



The nature of crustal reflectivity at the southwest Iberian margin



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ABSTRACT

Reprocessing of multi-channel seismic reflection data acquired over the northern margin of the Gulf of Cádiz (SW Iberian margin) places new constraints on the upper crustal structure of the Guadalquivir-Portimão Bank. The data presented have been processed with optimized stacking and interval velocity models, a better approach to multiple attenuation, preserved amplitude information to derive the nature of seismic reflectivity, and accurate time-to-depth conversion after migration. The reprocessed data reveal a bright upper crustal reflector just underneath the Paleozoic basement that spatially coincides with the local positive free-air gravity high called the Gulf of Cádiz Gravity High. To investigate the nature of this reflector and to decipher whether it could be associated with pieces of mantle material emplaced at upper crustal levels, we calculated its reflection coefficient and compared it to a buried high-density ultramafic body (serpentinized peridotite) at the Gorringe Bank. Its reflection coefficient ratio with respect to the sea floor differs by only 4.6% with that calculated for the high-density ultramafic body of the Gorringe Bank, while it differs by 35.8% compared to a drilled Miocene limestone unconformity. This means that the Gulf of Cádiz reflector has a velocity and/or density contrast similar to the peridotite at the Gorringe Bank. However, considering the depth at which it is found (between 2.0 and 4.0 km) and the available geological information, it seems unlikely that the estimated shortening from the Oligocene to present is sufficient to emplace pieces of mantle material at these shallow levels. Therefore, and despite the similarity in its reflection coefficient with the peridotites of the Gorringe Bank, our preferred interpretation is that the upper crustal Gulf of Cádiz reflector represents the seismic response of high-density intracrustal magmatic intrusions that may partially contribute to the Gulf of Cádiz Gravity High.

1. Introduction

The Gulf of Cádiz is located at the eastern end of the Azores-Gibraltar Fracture Zone (AGFZ) at the complex plate boundary between the African and Eurasian plates. Its present day crustal and lithospheric structure is the result of a geodynamic history that started with the rupture of Pangaea and the subsequent opening of the Central Atlantic in the Early-to-Middle Jurassic (Schettino and Turco, 2009). This led to the development of conjugate passive margins and to the opening of the Ligurian-Alpine-Tethys Ocean (Jiménez-Munt et al., 2010). Thus, the crustal structure of the northern Gulf of Cádiz (SW Iberian margin) may enclose remnants of the Middle-Late Jurassic rifting stage (Sallarès et al., 2011). In places, we may expect some extensional features to be overprinted by compressional structures that started developing in the Late Upper Cretaceous (e.g., Vergés and Fernández, 2012) because of the convergence of Iberia and Africa. This convergence, which continues at present, is responsible for subduction in the Betic and Rif domains of the Ligurian-Alpine-Tethys with the associated emplacement of the Gulf of Cádiz Imbricated Wedge (GCIW) along its front, and

the formation of the Alboran back-arc basin behind it (e.g., Iribarren et al., 2009; Vergés and Fernández, 2012; and references therein).

In the study area, the Eurasian-African plate boundary is diffuse, unlike in its westernmost segment near the Azores, where it is narrow and well defined. The boundary can be traced from the Azores across the seafloor eastward of the Madeira-Tore Rise to the Gorringe Bank (GB), where its deformation extends over a broader region approaching the Gulf of Cádiz. At the Mid-Atlantic Ridge, the tectonic behavior is mainly transtensive, becoming strike-slip along the Gloria fault. In contrast, it is mainly transpressive at the Gulf of Cádiz (e.g. Zitellini et al., 2009; Jiménez-Munt et al., 2010) (Fig. 1).

Three abyssal plains surround the SW Iberian margin (the Tagus to the north and the Horseshoe and Seine to the south) (Fig. 1). They span the plate boundary and are underlain by thin, Late Jurassic-Early Cretaceous oceanic crust (Zitellini et al., 2009). Several prominent seamounts protrude from these abyssal basins, most notably the GB (comprised of the Gettysburg and Ormonde seamounts), which straddles the boundary of the African and Eurasian plates. The GB, mainly composed of peridotites, gabbros and extrusive rocks (Ryan et al., 1973;

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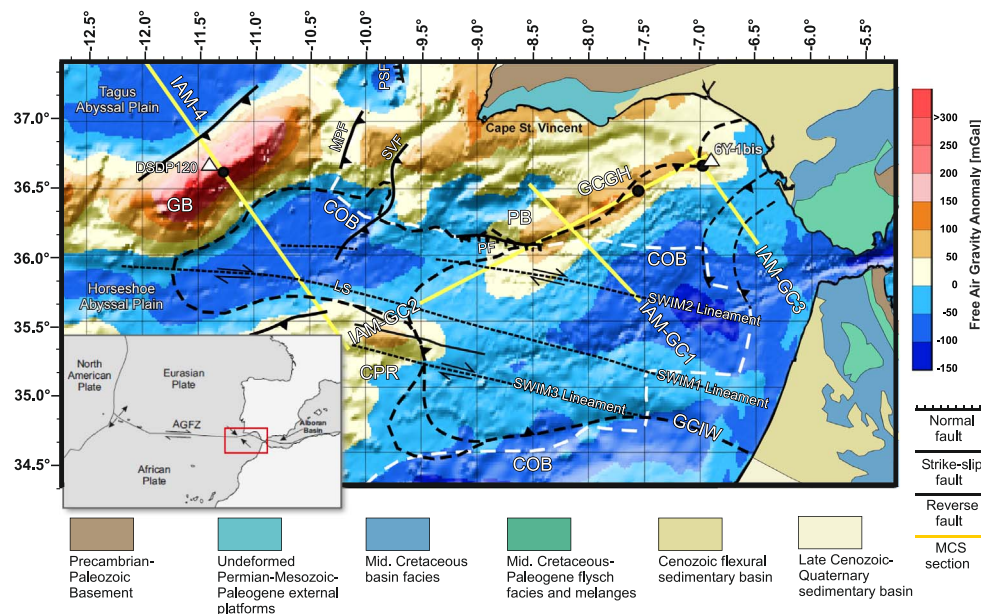


Fig. 1. Location of seismic lines analyzed (IAM-4) and reprocessed (IAM-GC1, GC2 and GC3) and their spatial relation to sea floor bathymetry and the Goringe Bank (GB) and Gulf of Cádiz Gravity highs (GCGH). Well sites are indicated by white triangles. Black dots on seismic profiles IAM-4, IAM-GC2 and IAM-GC3 show where reflection coefficients were calculated. Geological units modified from Mascle and Mascle (2012); tectonic domains and faults from Martínez-Loriente et al. (2014). Black dashed line indicates Gulf of Cádiz Imbricate Wedge; white dashed line indicates Continent-Ocean Boundary (COB); CPR: Coral Patch Ridge; GCGH: Gulf of Cádiz Gravity High; MPF: Marqués de Pombal; PB: Portimão Bank; PF: Portimão Fault; PSF: Pereira de Sousa Fault; SVF: St. Vicente Fault; AP: Abyssal Plain. Inset shows regional plate boundaries and relative movements. AGFZ: Azores-Gibraltar Fracture Zone. Tectonic background from Iribarren et al. (2007) and Zitellini et al. (2009).

Auzende et al., 1978; LaGabrielle and Auzende, 1982; Auzende et al., 1984) is one of Earth's largest positive free-air gravity anomalies, rising nearly 4500 m from abyssal depths to just 25 m below the sea floor (Fig. 1). Seismic reflection and refraction and integrated 2D lithospheric modeling (Purdy, 1975; Jiménez-Munt et al., 2010; Sallarès et al., 2013) show that the GB is a transtensional basin characterized by the presence of oceanic crust and exhumed serpentinized mantle material that was uplifted and thrust in a NW direction during the Late Oligocene-Early Miocene because of Africa-Eurasia convergence (e.g., Jiménez-Munt et al., 2010; Sallarès et al., 2013). Jiménez-Munt and co-authors estimate that the total shortening along the sub-crustal thrust fault is at least 20 km, shortening that is sufficient to explain the presence of mantle rocks at seafloor levels.

Three hundred kilometers directly east of the Goringe Bank, there is another large gravity anomaly oriented in a similar NE-SW direction (Fig. 1), the so called Gulf of Cádiz Gravity High (GCGH) located along the Guadalquivir-Portimão Bank (e.g., Roberts, 1970; Sandwell and Smith, 1997; Gràcia et al., 2003; Sandwell et al., 2014). This gravity anomaly also overlaps with a large magnetic anomaly (Dañoibeitia et al., 1999). Unlike the GB however, it lacks the same prominent seafloor morphology, being overlain by the thick GCIW. Gràcia et al. (2003) posits that this anomaly is a result of the combination of a shallow Paleozoic basement high and a local thinning of the crust, which is estimated by the authors to be no more than 10 km thick (see Fig. 10 of Gràcia et al., 2003). More recently, Ramos et al. (2017a) based on the interpretation of a 2D regional multichannel seismic survey acquired by the industry tied with onshore geology, exploratory wells, and 2D gravimetric modeling conclude that the GCGH is the result of the combination of a basement high, crustal thinning and the presence of exhumed sub-continental mantle material in the distal parts of the margin. In addition, Ramos et al. (2017a, 2017b) suggest that

Cenozoic shortening, which amounts to less than 10 km, is responsible for major uplift along the Guadalquivir Bank.

