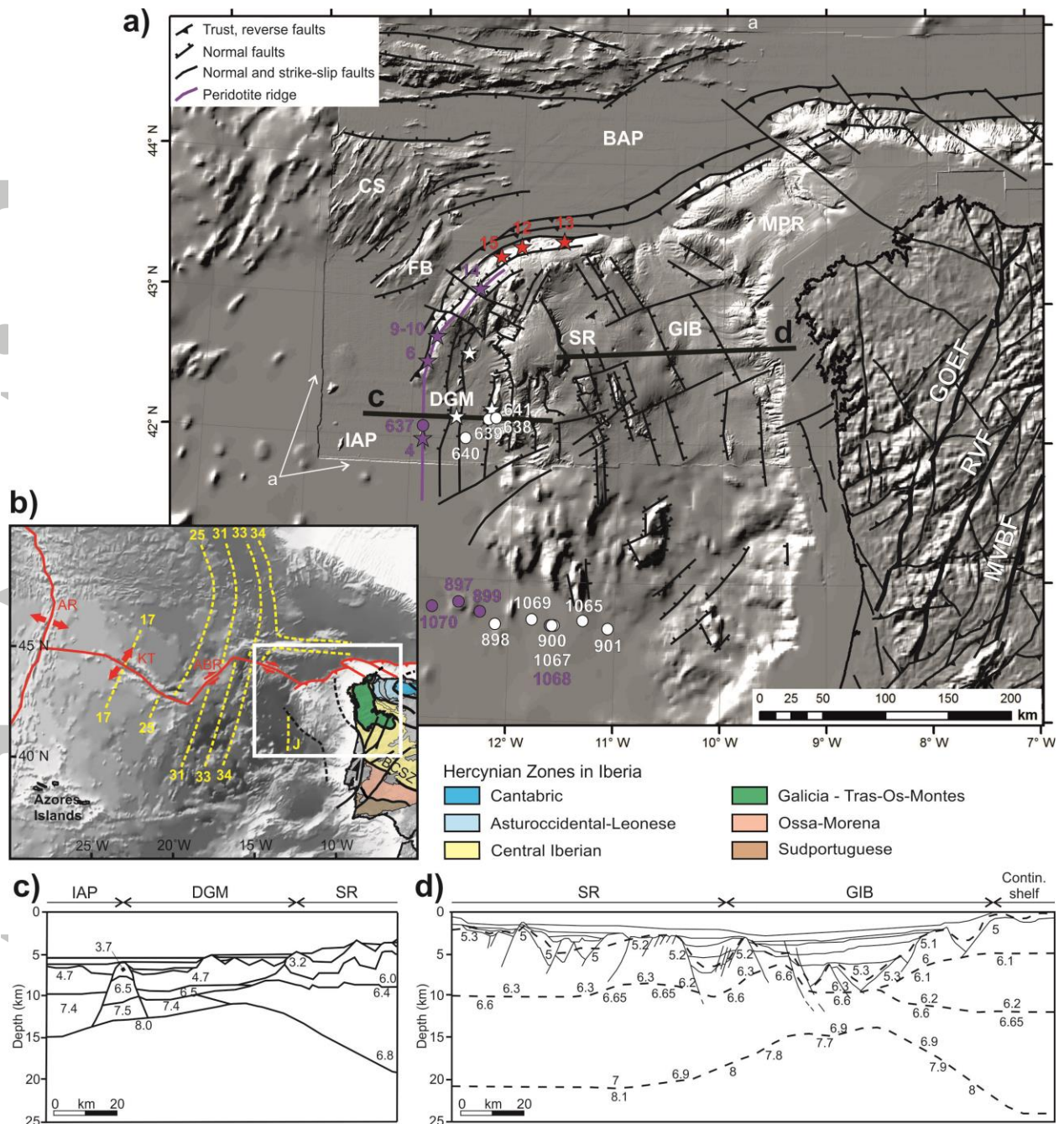
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**Table 1:** Density characteristics of the units modeled (Figs. 7-11). Seismic velocities from Córdoba *et al.* [1987], Sibuet *et al.* [1995], Whitmarsh *et al.* [1996], González *et al.* [1999] and Pérez-Gussinyé *et al.* [2003]. Correlation between seismic velocities and densities made with the aid of the empiric relationship curves by Ludwig *et al.* [1970], Barton [1986], Christensen and Mooney [1995] and Brocher [2005].

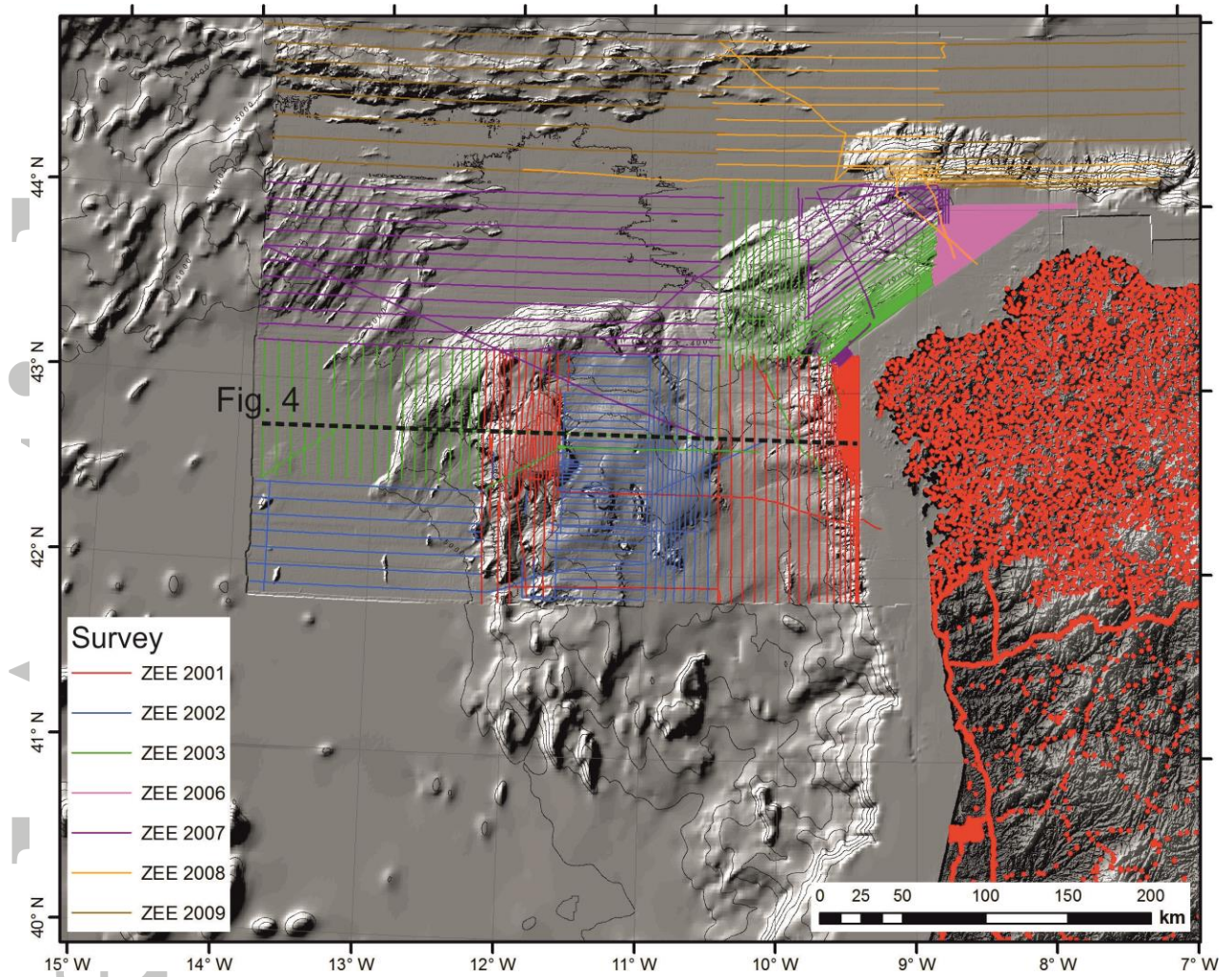
<i>Unit</i>	<i>P-wave velocity (km/s)</i>	<i>Density (g/cm<sup>3</sup>)</i>
Mantle	8	3.3
Serpentinized mantle	5.5 - 7.5	2.6 - 3.15
Lower oceanic crust	7.4	3.1
Upper oceanic crust	6.5	2.8
Lower continental crust	6.6	2.85
Upper continental crust	5.5	2.6
Syn-rift sediments	2.5	2.4
Post-rift sediments	2	2.1



