Geomorphology and Neogene tectonic evolution of the Palomares continental margin (Western Mediterranean)

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Abstract

The Palomares continental margin is located in the southeastern part of Spain. The margin main structure was formed during Miocene times, and it is currently part of the wide deformation zone characterizing the region between the Iberian and African plates, where no well-defined plate boundary occurs. The convergence between these two plates is here accommodated by several structures, including the left lateral strike-slip Palomares Fault. The region is characterized by sparse, low to moderate magnitude (Mw < 5.2) shallow instrumental earthquakes, although large historical events have also occurred. To understand the recent tectonic history of the margin we analyze new high-resolution multibeam bathymetry data and re-processed three multichannel seismic reflection profiles crossing the main structures. The analysis of seafloor morphology and associated subsurface structure provides new insights of the active tectonic features of the area. In contrast to other segments of the southeastern Iberian margin, the Palomares margin contains numerous large and comparatively closely spaced canyons with heads that reach near the coast. The margin relief is also characterized by the presence of three prominent igneous submarine ridges that include the Aguilas, Abubacer and Maimonides highs. Erosive processes evidenced by a number of scars, slope failures, gullies and canyon incisions shape the present-day relief of the Palomares margin. Seismic images reveal the deep structure distinguishing between Miocene structures related to the formation of the margin and currently active features, some of which may reactivate inherited structures. The structure of the margin started with an extensional phase accompanied by volcanic accretion during the Serravallian, followed by a compressional pulse that started during the Late most Tortonian. Nowadays, tectonic activity offshore is subdued and limited to few, minor faults, in comparison
with the activity recorded onshore. The deep Algero-Balearic Basin is affected by surficial processes, associated to halokinesis of Messinian evaporites.

Keywords: Multichannel Seismic reflection, swath-bathymetry, geomorphology, SE Iberia margin, geodynamic evolution.

1. Introduction

The boundary between the Eurasian and African plates in the westernmost Mediterranean Sea runs between the southern margin of the Iberian Peninsula and the northern margin of Africa. This boundary corresponds to a wide deformation zone, controlled by the NW-SE trending convergence (4.5 – 5.6 mm/yr) between the Eurasian and African plates (e.g. Booth-Rea et al., 2007; Serpelloni et al., 2007; Nocquet, 2012). Deformation is distributed over a large number of faults onland with terminations offshore, forming what has been defined as the Eastern Betics Shear Zone (EBSZ) among other structures (e.g. Bousquet, 1979; Sanz de Galdeano et al., 1990; Andeweg and Cloetingh, 2001). The EBSZ corresponds to a large, active seismogenic fault-system that from north to south is composed by the Bajo Segura Fault, the Alhama de Murcia Fault, the Palomares Fault and the Carboneras Fault, which continues into the Alboran Sea (Fig. 1) (e.g. Gràcia et al., 2006; Moreno et al., 2016). There is limited evidence of relevant seismic activity offshore the southeastern Betics (e.g. Stich et al., 2003; García Mayordomo, 2005; Martín et al., 2015). Instrumental events of low to moderate magnitude (MW < 5.2) have been recorded, and historical earthquakes, such as the 1518 Vera Earthquake of I0=X (Fig. 1), indicate slip along large structures. Recently acquired geodetic data also show evidence of ongoing deformation in the area.
(Echeverría et al., 2013, 2015). Strain rates deduced from geodetic measurements are around 2 mm/yr (e.g. Nocquet, 2012; Echeverría et al., 2013). These low values, combined with the erosive and depositional processes occurring at the Palomares margin, may explain the few evidences of active deformation in the offshore areas, as well as in the narrow continental shelf. New bathymetric maps and multichannel seismic reflection profiles have been collected along the margin to unveil the detailed geomorphology of this area and to reconstruct its recent tectono-sedimentary evolution. The Palomares continental margin is heterogeneous and displays a rough bathymetry due to the presence of prominent ridges and deep canyons, thus marking a difference with adjacent western Mediterranean margins (e.g. Pérez-Hernández et al., 2009, 2014). The deep structure of the Palomares margin has recently been presented and discussed by Giaconia et al. (2015). The aim of this work is to further extend the characterization of the margin, focusing on the shallow structure using new high-resolution data. This approach is essential to characterize the active tectonic and sedimentary processes occurring along the Palomares continental margin. In this work we present full-coverage high-resolution bathymetric data from the continental shelf to the deep basin along the Palomares margin. In addition, we have re-processed multichannel seismic reflection (MCS) profiles TM23 and TM24 presented in Giaconia et al. (2015) to improve the resolution of the shallow structure and added profile TM25 to our study.