Mantle rocks at different crustal levels are found all along the Western Iberian Atlantic margin (WIAM) and along its northern boundary. Along the WIAM, the distal margin zone is formed by hyper-extended crust characterized by the presence of a strong reflector (the “S” reflector of e.g., Hoffmann and Reston, 1992) that marks the transition from fractured crustal rocks to partially serpentinized peridotites (Reston, 1996). Serpentinized peridotites are also found at the northern border of Iberia, along the Cantabrian continental margin and along the northern Pyrenees, although in the Pyrenees these ultra-mafic mantle rocks are emplaced at crustal levels (Torre et al., 1989; Torre et al., 2015 and references therein). The crustal emplacement of these mantle rocks occurred because of the convergence of Iberia and Africa that resulted in the inversion of the early Cretaceous rifted relief and rising of the mountain range mainly from 85 to 30–25 Ma (e.g., Vergés and Fernández, 2012). The transtensional regime and the slow lithospheric extension prevailing in the Middle-Late Jurassic in a corridor connecting the Central Atlantic and the Alpine Tethys (southern Iberia) resulted in the opening of small basins separated by transform faults, many of them characterized by the presence of exhumed mantle peridotites that were further serpentinized (e.g., Jiménez-Munt et al., 2010). In addition, since some local gravity highs located along northern Iberia and its western margin are related to the presence of intracrustal or buried peridotitic bodies and that a NE-SW oriented gravity high is also found in the Guadalquivir-Portimão Bank of the northern Gulf of Cádiz (SW Iberia margin), we reprocessed three regional multichannel seismic (MCS) profiles from the Iberian-Atlantic margin (IAM) survey (Banda et al., 1995), located along and across the Gulf of Cádiz Gravity High to study the reflectivity of the crust and its possible causes. Our main aim was to address if along the Guadalquivir-

Table 1
Acquisition parameters.

Shot point interval	75 m
Record length	25 s
Sample rate	4 ms
High cut field filter	90 Hz–72 dB/oct ramp
Low cut field filter	Out
Streamer type	HSSQ/GX600 analog
Number of geophone groups	192 (40 hydrophones per group)
Group length	25 m
Group interval	25 m
Active streamer length	4800 m
Nominal streamer depth	15 m
Source type	BOLT Airgun array
Source volume	0.123 m ³
Source depth	10 m (± 1 m)

Portimão Bank there was also evidence for the presence of peridotitic bodies thrust at crustal levels that could partially contribute to the gravity high. For that purpose, we generate and analyze true amplitude stacked MCS profiles IAM-GC1, GC2, and GC3 (Fig. 1) that run along and across the NE-SW oriented GCGH to a depth of 12 s two-way travel time, or about 9500 m.

Reflection coefficients were also extracted from raw field data records for a prominent reflector located in the Guadalquivir-Portimão Bank that had been previously interpreted as the top of the Paleozoic basement (Gràcia et al., 2003), and which the new processing approach reveals to be a separate reflection event. Its reflection coefficient is compared with that of the ultramafic rock units of the Gorringe Bank (GB) (IAM-4 MCS profile, see Fig. 1 for location).

The newly applied processing flow improves the images of previous interpretations (Tortella et al., 1997; Gràcia et al., 2003), by more precisely defining the structure of the upper-middle crust along the Guadalquivir-Portimão Bank. These previous studies analyzed the same seismic data reported here. However, both interpretations relied on the post-cruise processing sequence carried out by the acquisition company Geco-Prakla over 20 years ago. The presented processing flow improves multiple noise attenuation and estimation of seismic velocity models that have allowed us to preserve the amplitude information—avoiding conventional automatic gain control (AGC); see Section 3.2. The trace amplitude variations across acoustic impedance (density–velocity) interfaces are then analyzed by calculating instantaneous amplitude and reflection coefficients. When the amplitude relationship is preserved, instantaneous amplitude can be regarded as a measure of reflection strength (Yilmaz, 2001).

2. Data and methods

2.1. Seismic data acquisition

The data used were acquired during the IAM survey conducted from August to September 1993. In this contribution, we reprocess and analyze the seismic lines IAM-GC1, IAM-GC2, IAM-GC3, and IAM-4 (Fig. 1). IAM-GC1 and IAM-GC3 cross the GCGH approximately orthogonally. IAM-GC2 bisects the gravity anomaly along its length. IAM-4 crosses the center of the GB. The seismic data were acquired by Geco-Prakla onboard the vessel Geco Sigma. Summarized acquisition

parameters for the IAM survey are found in Table 1. Complete details on survey parameters can be found in Banda et al. (1995).

2.2. Seismic data processing

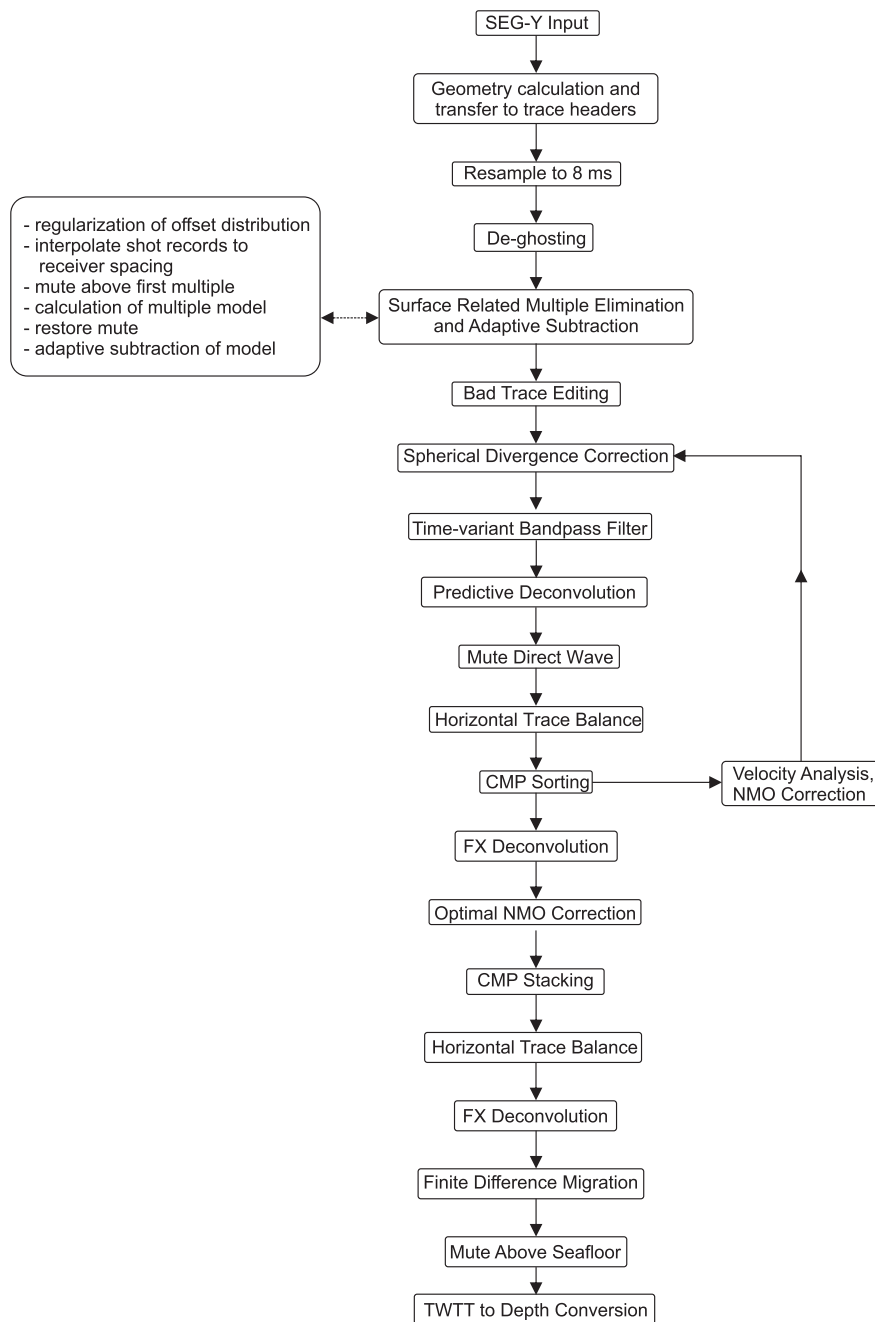
Shot records were first sorted, resampled, and geometry was applied using the onboard GPS navigation data. A small number of noisy channels were zeroed from each shot to optimize signal-to-noise ratio. Strong sea floor multiple energy was suppressed using a Surface Related Multiple Elimination (SRME) algorithm (Verschuur et al., 1992). To apply this method, we first needed to ensure that there were no source and receiver ghosting issues, which would affect the source wavelet shape. We did this by calculating the source and receiver delay times that correspond to the tow depths of the source and streamer. Knowing this delay time, we shaped the source wavelet by applying a band-pass filter to exclude noise above the ghost notch. Next, we regularized the offset distribution by choosing near and far offset values, as well as a fixed offset increment. This is done to regularize the natural variability of offsets due to streamer feathering, variations in tow depth and cable stretching. Next, SRME requires that we interpolate each shot to every receiver location. To do this, we identified the first seafloor multiple and temporarily muted above it in order to allow SRME to properly calculate the multiple model. After the multiple model was created, we restored the mute, uninterpolated the offsets and restored the original offset distribution. The approach is completed after adaptive subtraction (Wang, 2003) of the multiple model. This is one of the innovative processing steps applied that help preserve the amplitude information and increase the lateral resolution.

As the aim is to assess the nature of the reflectivity, reliable amplitude corrections were carefully designed. The processing sequence included a spherical divergence correction using the most accurate root-mean square (RMS) velocity model derived from velocity analysis. It was configured to compensate for geometrical spreading as well as for physical absorption using an average Q factor of 300 for the upper 5 km of crust with a high Q below (Warner, 1990). Predictive deconvolution was used to compress the seismic wavelet and to help suppress residual multiple energy. We applied a band pass filter of 4/6–50/60 Hz, which avoided oceanic swell noise at the low end and the Nyquist resampling frequency limit at the high end. We then muted the direct wave and applied a normalized lateral trace equalization. The latter specifically aims to preserve relative amplitude relationships assuming similar coupling for all hydrophones. Importantly, this process does not scale amplitudes in the vertical, along each seismic trace. The data traces were then sorted by common midpoint (CMP).