**Figure 1.** a) Structural sketch of northwest Iberia, over the digital terrain model (DTM) obtained by the combination of ship data and the GEBCO Digital Atlas. Modified from *Groupe Galice* [1979], *Boillot et al.* [1988b, 1995], *Grimaud et al.* [1982], *Boillot and Malod* [1988], *Murillas et al.* [1990], *Malod et al.* [1993], *Álvarez-Marrón et al.* [1997], *Ramírez et al.* [2006] and *Vázquez et al.* [2008]. BAP, Biscay abyssal plain; MPR, marginal platforms region; CS, Coruña seamounts; FB, Finisterre bank; SR, Seamounts region; DGM, deep Galicia margin; GIB, Galicia interior basin; IAP, Iberia abyssal plain; GOEF, Guimaraes-Orense-Eo fault; RVF, Regua-Verín fault; MVBF, Manteigas-Vilariça-Bragança fault; a, artifact on the DTM by the ship and GEBCO assemblage. Labeled circles show locations of DSDP and ODP drill holes [*Groupe Galice*, 1979; *Boillot et al.*, 1988a; *Whitmarsh et al.*, 1998] whereas stars show locations of dredge sampling [*Boillot et al.*, 1988b]. Purple locations indicate sampling of the exhumed mantle (serpentinite, serpentinized peridotite), while red locations indicate sampling of oceanic basement rocky outcrops. White locations

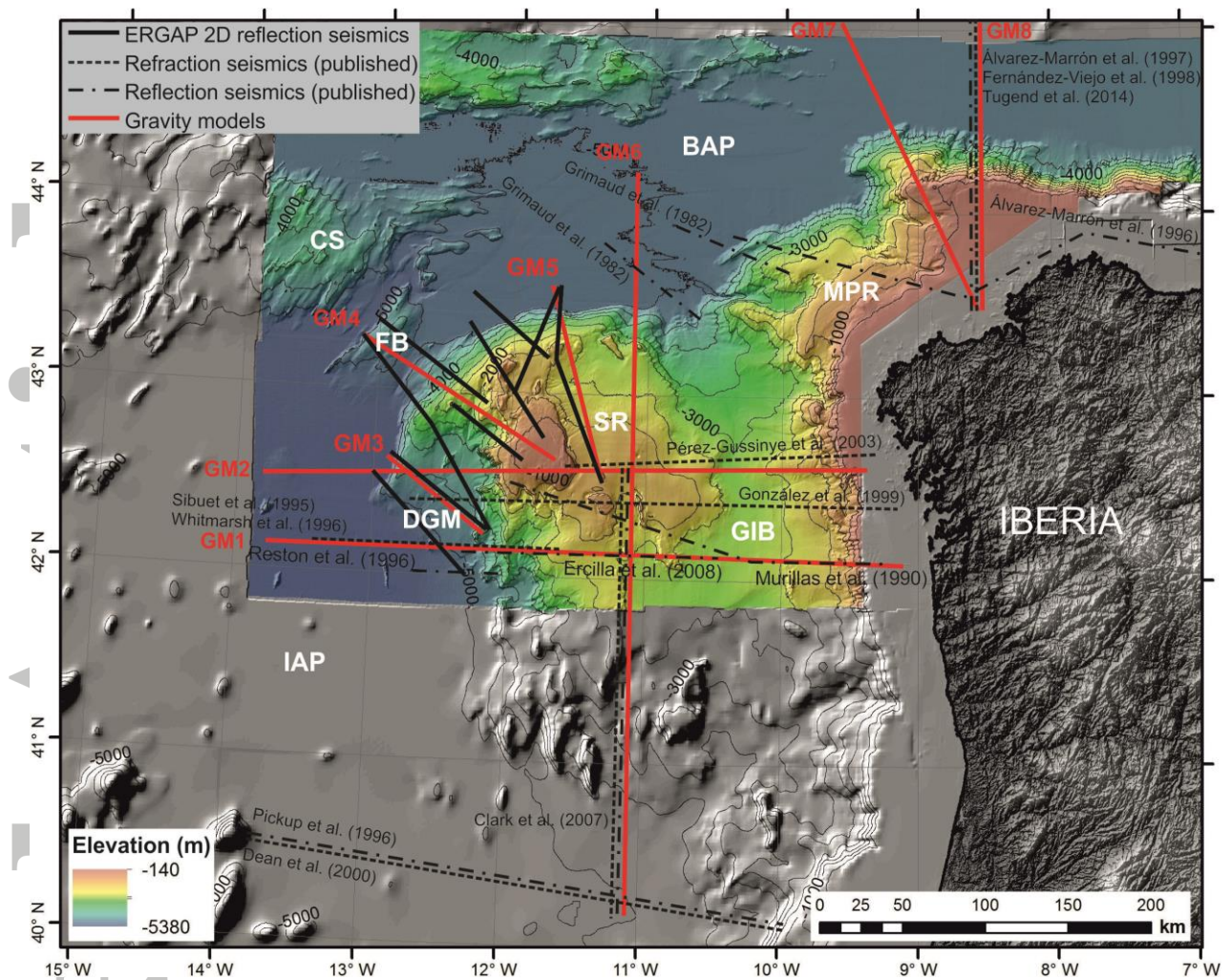
indicate either that continental basement was sampled or that the sampling didn't reach the basement. **b)** Regional map with location of the study area (white rectangle), main fracture zones and plate boundaries during Eocene times (red solid lines), magnetic anomalies on the oceanic domain (yellow dashed lines), and, colored, tectono-metamorphic zonation of the Hercynian belt in Iberia [*Pérez-Estaún and Bea, 2004*], with some of their offshore prolongations (black dashed lines), based on *Capdevila and Mougenot [1988]*, *Martínez Catalán et al. [2009]* and *Manatschal et al. [2015]*. AR, Atlantic ridge; KT, King's trough; ABR, Azores-Biscay rise; BCSZ, Badajoz-Córdoba Shear Zone. **c)** Refraction seismic velocities model on the continent-ocean transition, modified from *Sibuet et al. [1995]*. **d)** Refraction seismic velocities model on the GIB, and geological interpretation, modified from *Pérez-Gussinyé et al. [2003]*. Seismic velocity unit on **c** and **d** is km/s.