The objectives of this work are: (i) To characterize in detail the geomorphologic features of the margin, which is now possible on the basis of new high-resolution bathymetric data; (ii) To link the shallow expression of these structures with their origin at depth, through the analysis of the seismic images; and (iii) To explain these structures in their present-day compressive geodynamic framework, linking them with the geodynamic evolution of the margin.
2. Geological setting

The Gibraltar Arc system is the western termination of the Alpine orogen in the Mediterranean. Most models propose initial subduction of the Tethys lithosphere towards the north under the Balearic Islands during the Eocene-Oligocene. It later followed by subduction towards the east, as the Tethys oceanic lithosphere between Iberia and Africa rolled back towards the west (e.g., Lonergan and White, 1997; Booth-Rea et al., 2005; Chertova et al., 2014). This subduction process led to the later collision of the Alboran domain, consisting of a thrust stack of metamorphic complexes (e.g. Sanz de Galdeano, 1990), with the passive margins of European (Iberia) and African plates, generating the Betic (South Iberia) and the Rif (North Africa) orogenic belts and the Alboran Basin within them (e.g., Jolivet et al., 1999; Rosembaum et al., 2002; Faccenna et al., 2004; Schettino and Turco, 2006, 2011). Traditionally, the Alboran domain has been considered as a stack of three polymetamorphic terrains, referred to as Malaguide, Alpujarride and Nevado-Filabride complexes (e.g. Sanz de Galdeano, 1990). Recent studies reveal that Malaguide and Alpujarride thrust stack underwent high pressure (HP) metamorphism during the Eocene and represent a collapsed orogen (e.g. Azañón et al., 1997, Platt et al., 2005), being the Alboran domain sensu stricto, while the Nevado-Filabride complex is formed by subducted parts of the South Iberian margin and it is affected by a later HP metamorphism, during the Lower to the Middle Miocene (Platt et al., 2006; Booth-Rea et al., 2015).

The Betics are the northern branch of the Gibraltar Arc System. The Betic thrust belt consists of the superposition of different tectonics domains, and is mainly formed by the External Zones and the Internal Zones. The External Zones consist of the Mesozoic and Tertiary sedimentary cover of the South Iberian paleomargin, and the Internal Zones are composed by stacked sequences of the polymetamorphic terrains referred above (i.e.
Malaguide, Alpujarride and Nevado-Filabride complexes; e.g. Sanz de Galdeano, 1990; Vera et al., 2004; Balanyá et al., 2007). In this convergent setting (e.g. Dewey et al., 1989; van Hinsbergen et al., 2014) other tectonic mechanisms such as slab roll-back, or slab tearing under the Maghrebian and Betic margins have driven both, shortening in a direction perpendicular to plate convergence and extension that exhumed HP rocks in the core of the orogen (e.g. Lonergan and White, 1997; Martínez-Martínez et al., 2002; Faccenna et al., 2004; Booth-Rea et al., 2015; Mancilla et al., 2015). Due to slab roll-back and slab tearing under the southeastern Iberian and north African margins, both shortening and extension perpendicular to the plate convergence direction occur (e.g. Lonergan and White, 1997, Faccenna et al., 2004; Booth-Rea et al., 2015). The extensional process that led to the basin formation took place mainly during the Early and the Middle Miocene in the Western Alboran Basin, and from the Late Miocene until the latest Tortonian in the Eastern Alboran and the Algero-Balearic basins (e.g., Krautworst and Brachert, 2003; Booth-Rea et al., 2004; Giaconia et al., 2014). Onshore, in the eastern region, different mechanisms of extension have been proposed. The extension was mainly driven by magmatic accretion south of the Carboneras Fault (Rutter et al., 2012), and by normal faulting north of it (e.g., Krautworst and Brachert, 2003; Booth-Rea et al., 2004; Giaconia et al., 2014). Afterwards, from the Late Tortonian to present, the dominant mechanism affecting the Alboran Basin has been the NW-SE trending plate convergence that defined the basin’s present-day physiography and structures (e.g., Watts et al., 1993; Comas et al., 1999; Wortel and Spakman, 2000; Rosenbaum et al., 2002; Gràcia et al., 2006, 2012; Moreno et al., 2016).