The CMP gathers were then corrected for normal moveout after several iterations of velocity analysis. A surface consistent fx-deconvolution was applied before the CMP gathers were stacked. The horizontal trace balance and fx-deconvolution operators improved lateral correlation and attenuated random noise.

Finally, a finite-difference time migration algorithm was applied using interval velocities (see Supplementary material for the interval velocity function table) derived from the normal moveout (NMO) velocity field (Dix, 1955). The model was muted above the seabed. The migrated image was time-to-depth converted using a smoothed interval velocity function (Fig. 2).

Fig. 2. Processing scheme employed in this study. See Section 2.2 for details.



2.3. Calculation of reflection coefficients

The amplitude of seismic waves decreases as a function of the distance traveled. This is due to geometric spreading, physical attenuation through anelasticity (Jeffreys, 1965; Liu et al., 1976), scattering, mode conversion, and heat loss (Anderson and Hart, 1978). To characterize an interface by computing its reflection coefficients, seismic data need to compensate for this energy loss. In the current case, we also minimize for different source-receiver distances, and hence different travel paths so that the vertical incidence hypothesis would hold. We analyzed the nearest common offset receiver group (633.7 m for the Gorrige Bank and 631.0 m for the Gulf of Cádiz Gravity High), which were the nearest-to-vertical ray paths.

With the above considerations, geometrical divergence compensation and true amplitude recovery can be used to faithfully restore

modified amplitudes to their (most approximate) true amplitude had there been no loss of energy within the overlying layers. To minimize the amount of error bias in true amplitude recovery, we chose a location with a thin sediment cover and minimal dip. Therefore, most of the loss of amplitude observed was due to geometrical spreading in the water column, which does not significantly attenuate compressional seismic waves. Moreover, the sediment over the reflector in question is composed of unconsolidated, water-saturated sediments with well-constrained seismic velocities.

Since amplitudes are a function of the acquisition system, source signal, the path taken to the reflector and back, as well as the physical properties of the overlying media and the reflecting interface, MCS does not generally preserve the absolute amplitudes of reflectors. This is primarily due to the stacking of different data offsets from imperfectly corrected hyperbolic moveout. Therefore, while we herein present true

relative amplitude seismic profiles, we additionally analyze absolute amplitudes on common-offset sorted gathers to calculate and compare reflection coefficients, with the aim of estimating reflector density contrasts. We base our calculations on the method of Warner (1990), as detailed below.

We calculated R for regions of each seismic profile at two depths: at the sea floor (used as reference) and at the reflector in question. That is,

$$R = \frac{dA}{k} = \frac{\rho_2 v_2 - \rho_1 v_1}{\rho_2 v_2 + \rho_1 v_1} \quad (1)$$

where ρ and v represent the densities and velocities of the overlying and underlying media (1 and 2 respectively), A is the average amplitude of a given reflector, and k is an acquisition calibration factor. We determined the geometrical divergence compensation factor, d as:

$$d = 1/G(t) \quad (2)$$

where $G(t)$ is a time-dependent function given by:

$$G(t) = V(t)^v \cdot t^T \cdot e^{V(t) \cdot t \cdot \alpha} \cdot x^X \quad (3)$$

where v is the power value of the average velocity at t (a given two-way time), T is the power value of the two-way time, and X is an offset parameter used to calculate trace-by-trace scalars. The exponent α is a physical absorption coefficient (Sheriff, 1991). To simplify the calculation, we used a value of unity for the exponents v and T , because for the case of propagation in the water column, velocity is only variable by about 2% (Fortin and Holbrook, 2009). In addition, we observed velocities in the near surface sediments to increase by only 7% over water velocity. The offset term also reduces to unity because we have constrained our analysis to a single offset. The exponent α relates to seismic Q as:

$$\alpha = \frac{\pi f}{QV} \quad (4)$$

where f is the dominant frequency, V is the average wave propagation velocity, and Q is an attenuation factor that is the ratio of 2π times the peak energy to the energy dissipated in a cycle (Sheriff, 1991). Warner's method computes reflection coefficients for deep crustal reflectors by first computing k . This calibration factor is a function of the acquisition configuration (source strength and wavelet shape, streamer geometry, et cetera), and spherical divergence correction parameters that must include a value of the bulk seismic Q of the crust. In Warner's study, k is the greatest source of uncertainty to estimate deep crustal reflection coefficients because of the passage of the source wavelet through tens of kilometers of rocks with unknown properties. The calibration factor therefore adjusts recorded amplitudes by compensating for the unconstrained filtering effects of the Earth. We calibrated the acquisition system by considering a simplified case of how seismic amplitudes vary in a minimally dissipative scenario, appropriate for the near-surface reflectors in this study. We compared the amplitude of a primary sea-floor reflection and its first multiple reflection, which should occur at twice the two-way travel time. Warner (1990) gives the amplitude, A_p of a primary reflection as:

$$A_p = \frac{k R_{sf}}{d} \quad (5)$$

where R_{sf} is the reflection coefficient of the sea floor. Similarly, the amplitude of the seafloor's first multiple, A_m is given by:

$$A_m = \frac{R_{sf}^2 k}{2d} \quad (6)$$

A_p and A_m are measured directly from the seismic traces. R_{sf} is found to be $2A_m/A_p$ by substitution of Eqs. (5) and (6). Knowing R_{sf} and having separately determined d , we therefore define the calibration factor as:

$$k = \frac{dA_p}{R_{sf}} \quad (7)$$

Which gives us a value of $k = 1.6$ and with $\sigma = 0.10$ when averaged across each trace of the seafloor reflector in the common offset gather.

In contrast to Warner's deep crustal study, the present study is simplified by analyzing: a) near-surface structures with thin layers of overburden, b) near-vertical incidence rays, c) flat reflectors to minimize losses due to mode conversion that can occur at oblique incidence angles, and d) a low dominant frequency (15.7 Hz) to minimize attenuation loss via absorption. To be certain about the influence of Q values on reflection coefficient, we tested extreme values and found a negligible change in peak amplitude, as expected for a minimally dissipative scenario. Within these constraints then, the main effect of spherical divergence compensation is to adjust amplitudes for the loss due to geometric wave front spreading, which varies as a function of distance traveled to the reflector and back. The spherical divergence relationship thus becomes:

$$G(t) = V(t) \cdot t \quad (8)$$

allowing us to then compute the value of R using Eq. (1) (see Supplementary material for the Excel file used to calculate the calibration factor and reflection coefficients).

2.4. Sources of error

The main limitations of any MCS survey can be divided into those related to: i) estimation of the source wavelet signature, ii) the ray path taken (angle-dependent amplitude variation), and iii) the geometry and physical properties of the subsurface (convolutional model assumptions). In addition, there are processing-related sources of error such as non-optimal alignment of hyperbolic moveout during velocity analysis, the loss of amplitude information because of stacking, and the imperfect restoration of reflectors by migration.

The source variables are related to its configuration (amplitude, frequency band, and phase), which although constrained here within the same survey parameters, may vary from shot-to-shot because of the natural mechanical variability of the air guns, the compressor capacity (gun volume, initial firing pressure, port-closure pressure, et cetera), gun synchronization, and the geometry of the source array (Dragoset, 1990).

Secondarily, for a given seismic air gun source there is a spectrum of ray paths that spread out from the source. While the relative amplitude can be preserved even after stacking and migration, in stacked profiles the absolute quantitative meaning of the reflection signal is lost. Stacking uses data from many offsets to be combined to maximize the signal-to-noise ratio. Furthermore, P-S mode-conversions carry away some of the incident energy at larger incidence angles. Thus, to minimize energy loss from mode conversion at interfaces, we chose the nearest practical common offset for both seismic profiles (see Supplementary material for a segy file and jpg image of a representative raw shot record). Due to the geometry of each line, these values differed by 2.7 m–633.7 m for the IAM-4 (GB) and 631.0 m for IAM-GC2 and IAM-GC3 (GCGH).

Finally, there are unknown amplitude losses due to the geometry of the interfaces and the heterogeneous distribution of the reflection and transmission properties. These are the least constrained variables. A priori, we cannot measure the expected internal changes in amplitude of a wave front as it passes through the Earth using MCS data alone. Like the majority of MCS studies, we rely on the assumptions of the convolutional Earth model (Yilmaz, 2001). Changes to the wavelet occur due to geometric spreading, focusing and defocusing due to lateral velocity variations, anelastic attenuation and elastic scattering from simple specular reflectors, unknown random inhomogeneities (Warner, 1990), and from diffraction and wave front healing at sharp boundaries. True amplitude recovery for spherical divergence

spreading loss is only possible if the subsurface properties are constrained, including the depth and frequency dependence of natural absorption coefficients (Eq. (4)) (Lekić et al., 2009). Specifically, focusing/defocusing effects were present in this study because the GB and GCGH reflectors had different dips (3 ms/trace and 10 ms/trace, respectively) that, in principle scatter energy at differently oblique angles. When faced with the dilemma of minimizing all error criteria, we chose to control for reflectors with large dips, in order to mitigate the focusing/defocusing issues.

2.5. Instantaneous attributes

To quantify reflectivity over a broad area, we calculated the instantaneous amplitude of the final migrated stacked profile. Instantaneous amplitude is a measure of reflectivity strength, which is proportional to the square root of the total energy of the seismic signal at an instant of time (Bracewell, 1965; Taner et al., 1979). For instantaneous attributes to be physically meaningful, the CMP stacked profile must closely represent the subsurface in a way that the amplitude and frequency content are preserved, and there is minimal change in the shape of the waveform. In addition to preserving the waveform, multiples and random noise must also be minimized (Yilmaz, 2001).