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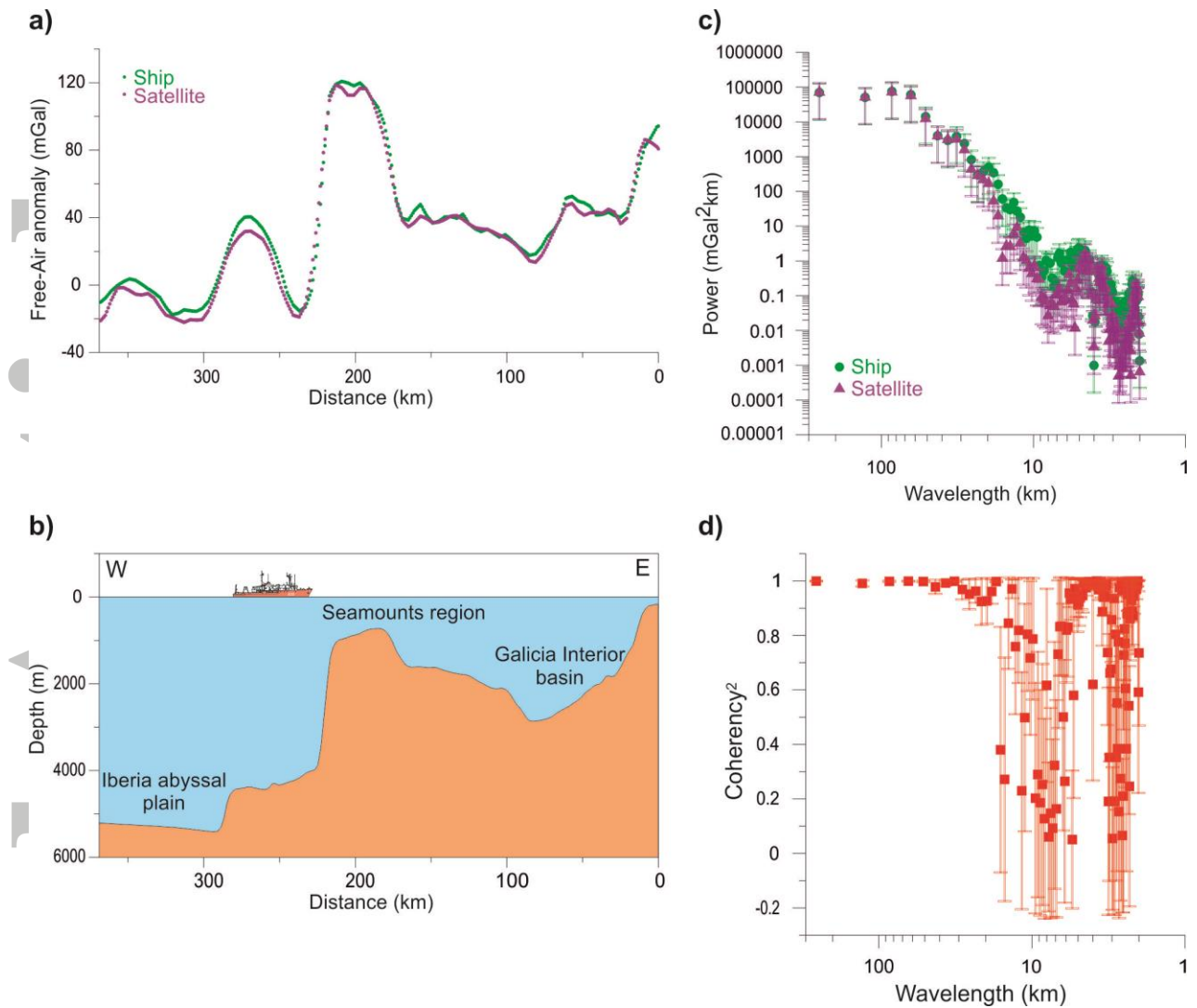
**Figure 2:** Shaded relief model from the DTM of the study area showing ship navigation lines with gravity (and swath bathymetry) data acquisition during the different surveys. On land, red points indicate the positions of the gravity measure stations. The black dashed line shows the location of the bathymetric and free-air anomaly profiling extracted for the coherency analysis (Fig. 4).

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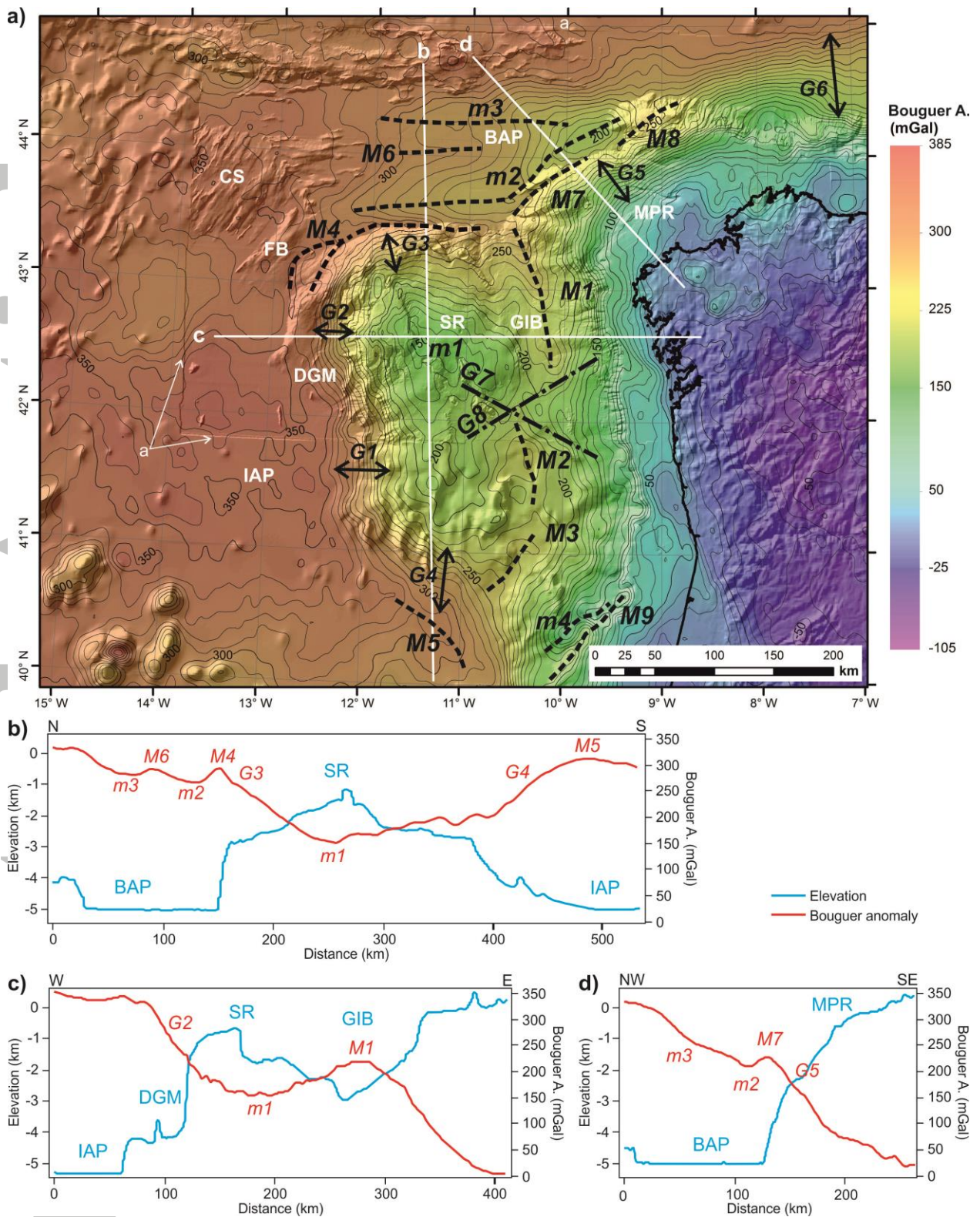
**Figure 3:** Digital terrain model of the study area with illumination from the NW. Contour each 500 m. Colored, swath bathymetry dataset (Spanish EEZ Project), surrounded by shaded relief model from GEMCO (offshore) and SRTM (onshore). Black thick lines show the location of the original multichannel reflection seismic profiles used for this work. Dotted black lines show the location of available refraction (wide angle) seismic profiles, published by others (see reference labels). Dash-dot black lines show the location of published reflection seismic profiles (see reference labels). Red thick lines show the locations of gravity models BAP, Biscay abyssal plain; CS, Coruña seamounts; FB, Finisterre bank; MPR, marginal platforms region; SR, seamounts region; DGM, deep Galicia margin; GIB, Galicia interior basin; IAP, Iberia abyssal plain.