Successive volcanic episodes in the basin are related to the evolution of the subduction zone. The first volcanic activity occurred during the Oligocene to the Early Miocene, and it is represented by tholeiitic dykes related with extension in the basin that are found
onshore in the Malaguide complex and in the Ronda peridotites (e.g. Torres-Roldan et al., 1986; Duggen et al., 2004; Marchesi et al., 2012). From the Middle Miocene until the Tortonian (Late Miocene), calc-alkaline, tholeiitic and shoshonitic volcanism related to the subduction process took place (Duggen et al., 2008). During the Messinian to Quaternary, volcanism changed to alkaline, reflecting a dominance of hot asthenosphere material, rather than subduction related fluids (Duggen et al., 2004, 2008).

The Palomares continental margin, with a NNE-SSW trend, is located in the southeastern part of the Iberian Peninsula (from 36°30′N to 37°15′N, and from 2°15′W to 1°W), including the eastern termination of the Betics and the easternmost part of the Alboran Basin. This margin is formed by thinned continental crust affected by volcanism (Booth-Rea et al., 2007). The volcanism is dated between 12 Ma and 6 Ma (Duggen, 2008), and the volcanic rocks sampled belong to tholeiitic, calc-alkaline and shoshonitic series (Comas et al., 1999; Duggen, 2004). The margin is composed by an extremely narrow linear shelf parallel to the coastline, suggested to be controlled by the Palomares Fault (PF) (Comas and Ivanov, 2006), and by a broad slope with a rough bathymetry with deeply incised submarine canyons and prominent volcanic and plutonic ridges (Comas and Ivanov, 2006). Strike-slip and reverse faulting and folding dominate the deformation in this region, with Carboneras Fault and PF as the most important structures onshore (Fig. 1). The PF is well defined onshore as a NNE-SSW left-lateral strike-slip fault, (e.g. Silva et al., 1993; Comas et al., 2000; Booth-Rea et al., 2004; García Mayordomo, 2005), but its potential extension offshore is unclear. Several segments form the PF, with a fault deformation zone about 4 km wide, and a lateral displacement of around 15 km. The PF has been active since the Tortonian, and during the Quaternary, only the easternmost segments have been clearly active (e.g. Booth-Rea et al., 2004). The PF reaches the coastline ending in reverse fault splays at the southern
and northern sides of the Sierra Cabrera (Giaconia et al., 2012). The presence offshore of other PF segments is not clear, and has been proposed that deformation is transferred to its conjugate dextral Polopos Fault zone and associated reverse structures in Sierra Alhamilla and Sierra Cabrera (Giaconia et al., 2012; 2013) (Fig. 1, 2).