Reflectivity strength is a proxy for reflection coefficient (R), which at normal incidence is a measure of the contrast in acoustic impedance across a boundary (Eq. (1)). A reflection coefficient of unity means perfect specular reflection, while zero indicates perfect transmission. In practice, neither is fully obtainable due to loss of signal through

geometric spreading and physical attenuation. However, sensible choices for the parameters help to preserve the relative amplitude information. The drawback of true amplitude preservation is that deeper, weaker structures are not generally visible. This is because at depth they are inherently less reflective due to the increasing loss of signal amplitude as it passes through the Earth. In addition, density and velocity contrasts decrease with depth due to an overall increase in compression, making only the largest acoustic impedance contrasts (e.g. the Moho) visible. The reader is referred to Tortella et al. (1997) for identification of deep structures on this data set.

3. Results and discussion

3.1. A brief description of the seismic profiles

All three seismic profiles (Figs. 3 to 5) show a 500–1000 m layer of sub-horizontal reflectors. A full description of the sedimentary sequence of the northern Gulf of Cádiz can be found in the works of Tortella et al. (1997); Gràcia et al. (2003) or Ramos et al. (2017a), among others. In general, the sedimentary layers are overlain on higher amplitude basement rocks with variable dips. These layers extend to approximately three kilometers below the sea floor, becoming less vertically and horizontally coherent and likewise losing reflection strength.

From NW to SE, IAM-GC1 (Fig. 3) crosses Portimão Canyon, near Cabo de São Vicente until the latitude of the Strait of Gibraltar. It is characterized by highly variable surface morphology that divides units of no more than 500 m thick, which are likely composed of water-

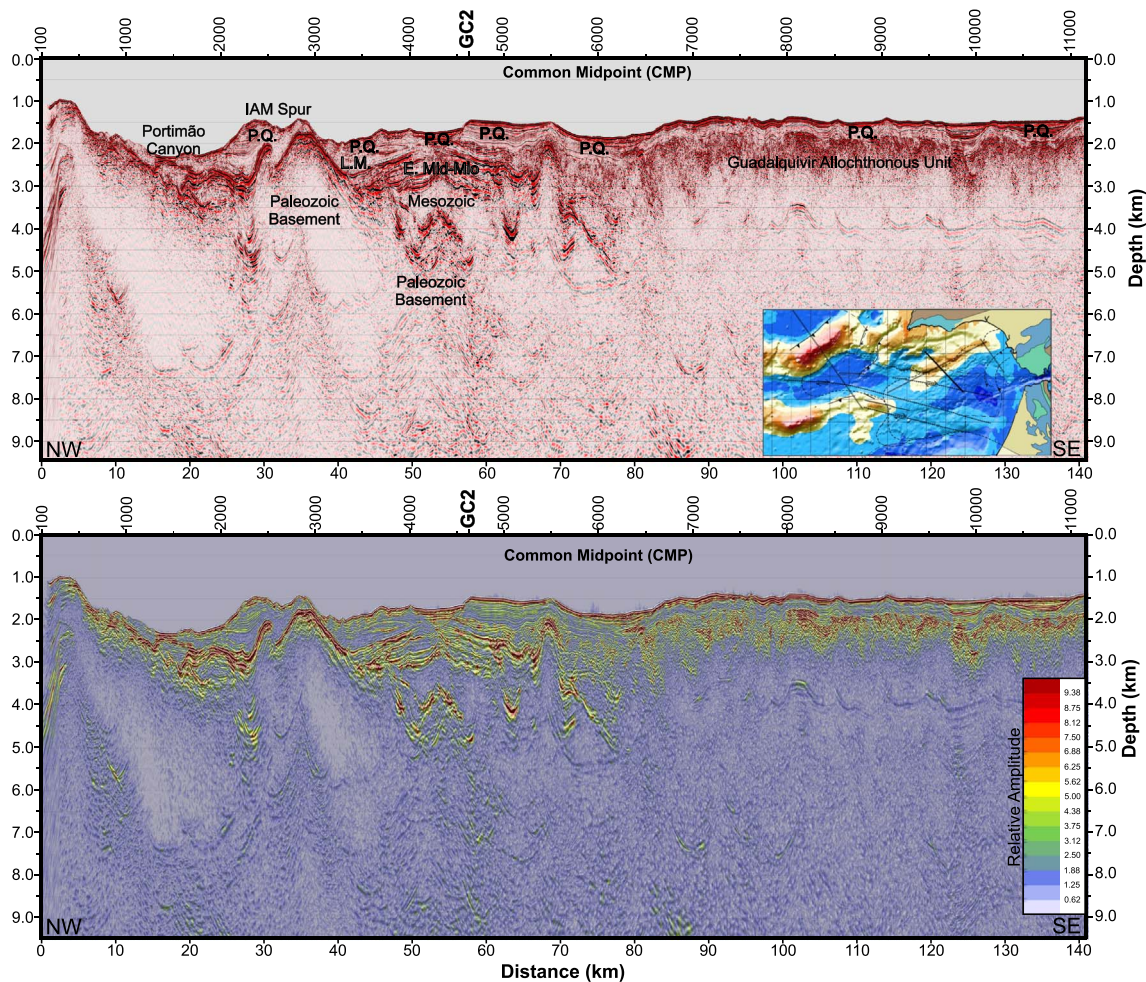


Fig. 3. MCS profile IAM-GC1. Top - final migrated depth profile. Black line in bottom right inset shows its location. Bottom - instantaneous amplitude attribute overlain on seismic data.

saturated unconsolidated Plio-Quaternary sediments. To the SE there are two distinct types of reflectors with different wavelet characteristics. Near the surface, thin, horizontally coherent, but weak reflectors represent relatively undeformed Plio-Quaternary sediments. These overlie thicker, higher amplitude, undulating Middle Miocene-to-Mesozoic units. Between CMP 4000 and 4500, the Paleozoic basement rocks and overlying Miocene and Mesozoic units are uplifted and symmetrically faulted. At the apex of this uplift there is a separated sedimentary basin bounded by normal faults that is characterized by horizontally coherent stratigraphy. At about CMP 5500 there is a distinct Triassic salt diapir, identified by Gràcia et al. (2003). Above the diapir, there is notable deformation of the Plio-Quaternary (and a thinner, less coherent layer—possibly Late Miocene) sediments. This indicates that the diapir continued its growth after the deposition of these sediments and possibly continues at present. The high amplitude reflectivity at the top of the diapir points to a sharp impedance contrast, supporting the notion that it is composed of salt. Moreover, the interior of the diapir is seismically transparent, suggesting that it is monocompositional.

Directly SE of the salt diapir, there is a dipping coherent reflector, which ends at about CMP 6300. From here until the end of the profile, the Plio-Quaternary sediments are more horizontal and thinner than 500 m; they are locally disrupted by some diapiric structures. Below these sediments (between about 2 and 3 km depth) there is a wide region of high reflectivity/low-horizontal coherency, which diminishes completely beneath 3.5 km. Below this depth the upper crust is nearly transparent when displayed with the true relative amplitudes.

Seismic profile IAM-GC2 (Fig. 4) coincides with the major axis of the gravity high (Fig. 1) extending from the NE to the SW. The sea floor

at the NE end of the profile is less than 500 m deep, while the SW extent reaches depths of up to 4000 m. A thick (1.5 km) Plio-Quaternary/Miocene sedimentary sequence is observed in the NE end of the profile (from CMP 100 to CMP 3000). This is part of the GCIW, which thins considerably until about CMP 4800, thickens again, then becomes very thin and almost outcrops in places. It bounds isolated basins that contain some normal faults (e.g. near CMP 7000). At the intersection with IAM-GC1 there is a large sedimentary unit that varies in thickness between about 300–1000 m. It thins to less than 500 m for the remainder of the profile. From CMP 13000 until CMP 16200, the near-surface sediments overlie about 500 m of units that display broken reflectivity; presumably Mesozoic units with tectonically altered fabric (Tortella et al., 1997). Here, there is an abrupt truncation between these units and the underlying ones, where they are in contact with a chaotic body with low horizontal coherency. We identify this as the Giant Chaotic Body outlined in Torelli et al. (1997) and Tortella et al. (1997).

From CMP 9000 until about CMP 12500 there are several clear reflectors. These may be Mesozoic units on top of Paleozoic basement rocks. As the seafloor descends from CMP 12000 towards the SW, we can clearly see two to three distinct units under the Plio-Quaternary (and possibly, Miocene) sediments. Here, part of the Giant Chaotic Body forms a wedge-like shape between tectonically altered Miocene fabric and Paleozoic basement reflectors. These reflectors dip gently at about 3° towards the SW until CMP 16400 where there is a major fault with a dip of approximately 30° (Fig. 4). Horizontal coherency is greatly diminished here but the sequence can still be followed until the end of the profile in a slightly up-dip direction. This normal fault appears to be at or near the Continent-Ocean Boundary (Fig. 1).