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**Figure 4:** Coherency analysis of free-air gravity anomaly data from ship and satellite-derived datasets. Sampling interval is 1 km. See location of the sampled profile on Fig. 2. **a)** Ship (green) and satellite altimetry (purple) free-air anomaly values along the profile. **b)** Bathymetric profile overlapping the gravity sampled profile. **c)** Comparison between the ship and satellite altimetry data power spectrums calculated along the profile. **d)** Coherency analysis between the ship and satellite altimetry datasets.

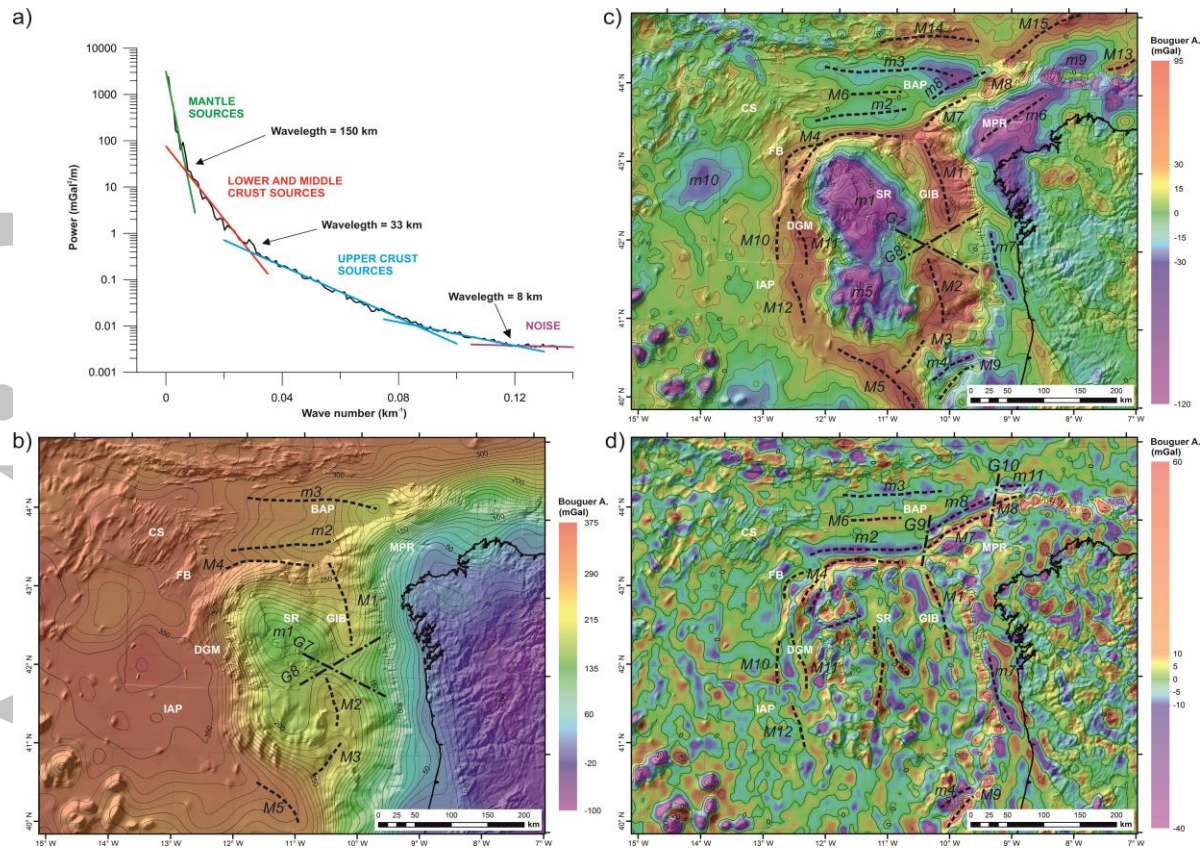
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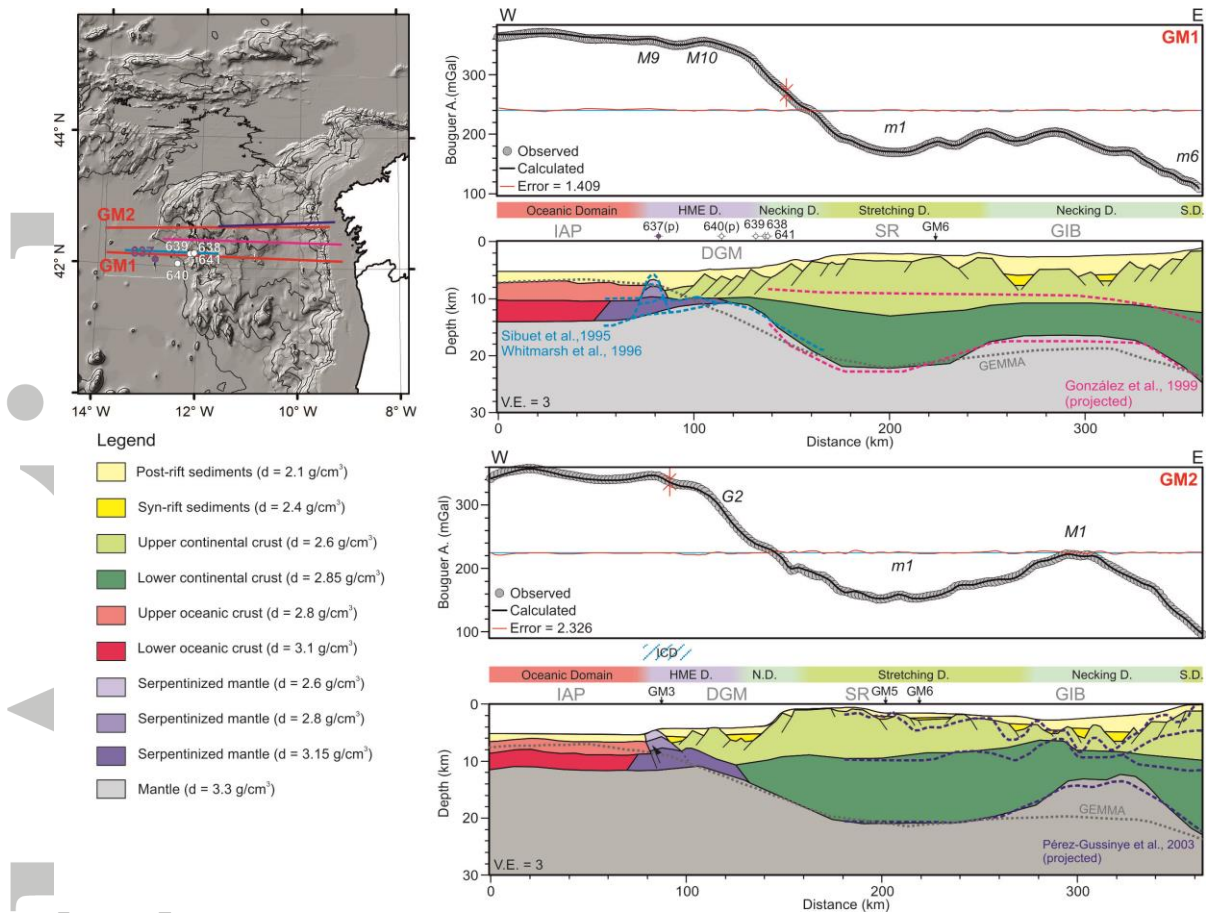
**Figure 5:** a) Bouguer anomaly map (color), presented over the shaded relief model, calculated from the DTM, to facilitate the connection between gravity anomalies and morphological features. White “a” indicates artifacts on the DTM by the ship and GEBCO assemblage. Contour each 10 mGal. Dashed black lines show relative minimum (*m*) and maximum (*M*) axis. *G* indicates gradient stripes on the continent-ocean transition (black arrows) or related to strike-slip tectonic features (dash-dot black lines). b-d) Elevation (blue) and Bouguer anomaly (red) profiles. Blue labels indicate toponymy, and red italic labels