3. **Data and methods**

This study results from the integration of two different datasets: a) High-resolution bathymetry and b) MCS profiles. The new swath-bathymetry, which covers from the continental shelf, across the slope and to the deep basin of the Palomares margin, was acquired on board of the Spanish R/V “Vizconde de Eza”. A Kongsberg EM-300 multibeam echosounder, operating 135 beams at 30 KHz frequency was used. Between years 2001 and 2003, a total of four oceanographic cruises were carried out in the framework of the project “Fishing Charts of the Mediterranean” (CAPERMA) by the Spanish Institute of Oceanography (IEO) and the General Secretariat of Maritime Fisheries (SGPM). Multibeam data have been processed using Neptune software, and interpolated to a 100x100 m side regular grid. Detailed bathymetric and slope maps have been obtained (Figs. 1, 2, 3).

The MCS profiles were acquired during the TOPOMED-GASSIS cruise (October 2011), on board of the Spanish R/V “Sarmiento de Gamboa”. The seismic data were acquired using a 5100 meters long active section of a Sentinel SERCEL streamer with 408 active sections (12.5 m channel interval), towed at 10 m depth, and a 50.15 l (3060 ci) air-gun source. The source array was composed of 8 G-GUN II guns deployed at 7.5 m depth, in a single cluster distribution. The airgun shots were fired every 50 m at a pressure of 2000 psi. The total record length was 14 s (Two Way Travel Time, TWTT), with a sample rate of 2 ms.
Three MCS profiles have been used for this study: TM23, TM24 and TM25 (Fig. 2). These profiles have been processed using “GLOBE Claritas” software. Lines TM23 and TM24 have been published by Giaconia et al. (2015) with a processing aimed at deep penetration. For this work, the processing flow was designed to obtain the maximum resolution of the shallow subsurface structure, to ensure a high-resolution imaging of the sedimentary cover and top of the basement (< 6 s TWTT). Processing steps include minimum-phase conversion, geometry definition accounting for streamer feathering, spherical divergence correction, predictive deconvolution in Tau-P domain (to eliminate the bubble and short periods multiple reverberations), surface consistent deconvolution, normal-move-out correction based on velocity semblance analysis, stretching mute, amplitude recovery, time migration and a high-pass frequency filter (1-3 Hz). Finally, we used the IHS Kingdom Advanced software to represent the stratigraphic and structural interpretation of the MCS dataset presented here.

4. Results

4.1. Seafloor morphology

The margin has three areas differentiated: (i) the continental shelf, (ii) the continental slope and (iii) the Algero-Balearic deep basin (Fig. 2).

The continental shelf is divided in two parts, the inner shelf and the outer shelf that are separated by a pronounced escarpment (Acosta et al., 2013) (Fig. 2). The inner shelf has an average width of 4 km, extending up to 19 km in the southern part of the study area, at the Cabo de Gata Spur. The slope angle is less than 1º and the maximum depth is close to 80 m. On its outer part, an undulated topography can be recognized (Fig. 3). The escarpment separating the inner and outer shelves is located between 80 and 120m depth, and has a dip between 7º and 15º (Fig. 2b). This scarp shows an undulating trace
north of 37°20’N. It is slightly concave from 37°20’N to 37°N and convex south of
37°N (Fig. 2a), and it is modelled by erosional structures, as gullies or submarine
canyon heads. The outer shelf, although locally eroded by canyons and gullies (Fig. 2a),
is relatively flat with dips ranging from 1° in the north to 2° in the southern part (Fig.
2b). Thus, the outer shelf has been extended down to 450 m water depth (Acosta et al.,
2013). The base of the outer shelf is defined by an undulated topography easily
recognized on the slope map (Fig. 2b).