Profile IAM-GC3 is 85 km long and is approximately parallel to IAM-

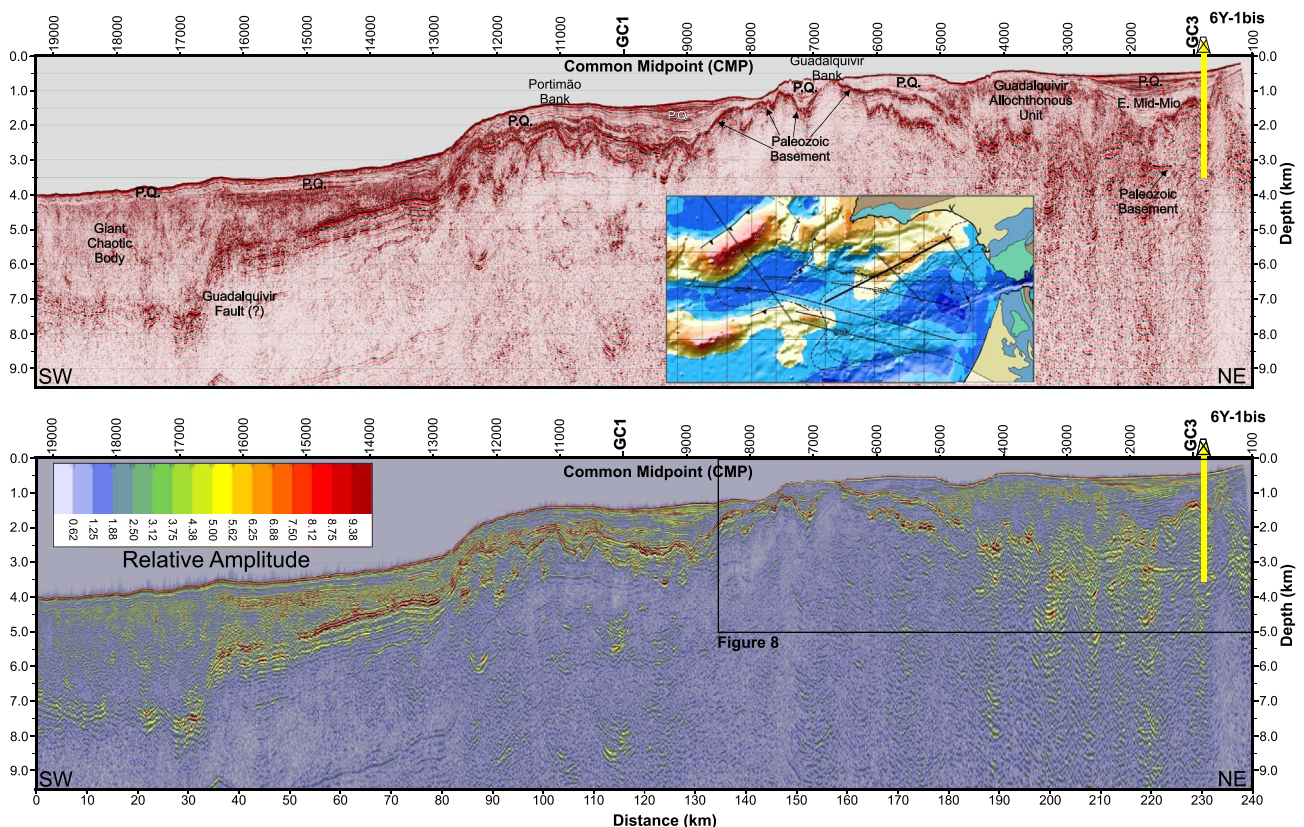


Fig. 4. MCS profile IAM-GC2. Top - final migrated depth profile. Black line in bottom right inset shows its location. Bottom - instantaneous amplitude attribute overlay on seismic data. Borehole location indicated in yellow. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

GC1, but situated higher on the accretionary wedge, close to the Guadalquivir basin. It has a shallow, roughly horizontal planar sea floor. It shows thicker near-surface Plio-Quaternary sediments than IAM-GC1, with thicknesses increasing towards the SE. In the NW, at the intersection with IAM-GC2, the lower boundary of the sediments overlies a high-amplitude antiform with its crustal domain at about 1.4 km depth. This antiform marks the upper boundary of deeper sub-horizontal reflectivity at approximately 2 km depth. This is the same diapir seen at CMP 1000 on profile IAM-GC2.

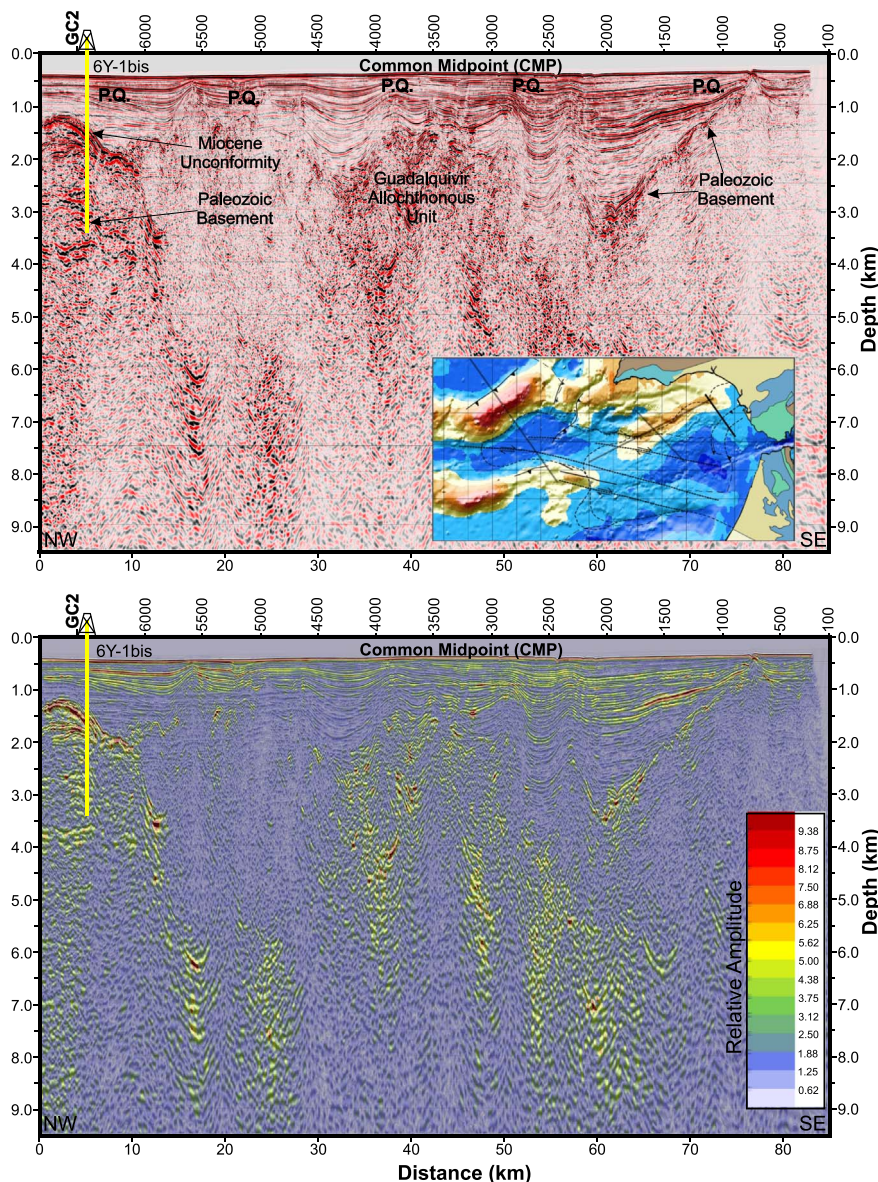
Beginning at CMP 6000 at depths between 2.0 and 5.5 km, there is a zone of low reflectivity that underlies the near-surface sediments, continuing until about CMP 4500. Beneath this seismically transparent zone there is some higher amplitude reflectivity (starting at about 6 km depth). In the center of the seismic profile, there is a large zone of strong reflectivity that contacts the Pliocene-Quaternary sediments and shows several examples of diapirism (e.g. CMPs 2300, 2800, 3800, and 4600).

3.2. Comparison of processing flows

Critical differences between our processing scheme and that of the original processing (Banda et al., 1995; Tortella et al., 1997; Gràcia et al., 2003) are: i) amplitude preservation and ii) multiple suppression/velocity analysis.

Automatic Gain Control (AGC) is commonly used to highlight weak signal by equalizing trace amplitudes in the temporal dimension. By its very nature, it destroys the relative amplitude relationship. Fig. 6 (top) shows the original processing result with AGC applied to the post-stack data. Although this scaling makes many weak events stand out, it is impossible to infer the reflectivity relationship of individual geological units. In contrast, the preserved amplitude profile shown in Fig. 6 (bottom), more accurately represents the actual reflectivity of the subsurface. For example, the near-surface sediments and the top of the antiform appear ‘bright’ because they represent a true impedance contrast with the low density/low velocity overburden. For the same reason, the sub-horizontal reflector at about 2.9 s two-way-travel-time appears weaker directly below the antiform than it does slightly to the SW. It is likely that its high

Fig. 5. MCS profile IAM-GC3. Top - final migrated depth profile. Black line of the lower right inset shows its location. Bottom - instantaneous amplitude attribute overlain on seismic data. Borehole location indicated in yellow. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)