indicate minimums, maximums and gradients seen also on **a**. BAP, Biscay abyssal plain; SR, Seamounts region; DGM, deep Galicia margin; GIB, Galicia interior basin; IAP, Iberia abyssal plain; MPR, marginal platforms region.

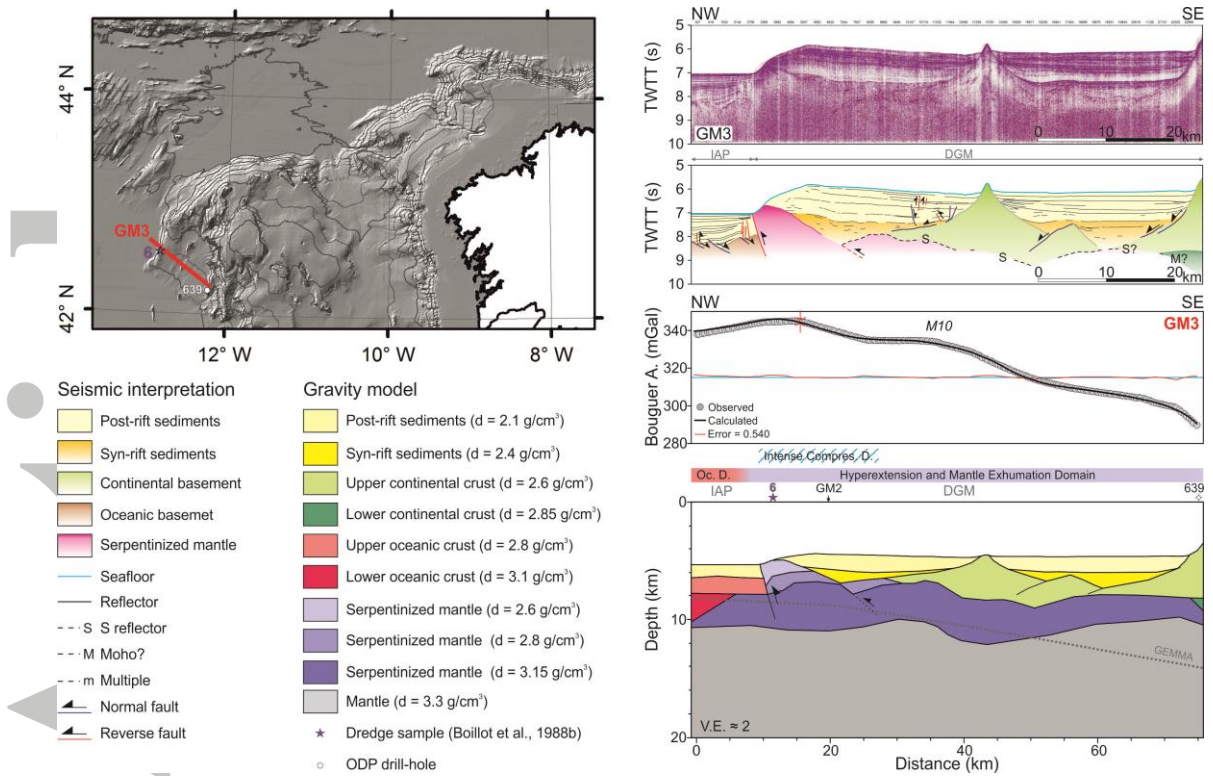
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**Figure 6:** a) Radial power spectrum of the complete Bouguer anomaly grid. Colored lines show the lineal approximations for every wavelength span related to each kind of gravity anomaly sources [Karner and Watts, 1983]. b) Long-wavelength Bouguer anomaly map ( $\lambda \geq 145$  km). Contour each 20 mGal. c) Mid-wavelength Bouguer anomaly map ( $145 \geq \lambda \geq 35$  km). Contour each 10 mGal. d) Short-wavelength Bouguer anomaly grid ( $35 \text{ km} > \lambda > 8$  km). Contour interval of 10 mGal. Filtered Bouguer anomaly grids are presented over the shaded relief model, to facilitate the connection between gravity anomalies and morphological features. Dashed black lines indicate Bouguer anomaly relative minimum ( $m$ ) and maximum ( $M$ ). Dash-dot black lines indicate gravity anomaly gradients ( $G$ ) related to strike-slip and/or transform faults.