The continental slope extends from ~450 to 2500 m depth, and corresponds to the
transition zone between the shelf and the basin. The outer shelf reaches 450 m and is
unusually deep for a continental shelf structure, but the term is used because of its dip,
which is much lower than at the continental slope (Acosta et al., 2013). Along the slope
there are three large topographic highs, which are the most characteristic features of this
part of the margin. From north to south, they are: the Aguilas High, the Abubacer High
and the Maimonides High. These elongated highs are separated by prominent submarine
canyons that will be described later (Fig. 2). The Aguilas High is 40 km long and 15 km
wide. Its deepest point is located at 2000 m water depth, and the shallower one is at 850
m water depth (Figs. 2a, 3a). The dip of the northern side of the Aguilas High varies
between 13° and 19°; being slightly steeper than the dip of the southern side, which
ranges between 8° and 12° (Fig. 2b). The Aguilas High is 25 km long and 12 km wide
and follows an east-west linear trend. It can be divided into two different sections: The
western part shows a flat top at a water depth of around 1000 m, with a south flank
highly affected by gullies, while the eastern part exhibits a more rugged topography
(Fig. 3a). The Abubacer High is 40 km long, 12 km wide and trends NE-SW (Figs. 2a,
3b). The shallower depth is 290 m and its base is located at 1600 m depth. The northern
part shows an irregular topography due to the presence of abundant gullies (Fig. 3b).
The southern part of the Abubacer High is named as Polacra Ridge, characterized by a sharp, narrow crest and steep slopes higher than 35° (Fig. 3b). To the east of the Polacra Ridge, a large (22 km wide x 12 km long) sedimentary lobe is observed (Fig. 3b). The W-E trending Maimonides High has a maximum length of 62 km, and it is up to 25 km wide (Fig. 2a). Its highest point is located at 820 m depth, while the base of the high is at 2300 m water depth. The Maimonides High shows a very rough, irregular relief, and the slope dip varies considerably, from 2° to 30° (Fig. 2b). These very steep slopes together with strong erosive processes cause a progressive dismantlement of the flanks of Maimonides High (Fig. 2a, 3b). Towards the east, the Maimonides High splits into two ridges, the Genoveses Ridges, separated by a narrow and elongated trough (Fig. 3b).

Along the Palomares margin, the Algero-Balearic Basin has an average depth of 2500 m and a subtle eastward dip of less than 1° (Figs. 2a, 2b). The basin is characterized by comparatively small-scale morphologies of swells and furrows (Figs. 2a, 3).

The most prominent erosional features affecting the entire margin are submarine canyons. The canyons run from the continental shelf to the Algero-Balearic Basin. The three main canyons are from north to south: Aguilas Canyon, Almanzora-Alias-Garrucha Canyon and Cabo de Gata Canyon (Fig. 2). The W-E trending Aguilas Canyon is 50 km long running from 100 m to 2400 m depth, with an average gradient of 2° (Figs. 2, 3a). The canyon’s maximum width, less than 3 km, is reached at the canyon mouth. The Aguilas Canyon runs between the northern flank of the Aguilas High to the south, and a trough terrain with a downward sequence of east-facing steps, to the north (Fig. 3). The Aguilas Canyon is composed of two small canyon heads, both cutting the escarpment between the inner and outer shelf (Fig. 3a). The Almanzora-Alias-Garrucha Canyon is a N45 trending, relative linear structure with a very asymmetric cross section. It is 80 km long, running from 50 to 2550 m depth, has an average gradient of 2°, and a
width of 0.5 to 3 km (Figs. 2, 3a). The northern flank of the canyon is characterized, from the shallower to the deepest part, by a gullied escarpment south of the Aguilas High, a 5 km wide terraced scarp and a steep embayed scarp (Fig. 3a). The southern wall is narrow, steep and linear, locally disrupted by gullies. The Almanzora-Alias-Garrucha Canyon is composed by two deeply-incised canyon heads that feed a meandering system made of multiple tributary channels. The canyon mouth is characterized by a succession of sedimentary lobes, stepping towards the east (Fig. 3a).