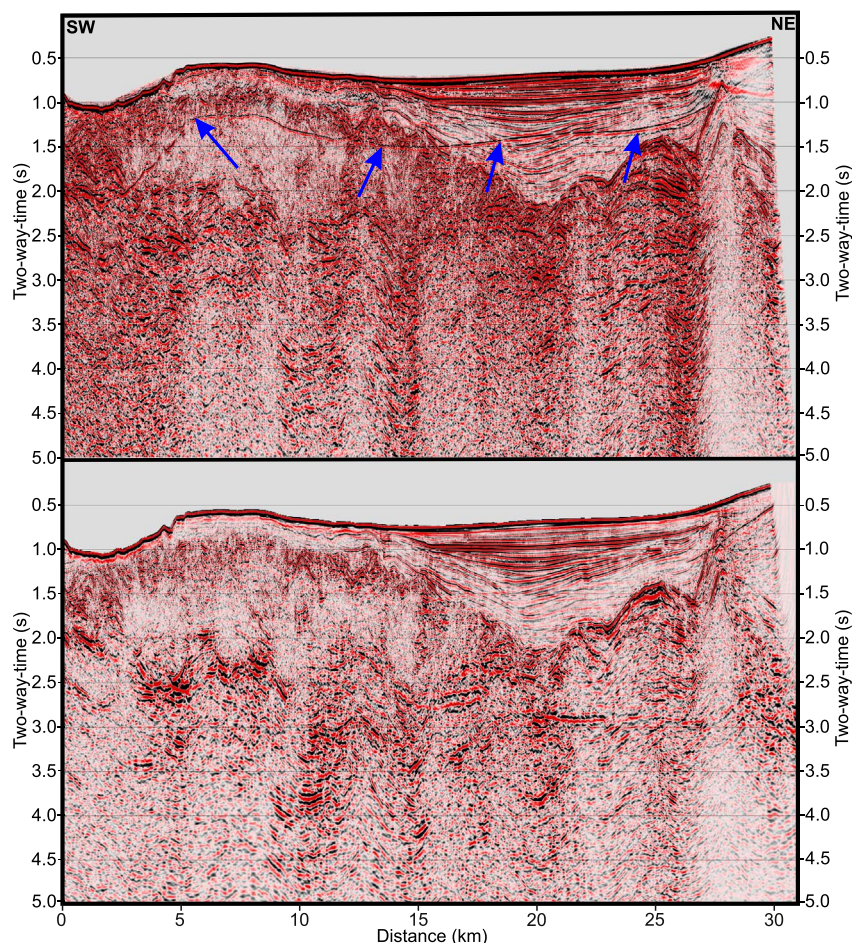


Fig. 6. Comparison between multiple suppression approaches for a portion of seismic profile IAM-GC2. Top - original processing including trace equalization and amplitude compensation. Bottom - the re-processed data with amplitude preservation showing the relative variation of seismic reflectivity. The suppression of the surface multiples improves the final image. In addition, it also increases the accuracy of the velocity model, since multiples do not bias hyperbolic moveout on the CMP gathers. Blue arrows indicate the position of the surface multiple in the original onboard processing. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

impedance and the antiform's geometry partially scatter impinging wave fronts, thereby 'shielding' the reflector directly below and so, diminishing its reflected energy (Section 2.4, limitation iii).

The second important distinction is the way in which multiple energy was identified and suppressed. The post-cruise processing by Geco-Prakla involved testing three seafloor demultiple approaches: i) the slant-stack method (Tatham et al., 1983), a slowness-frequency domain estimation, ii) wave equation (Monk, 1993), with a model-based adaptive subtraction, and iii) frequency-Wavenumber (FK) demultiplying (Wu and Wang, 2011). In the present study, we employed a modeling and adaptive subtraction approach called surface-related multiple elimination (SRME), which was developed by Verschuur et al. (1992).

SRME is a completely data-driven convolution process that can model the multiple wave-field directly from the data without an a priori sub-surface model. It does this by subdividing ray paths into their primary and multiple components, treating multiples like primaries, but with their source at the subsurface reflection point with the surface. Convolving these two events then permits the multiple to be predicted. SRME takes each different type of surface-related multiple and convolves the traces to make the prediction. The process considers all possible reflection points at the surface and sums over their contributions after combining common receiver and common shot gathers. In this way, each possible surface-related multiple is in theory predictable, and can be attenuated.

The most obvious difference between the two processing sets is the prominent multiple that remained in the original processing (blue arrows in Fig. 6 – top). The images in the present study show much better multiple attenuation (Fig. 6 – bottom). Nevertheless, our data set does contain some multiple energy at the extreme NE end of the line. This is due to the low-fold at the end of all seismic lines and how, for SRME,

shots must be interpolated to each receiver location.

The third difference pertains to velocity analysis. We paid careful attention to optimizing the velocity field. Velocities were determined using semblance and mini-stacks picked every 100th CMP and/or every 50th CMP for areas of sharper topographic relief. This corresponds to lateral distances of 1.25 km and 625 m, respectively. In contrast, the post-cruise original velocities were picked only every 6 km. We also fine-tuned the velocities by observing a 'near-far stack', which is a tool that shows color overlays of weighted near and far offset traces to reveal regions where the velocity model needs improvements. The offset weighting applied varies as $1/x^2$ (Ravens, 1995), where x is the source-receiver offset. The near-stack is weighted from 1.0 at the minimum offset to 0.0 at the far offset and is interpolated in between. The far-stack is weighted inversely.

A well-constrained velocity model is important because, in addition to optimizing the stack response, it permits proper conversion of two-way time to depth. We verified our time-depth conversion by comparing it the depths obtained for well 6Y-1bis (Fig. 7). Constrained velocities also improve confidence that the true-amplitude acoustic impedance contrasts better map density contrasts because acoustic impedance depends on both density and velocity. In this way, we can better infer contrasts in lithology and the nature of the reflectors (see Section 3.4).

3.3. Correlation with well 6Y-1bis

Fig. 7 demonstrates a clear correlation between the depth-converted seismic profiles and the stratigraphic core data of well 6Y-1bis, which is located near the intersection of IAM-GC2 (Fig. 4) and IAM-GC3 (Fig. 5).

The well data show that the antiform seen in both profiles is a Miocene unconformity composed of limestone. Several reflectors at depth show varying strength but fit the stratigraphic core data identified as Jurassic units. At approximately 3000 m, there is a drop in reflectivity that corresponds to a reported fault in the well data. At about 3200 m on both profiles, we identify the reflector that corresponds to the top of the Paleozoic basement. On IAM-GC3, we also report a distinct band of reflectors at a depth of 3600 to 4000 m, just below the maximum depth of the well.

The good fit of the well data to the seismic data gives us high confidence in our seismic velocity model, meaning that the reflectors across all profiles were optimally imaged. Moreover, it suggests that the calculated depths are well constrained and that the presented structures accurately represent the subsurface in a way that may be used to verify tectonic models.

3.4. Where is the Paleozoic basement under the GCGH?

One of the more interesting features of IAM-GC2 is the stratigraphic sequence between approximately CMP 5200–6200 (Fig. 8). Gràcia et al. (2003) identified the units in this area (from top to bottom) as Plio-Quaternary and Miocene sediments, and Paleozoic basement. The newly processed data set presented here calls into question this aspect of their interpretation. Rather, we suggest a sequence of Plio-Quaternary sediments, Paleozoic basement (henceforth Reflector 1), and an unknown body (henceforth Reflector 2). This interpretation derives from the re-assessment of the reflectors based on their instantaneous amplitude and phase attributes, as well as the comparison of reflection coefficients with the GB. We agree with Tortella et al. (1997) and Gràcia et al. (2003) that the bright reflector in the SW portion of Fig. 8 is a Paleozoic basement unit (henceforth SW Reflector). However, we

differ with them for the CMP range 5200–6000 because the instantaneous phase data of the SW Reflector and Reflector 1 clearly define a continuous event, spanning the location of the SW edge of Reflector 2 (Fig. 8, Inset A). In addition, the instantaneous amplitude data shown for Reflector 2 depicts a high impedance interface. So, if Reflector 2 were the same event as the SW Reflector, as previously reported, then it begs the question: why does Reflector 2 have similar amplitudes when overlain by Reflector 1 plus Plio-Quaternary units, as it does when it is overlain by only Plio-Quaternary units (SW Reflector)? Reflector 2 receives proportionally less wave front energy than the SW Reflector because it is overlain by Reflector 1 and the near-surface sediments—a significant proportion of energy is reflected at Reflector 1 and never reaches Reflector 2. Therefore, to account for the high reflectivity of Reflector 2, its impedance contrast must be proportionally higher, meaning that it is either denser, has a higher seismic velocity, or both.

Additionally, the instantaneous phase spectrum (Fig. 8, Inset B), although of mixed phase, defines the NE extent of Reflector 2. The instantaneous amplitude data (Fig. 4 and main image of Fig. 8) show weaker amplitudes when directly under the Guadalquivir Allochthonous Unit (GAU). This is most probably because the chaotic fabric of the GAU scatters seismic energy, having the effect of reducing the incident wave energy arriving at the units below. To the NE of Inset B there is discontinuous reflectivity that may either be of Paleozoic basement composition, or be similar to Reflector 2. Under the NE end of the GAU on IAM-GC2, and continuing under the antiform, we observe two weak but continuous phase events that are separated by a mixed-phase zone. The mixed-phase zone is located precisely at the NE boundary of the GAU and is most likely related to normal-faulted Paleozoic basement rocks.