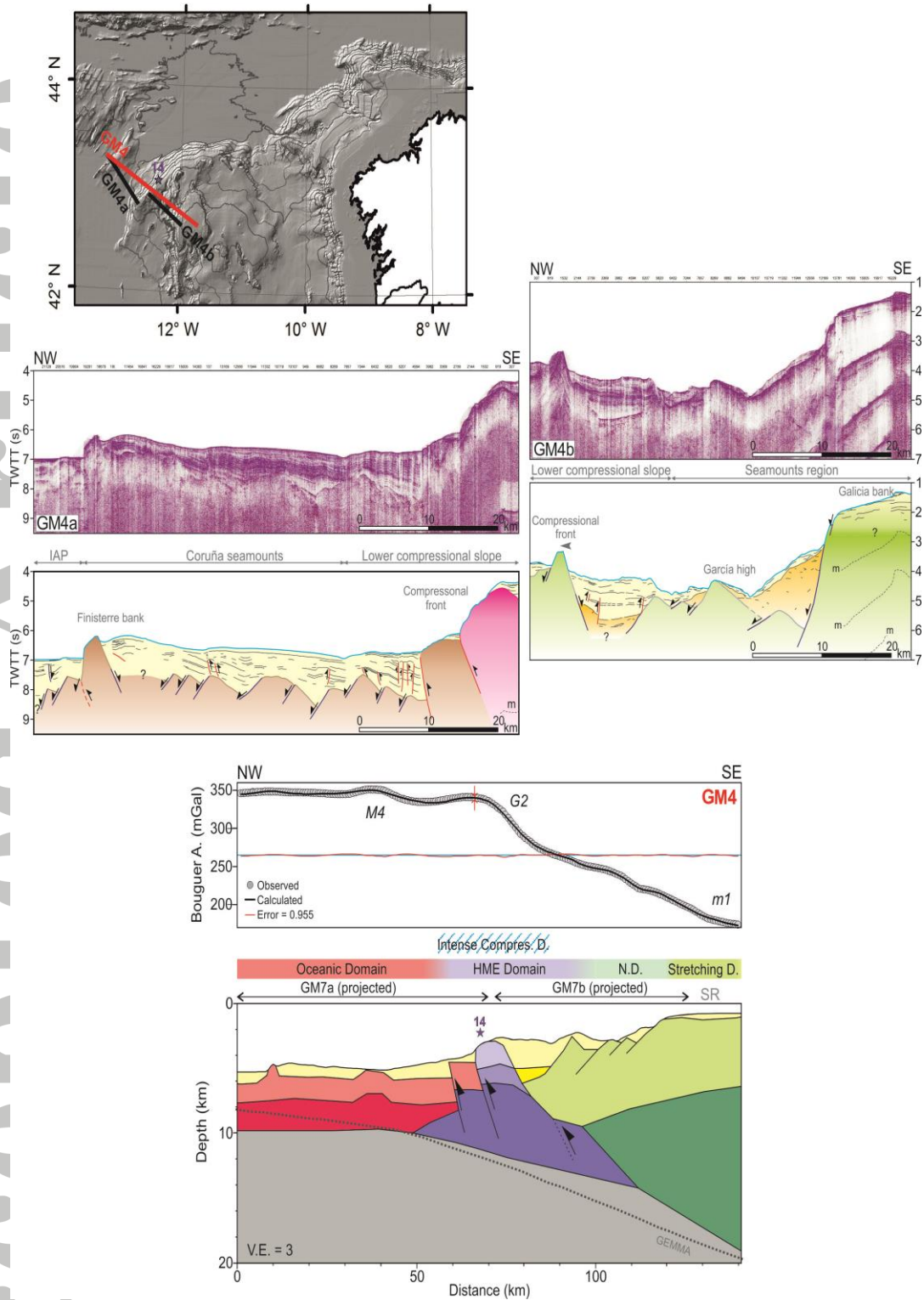


**Figure 7:** Gravity models GM1 and GM2, crossing the western margin. Italic labels on the Bouguer anomaly curves indicate relative maximums (*M*), minimums (*m*) and gradient (*G*) also labeled on the Bouguer anomaly map (Fig. 5a). Some horizons from published refraction seismic models are plotted (and referenced) over the gravity models (dashed colored lines, see location on the map and on Fig. 3). GEMMA Moho depth model profiles are also plotted on the gravity models (dashed grey line). ODP Leg 103 drill-holes locations are plotted on the insert and along GM1, p: location projected.

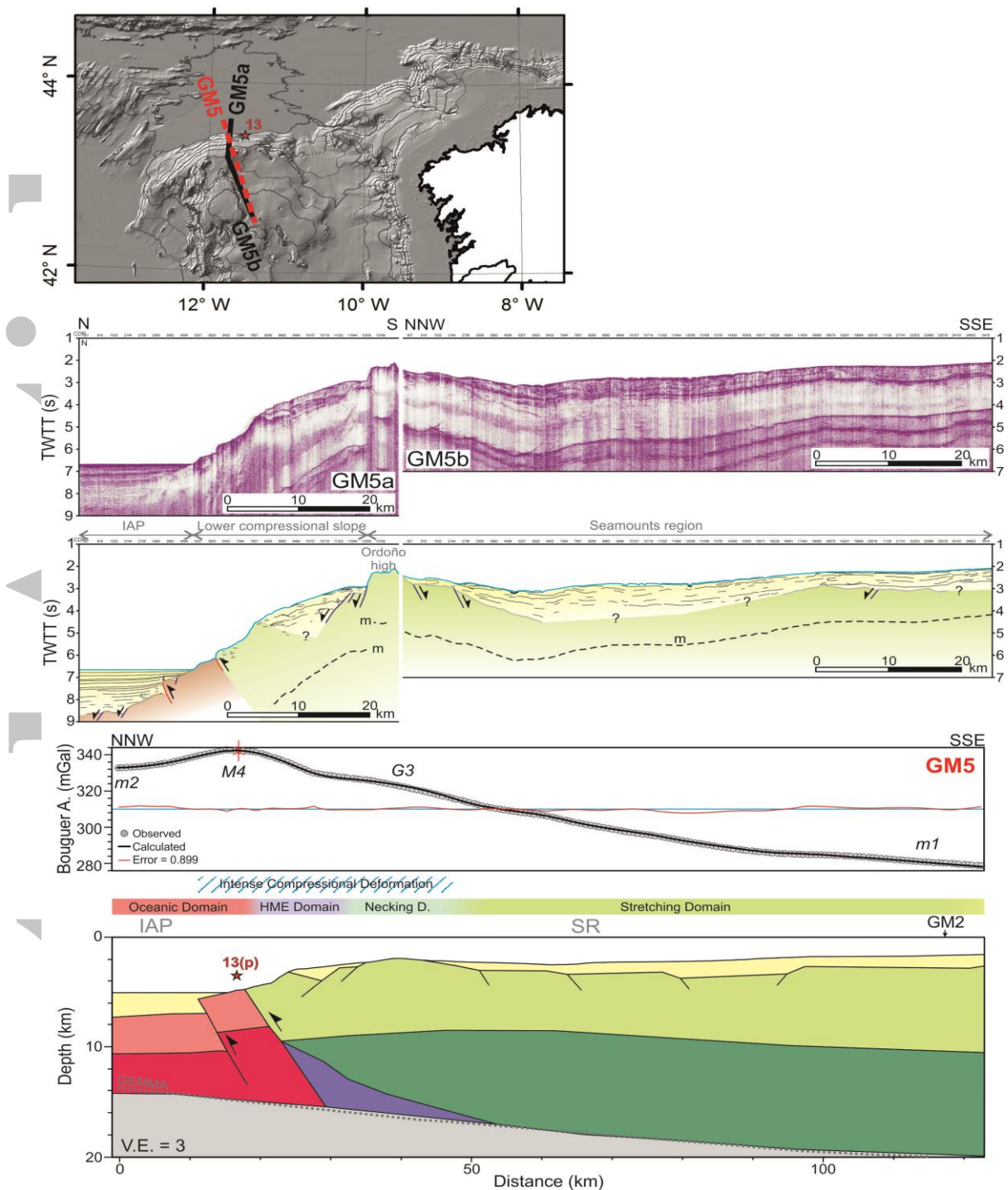


**Figure 8:** Coincident seismic profile and gravity model GM3. The seismic profile is shown with its line-drawing interpretation below. Dredge sample by *Boillot et al.* [1988b] and ODP Leg 103 drill-hole (white circle) are located on insert and on the interpreted seismic profile (purple star). GEMMA Moho depth model profile is plotted on the gravity model (dashed grey line).

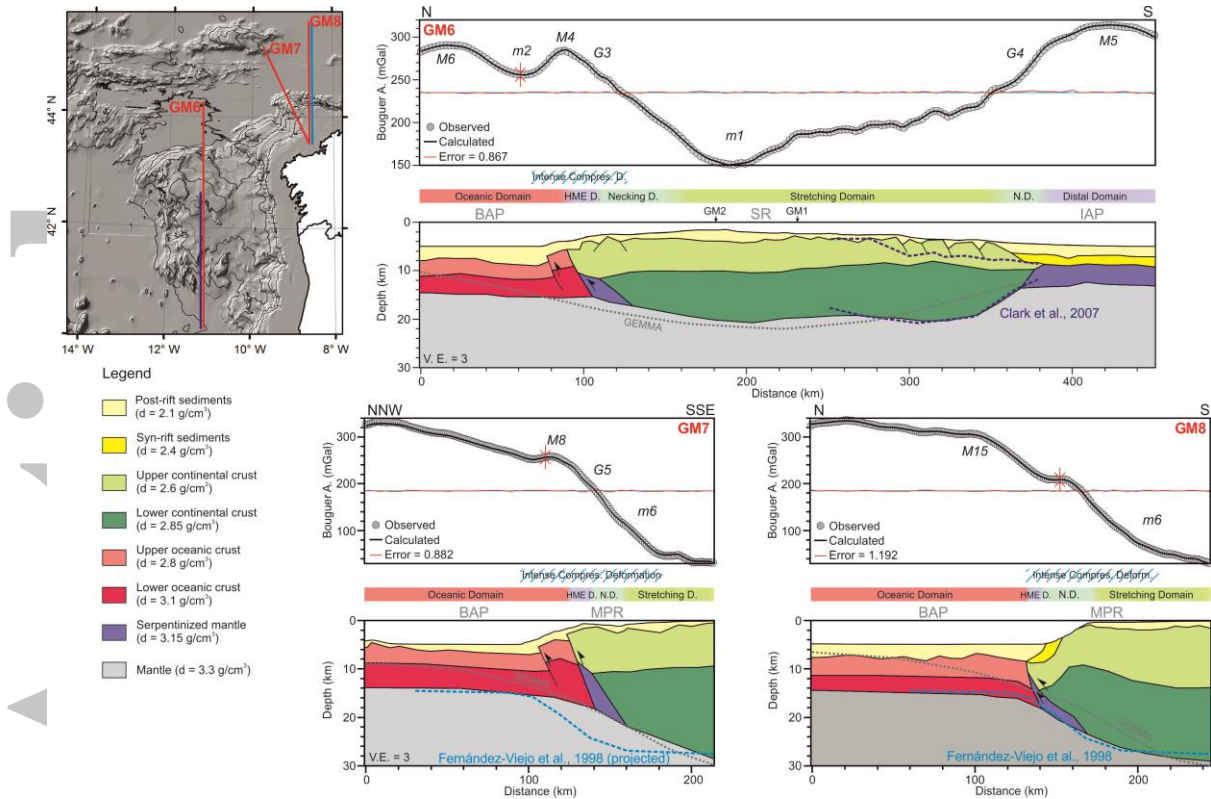
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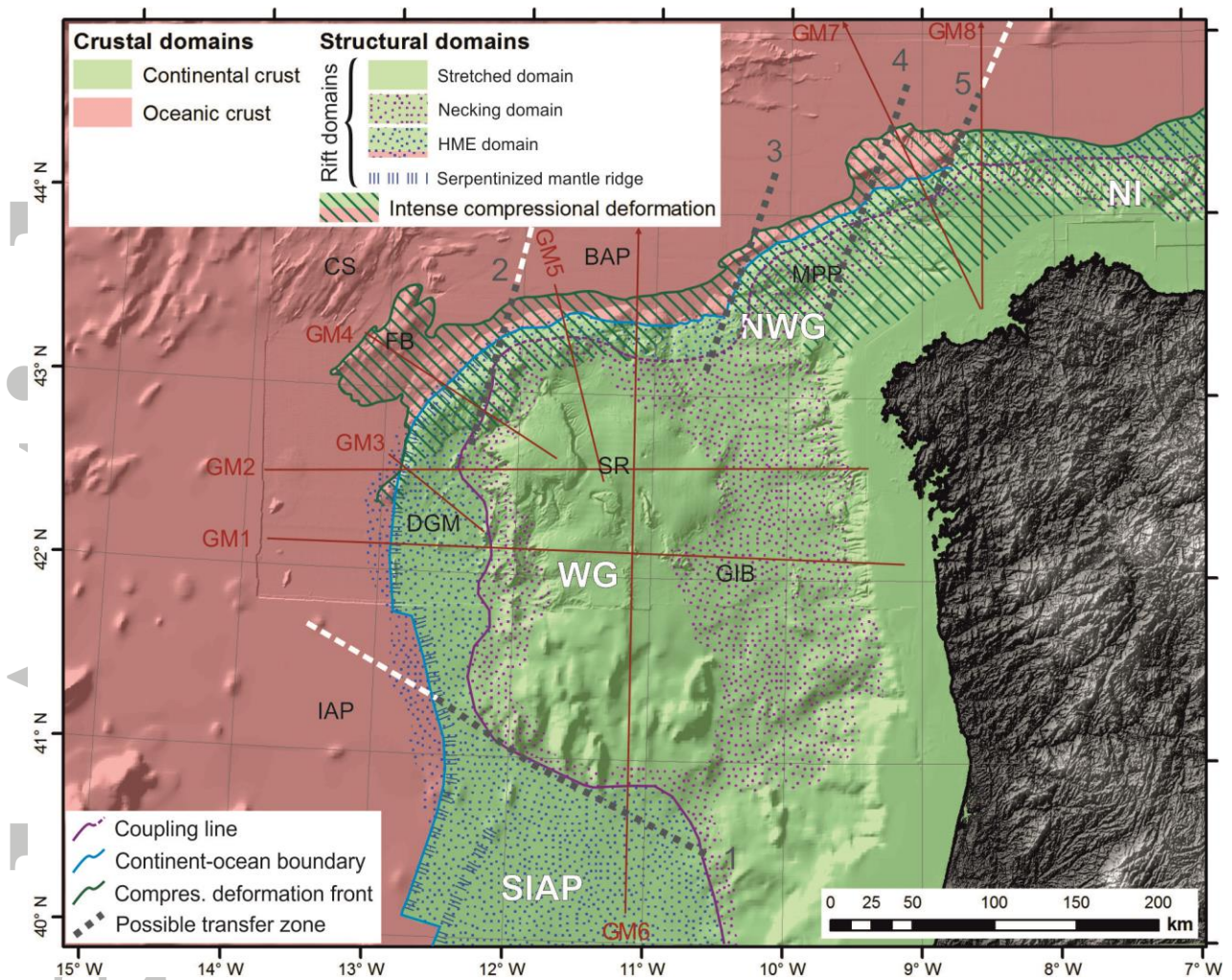
**Figure 9:** Seismic profiles GM4a and GM4b with their line-drawing interpretations below, and the nearby GM4 gravity model. In the location map, both the seismic profiles (black lines) and the gravity model (red line) are shown. See legend for seismic interpretation and gravity model on Fig. 8. Dredge sample by *Boillot et al.* [1988b] located on the insert and on the interpreted seismic profile (purple star). GEMMA Moho depth model profile is plotted on the gravity model (dashed grey line). Italic labels on the Bouguer anomaly curve indicate relative maximum *M4* and gradient *G2* also labeled on the Bouguer anomaly map (Fig. 5a).