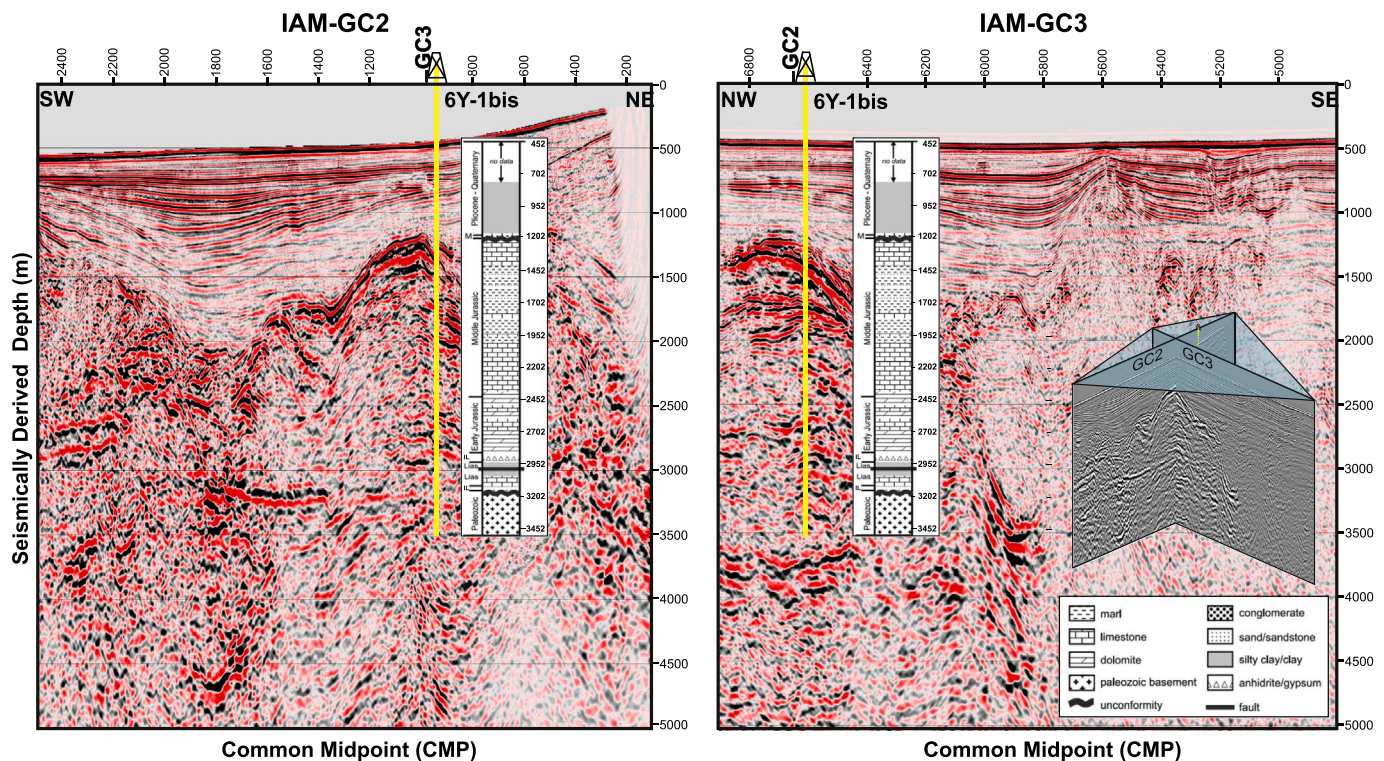


Fig. 7. Correlation of well log data from well 6Y-1bis (modified from Gràcia et al., 2003) to seismic depth profiles, IAM-GC2 and IAM-GC3. Well depths are shown adjusted with respect to the sea surface to correspond to seismic data. Well location does not correspond precisely with seismic profiles. For IAM-GC2, the tie point is 500 m SW and 2 km NW of the well. For IAM-GC3, the tie point is 2 km NW and 500 m SW of the well. Greyscale inset at right shows the location of the well and the seismic data tie point.

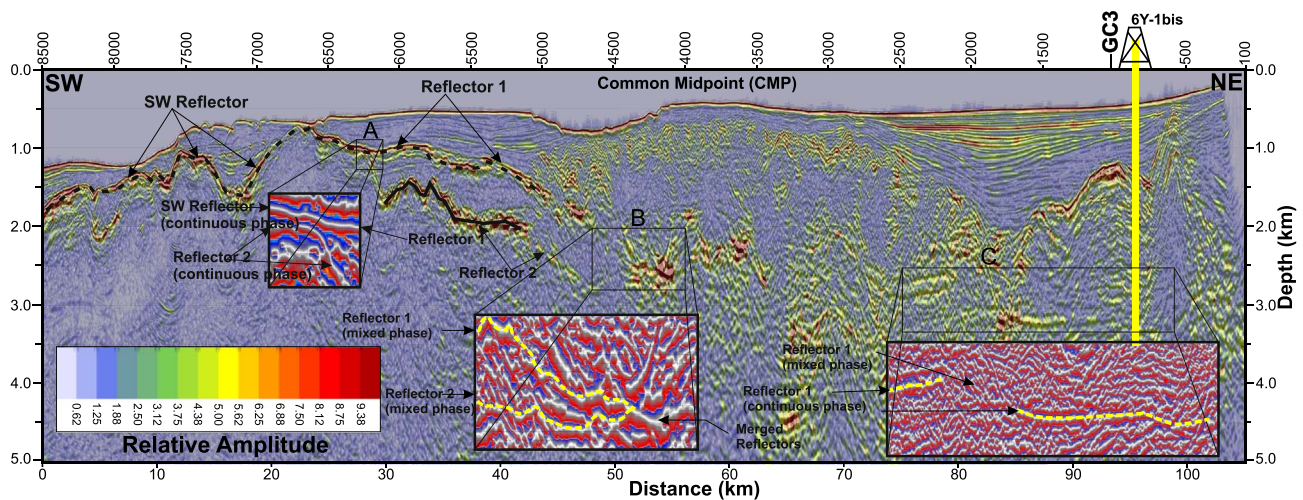


Fig. 8. A portion of the seismic profile IAM-GC2 showing SW Reflector, Reflector 1, and Reflector 2, as per discussion in Section 3.4. Zoom inserts show the instantaneous phase spectra, which is a measure of horizontal continuity.

3.5. Comparison of reflection coefficients

To better quantify the nature of Reflector 2, we calculated its reflection coefficient and compared it to a known high-density body from the same seismic survey (profile IAM-4, crossing the GB, Fig. 1) where extensive peridotite has been confirmed. Fig. 9 compares the GB to the antiform of IAM-GC3 and the GCGH. We chose the analysis location for the GB to be approximately 5 km to the SE of DSDP Site 120 because it was a more reliable location to identify the lithospheric mantle interface reflector reported by Sartori et al. (1994) for seismic profile AR92-3, and Jiménez-Munt et al. (2010) for profile IAM-4. Moreover, Reflector 2 at the GCGH was an interesting target to scrutinize because it spatially corresponds to an elevated local positive free-air gravity high (~ 100 mGal). Gràcia et al. (2003) identified the Paleozoic basement high as the cause of this gravity anomaly (therein, Fig. 9).

For this analysis, we only considered events on the common offset gathers that showed trace-to-trace amplitude stability. In this way, we are more confident that we are measuring the Earth's response at a single impedance interface, as opposed to regions where there are mixed-phase peaks or troughs. Mixed-phased events can indicate responses from a range of rock units, or can be noise-related artifacts. Hence, the complexity of the reflectors that were analyzed limited the areas we could confidently investigate. Some amplitudes for instance, yielded reflection coefficients of more than double or even greater than one (perfect specular reflectivity). We considered these to be outliers and therefore they were rejected from the calculation.

Using the method of Warner (1990), we quantified the similarity in reflection intensity (Section 2.3) by calculating reflection coefficients — normalized with respect to the seafloor sediment reflectivity in order to reduce the uncertainty introduced by the influence of the overburden. We report a 4.6% difference in R_{sf}/R_{body} ratios (Fig. 9 and Table 2). Assuming that the near-surface sediments have similar density and velocity characteristics, this implies that the acoustic impedance contrasts of the bodies are similar. In contrast, the R_{sf}/R_{body} ratio for IAM-GC3, which is near the location of the cored limestone Miocene unconformity (Gràcia et al., 2003) differs from IAM-4 by 35.8%, indicating that the composition of Reflector 2 is dissimilar from limestone.

Using the calculated values of reflection coefficient for Reflector 2 it is theoretically possible to calculate its density by rearranging Eq. (1) to solve for ρ_2 . However, in practice the set of unconstrained variables (precise velocity and density of the overburden, velocity at the reflector, geometry of the imaged structures, convolutional model assumptions, estimation of the calibration factor), does not provide a conclusive result.

The assessment of the nature of the reflectivity carried out in this contribution can be extended to any other regions or areas of interest taking into account the issues discussed in the methodological section of this study. Main points to consider are: use reflectors with minimal dips, near-vertical incidence ray paths, and well constrained calibration factors.