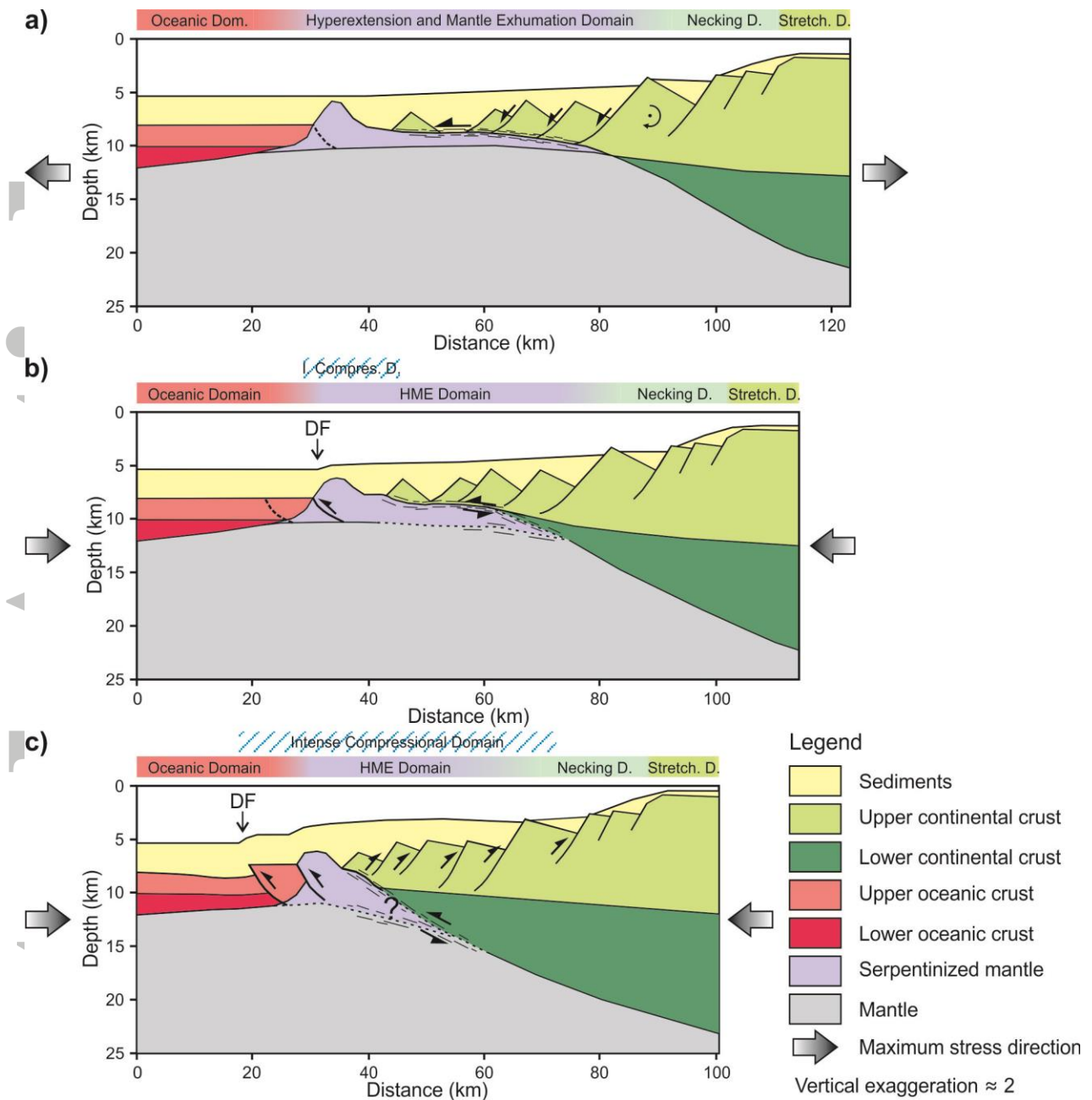
**Figure 10:** Seismic profiles GM5a and GM5b with their line-drawing interpretations below, and the nearly coincident gravity model GM5. In the location map, both the seismic profiles (black lines) and the gravity model (red dashed line) are shown. See legend for seismic interpretation and gravity model on Fig. 8. Dredge sample by *Boillot et al.* [1988b] located on the insert and on the interpreted seismic profile (red star). GEMMA Moho depth model profile is plotted on the gravity model (dashed grey line). Italic labels on the Bouguer anomaly curve indicate relative maximum *M4*, gradient *G3* and relative minimum *m1* also labeled on the Bouguer anomaly map (Fig. 5a).



**Figure 11:** Gravity models GM6 (north-south directed, crossing the whole margin from the Biscay abyssal plain to the southern Iberia abyssal plain), GM7 and GM8 (northwest of Galicia). Italic labels on the Bouguer anomaly curves indicate relative maximums (*M*), minimums (*m*) and gradients (*G*) also labeled on the Bouguer anomaly map (Fig. 5a). Some horizons from published refraction seismic models are plotted (and referenced) over the gravity models (dashed colored lines, see location on the map and on Fig. 3). GEMMA Moho depth model profiles are also plotted on the gravity models (dashed grey line).



**Figure 12:** Crustal and structural domains map (color), over shaded relief map from the DTM. CS, Coruña seamounts; BAP, Biscay abyssal plain; MPR, marginal platforms region; FB, Finisterre bank, SR, seamounts region; DGM, deep Galicia margin; GIB, Galicia Interior basin; IAP, Iberia abyssal plain. The coupling line indicates the border between coupled and decoupled crustal deformation. Continent-ocean boundary is interpreted from all the available data, including gravity, seismic, bathymetric and sampling information (See Figs. 1-6). The deformation front is the distal border of the area where intense compressional deformation is interpreted. The possible transfer zones are areas where gravity data, and also morphology, indicate the existence of lineal features that could play as transfer zones during rifting. Dashed white lines indicate approximate limits between the different segments of the margin: SIAP, south Iberia abyssal plain segment; WG, west Galicia segment; NWG, northwest Galicia segment; NI, north Iberia segment.



**Figure 13:** Simplified conceptual model for the initiation of the tectonic inversion of the north and northwest Galicia margin. **a)** Starting point, at the end of the rifting episode. **b)** Initiation of the tectonic inversion. Most of the compressional deformation is mainly assumed by the serpentinized mantle unit and by the lithospheric detachment, which is reversely reactivated. A thrust is originated on the distal flank of the peridotite ridge. **c)** Present stage of the tectonic inversion of the margin. When most of the serpentinized unit has been consumed by the tectonic inversion of the detachment, compressional deformation propagates towards the foreland by the generation of new thrusts affecting the proximal thin and weak oceanic crust. Tectonic compressional deformation progresses also towards the continent by the partial inversion of previous normal faults. DF, deformation front.