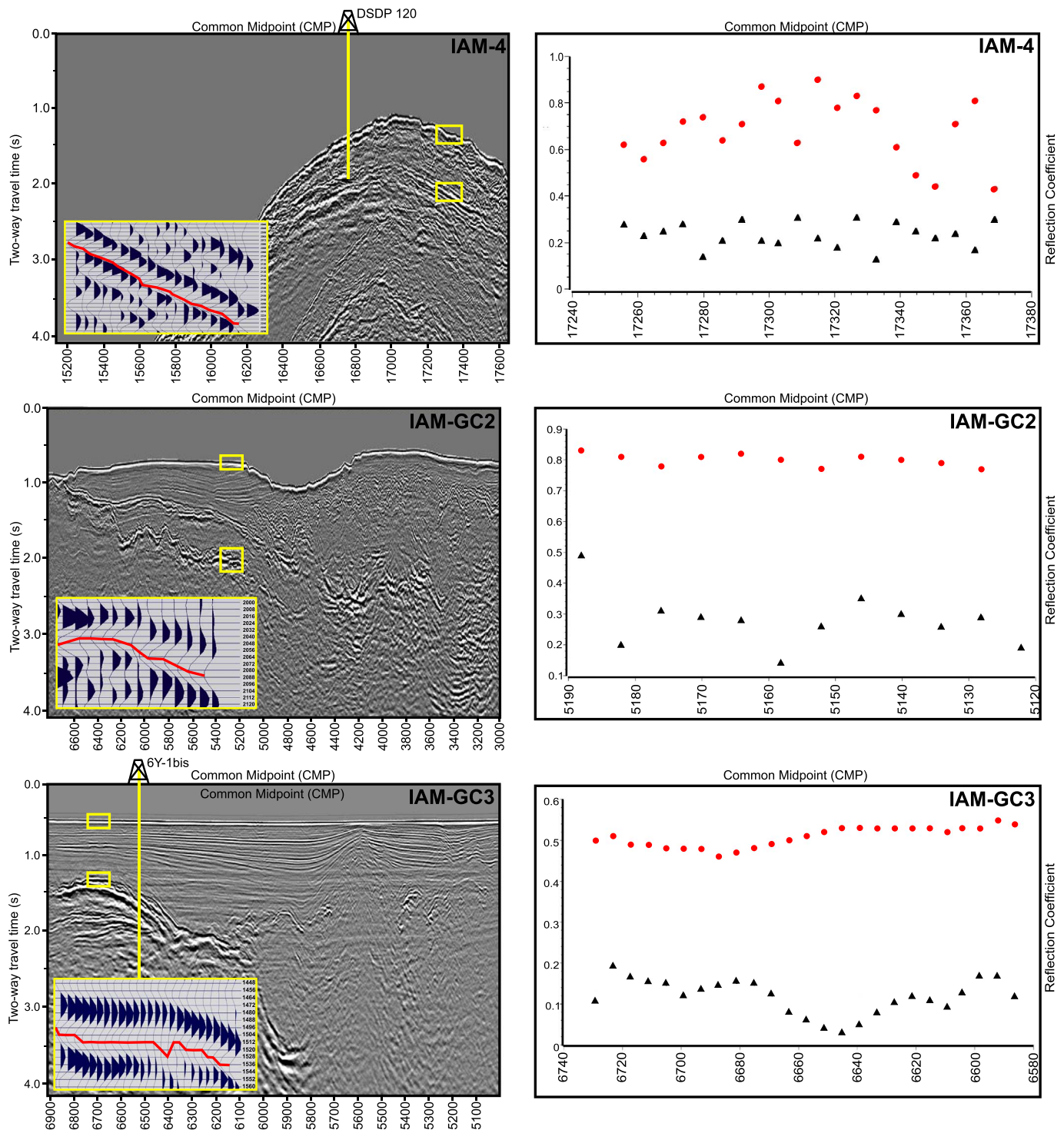


Fig. 9. Left - analysis locations where reflection coefficients (R) were calculated on common offset gathers (yellow boxes/insets) for the Goringe Bank (IAM-4), Gulf of Cádiz Gravity High (IAM-GC2), and the Miocene limestone unconformity from IAM-GC3. Right - plots of R for the sea floor (red circles) and for analyzed reflectors (black triangles) for corresponding regions. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

3.6. What is the nature of reflector 2?

We consider three plausible scenarios that could explain the nature of Reflector 2:

a) It is a fragment of ultramafic rocks exhumed/emplaced during the Jurassic extension that was later uplifted and thrust during the Paleogene-Neogene compressional stage that affected the whole area.

b) It is comprised of volcanic rocks. These may be a result of Central Atlantic Magmatic Province (CAMP) hydroclastic eruptions of basic volcanic rocks in the Gulf of Cádiz, which are related to Late Triassic volcanism.

c) It is composed of high-density intrusions inherited from the Variscan crust.

The first scenario considers Reflector 2 as an intracrustal fragment of continental mantle (serpentinized peridotite) exhumed during the

opening of pull-apart basins from the Late Jurassic to Lower Cretaceous that controlled the formation of the South Iberian margin, and was later emplaced at upper crustal levels during the Paleogene-to-Neogene compressional phase. As reported by Manatschal et al. (2010), rifting along the southern North-Atlantic seems to have been localized from Late Jurassic to Early Cretaceous along several local basins, resulting in a very thin crust (less than 10 km thick) that allows for local mantle exhumation. More recently, Ramos et al. (2017a), based on integrated MCS data tied with onshore geology, exploratory wells and gravity modeling, concluded that the SW Iberia margin is characterized by a pronounced thinning of the crust (~10 km) under the outer portions of the margin, with exhumed sub-continental mantle at lower crustal/base-of-the-crust levels in its distal parts. Although partial inversion of the SW Iberian margin is documented by Ramos et al. (2017b) and that the present day seismicity, as observed from focal mechanisms, indicates N-S to NW-SE directed shortening across the Gulf of Cádiz (e.g., Custódio et al., 2015 and references therein) the amount and extent of shortening does not seem sufficient to emplace pieces of exhumed mantle at upper crustal levels. Ramos et al. (2017b), based on interpretation of an industry 3D cube seismic data, conclude that the estimated horizontal shortening is less than 10 km, consistent with the ~20 km NW-SE estimated shortening for the Gorringe Bank, but not sufficient to emplace mantle material at shallow crustal levels as happens at the Gorringe Bank. Moreover, the analyzed MCS data do not show evidence of a mechanism (e.g. crustal thrust faults) that would be capable of carrying mantle rocks to the near surface. Therefore, despite the fact that the calculated reflection coefficient ratios differ by only 4.6% compared to the serpentinized peridotite that crops out at the Gorringe Bank, the lack of a clear geodynamic mechanism prevents us from supporting this explanation.

The second scenario postulates volcanic intrusions as the cause. These could be related to the CAMP rifting event, which is associated with the breakup of Pangaea and the opening of the Atlantic (Marzoli et al., 1999). The CAMP formation was a spatially extensive but temporally brief tectonic episode that may have allowed for the migration of volcanic dikes, sills and flood basalts of tholeiitic composition as far as 2000 km onto its passive continental margins. An example of CAMP-related magmatism observed in the present-day Iberia can be found in the Algarve Basin, which hosts units of subaerial lava flows, pyroclastic deposits, and peperites (Martins et al., 2008). Dating of these units by the $^{40}\text{Ar}/^{39}\text{Ar}$ method by Verati et al. (2007) shows ages of 198.1 ± 1.6 to 198.4 ± 2.8 Ma, which are almost coincident with the ages reported for the CAMP pulse elsewhere (e.g., Marzoli et al., 1999). CAMP volcanic sequences are widespread in the Atlas region and West African craton, showing that CAMP volcanism covers an extensive region of SW Iberia and NW Africa.

The third scenario considers the reflector to be related to the presence of mafic material inherited from the Variscan crust. Extensive Late Paleozoic mantle plume related volcanism and emplacement of dikes and sills took place in the Variscan Belt of Europe. Various authors (e.g., Carbonell et al., 2004; Palomeras et al., 2009) report the presence of mafic rocks at mid-crustal levels in the South Portuguese Zone of the southern Iberian Massif.

Salah (2014), based on a 3D inversion of P- and S-wave arrival times from local earthquakes beneath Southwest Iberia, conclude that at 4-km depth both P- and S-wave velocities take average to high values relative to the initial velocity model. The author attributes this to the possible presence of igneous intrusions as recorded in Southern Portugal. Most recently, Torne et al. (2015), based on integrated modeling of elevation, geoid and gravity data conclude that the gravity high observed in SW Iberia is partly associated with an increase in average crustal density between 15 and 30 kg m^{-3} .

Considering the above discussion and that the reflector is located at upper crustal levels (2–4 km), our preferred interpretation is that Reflector 2 is a volcanic intrusion that was emplaced because of the extensive CAMP, vestiges of which have been dated in the nearby

onshore areas of the Algarve Basin. However, due to the lack of age constraints from physical samples in the study area, we cannot completely rule out the possibility that it is an intrusion inherited from the Variscan crust.

4. Conclusions

One of the main findings of this study has been the identification of a seismic reflector underneath the Guadalquivir-Portimão Bank. We calculated and analyzed reflection coefficients for this feature and compared them with reflection coefficients for known serpentinized peridotites units at the Gorringe Bank and with a known Miocene limestone unconformity indicated by nearby well data.

The reflection coefficient comparison showed a difference of 4.6% between the Gulf of Cádiz Gravity High and the Gorringe Bank. In contrast, there was a 35.8% difference between it and a drilled Miocene limestone unconformity. This suggests that the reflector has a similar impedance to the Gorringe Bank reflector, which marks the presence of buried serpentinized peridotites. However, based on geological information it seems unlikely that the studied reflector has a similar geological origin because at present there is no evidence that shortening across the Gulf of Cádiz is sufficient to emplace ultramafic material at upper crustal levels. Moreover, the available MCS data do not show evidence of a mechanism (e.g. crustal thrust faults) that would be capable of carrying shallow mantle material to near the surface.

We consider the most likely scenario to be a magmatic intrusion that could have arisen from extensive CAMP volcanism that affected, among other regions, West Africa and South Iberia. Martins et al. (2008) conclude that the Lower Jurassic volcanic-sedimentary successions of the onshore Algarve basin are associated with the CAMP, favoring a passive rifting model to explain the preserved volcanostratigraphic sequence.

Reprocessing of the three MCS deep seismic regional profiles also shows a very poor seismic response at deep crustal levels, that it is likely related to the presence of high-velocity/high-density layered bodies at upper crustal levels, which would hinder the penetration of seismic waves at depth.

Reprocessing Gulf of Cádiz Iberian-Atlantic Margin multichannel seismic data improves on those done in the mid-to-late 1990s for several reasons. First, by carefully picking stacking velocities, testing wavelet-modifying parameters, and avoiding harsh trace-scaling processes, we approach as best as possible to a true relative amplitude stacked seismic profile. Second, we present our data with a demultiple scheme that greatly improves the interpretation, as well as the confidence in our velocity model, which has been verified against nearby well data. Finally, we present depth-converted profiles with twice the spatial resolution of the post-cruise processed data set.

The approach taken enhances lateral continuity and uncovers previously unseen structural details in the data. Amplitude preservation allows for the interpretation of true relative amplitudes using instantaneous attributes. As instantaneous amplitude represents seismic reflection strength, when velocities are well determined we are able to generate proxy maps for density contrasts, helping to constrain interpretations and geodynamic models.

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.tecto.2017.10.006>.

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Table 2

Calculated average reflection coefficient ratios and percentage difference in comparison to control case IAM-4 (Gorringe Bank) for seismic lines IAM-GC2 (Gulf of Cádiz Gravity High) and IAM-GC3 (Miocene unconformity – limestone (Fig. 7)).

	IAM-4	IAM-GC2	IAM-GC3
Average R_{sf}	0.69	0.80	0.51
Average R_{body}	0.23	0.26	0.12
R_{sf}/R_{body}	2.92	3.06	4.20
IAM-4 R ratio difference	Control	4.6%	35.8%

their constructive comments that greatly improved the first version of the manuscript. Fig. 1 was made using GMT (Wessel and Smith, 1998).

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