The W-E trending Cabo de Gata Canyon is located between the Abubacer and Maimonides highs. It is 72 km long, ranging from 80 to 2500 m depth, with an average gradient of 2°, and a width from 500 m at the upper-middle part to 2.5 km at the canyon mouth (Figs. 2, 3b). The Cabo de Gata Canyon is formed by two narrow canyon heads that cut the inner shelf (Fig. 3b). The geomorphology of both canyon walls is diverse, probably dominated by the different erosive processes between the Abubacer (northern wall) and Maimonides highs (southern wall). The lower part of the Cabo de Gata Canyon north wall is deeply incised in the sedimentary lobe located to the east of the Polacra Ridge (Fig. 3b). The south wall of the Cabo de Gata Canyon is wider, with a low-dip angle, and highly eroded by gullies (Fig. 3b).

4.2. Seismostratigraphic units

The post-Tortonian stratigraphy of the Eastern Alboran Basin is reasonably well established (e.g., Jurado and Comas, 1992; Comas et al., 1995; Alvarez-Marrón, 1997; Comas et al., 1999; Moreno, 2011; Booth-Rea et al., 2007; Giaconia et al., 2015; Moreno et al., 2016). These studies integrate commercial and scientific wells (ODP Leg 161, Comas et al., 1996), essential to calibrate the Alboran Basin sedimentary infill. A better definition of the Late Miocene series is achieved by the onshore-offshore correlation of the pre-Pliocene sedimentary units (e.g. Martín et al., 2003; Giaconia et
al., 2014; Giaconia et al., 2015). We have defined five main seismostratigraphic units on the basis of their seismic expression and correlation with previous works, allowing us to assign an age to the sedimentary units identified in the Palomares continental margin (Fig. 1). From top to bottom, the seismostratigraphic units are labeled I to V, and below them we found the basement of the margin (Figs. 4-7).

Unit I can be subdivided into two sub-units: Sub-unit Ia is characterized by parallel well-stratified reflections of variable lateral continuity, and corresponds with Late Pliocene to Quaternary age. It has a constant thickness, between 0.2 and 0.3 s in TWTT. Sub-unit Ib presents an overall more transparent character corresponding to Early Pliocene deposits (Fig. 4). Its thickness ranges between 0.35 and 0.45 s in TWTT (Figs. 5-7).

Unit II corresponds to Upper Messinian deposits bounded on their uppermost part by the Messinian Unconformity (M) (Figs. 4-7). This is a deeply erosive surface widespread in all the Mediterranean basins, generated during the Messinian Salinity Crisis, when the connection with the Atlantic Ocean was closed and a large part of the Mediterranean Sea desiccated (e.g., Hsü et al., 1973; Rouchy and Caruso, 2006; CIESM, 2008, Roveri et al., 2014). Because of its erosive top, it is not continuous along the profiles, and presents a remarkable variability in thickness. The seismic expression of this Messinian unit is characterized by strong, high-amplitude and low-continuity reflections alternating with transparent bodies. In the Algero-Balearic Basin, thick evaporitic deposits are found. These deposits mark a difference in Unit II between the western and the eastern part of the study area, and then allow us to subdivide Unit II in the Algero-Balearic Basin into three subunits: IIa, IIb and IIc (Fig. 4c). They can be correlated with the Messinian Salinity Crisis units defined throughout the whole Mediterranean Sea (e.g., Lofi et al., 2011), and referred to as Upper Unit, Mobile Unit
and Lower Unit, respectively (Fig. 4c). Subunit IIa, is equivalent to the Upper Unit (UU), and corresponds to the upper evaporites, characterized by continuous reflectors formed by salt and gypsum deposits with marls and limestone alternations. At its top, subunit IIa or UU is bounded by the Top Surface or Top Erosion Surface (TS/TES) depending on its erosive character and is a strong seismic reflection correlated with the M unconformity (Lofi et al., 2011) (Fig. 4c). Subunit IIb corresponds to the Mobile Unit (MU) that is the evaporitic deposit (mainly halite) (Lofi et al., 2011), showing an overall transparent character. Finally, Unit IIc is equivalent to the Lower Unit (LU), and corresponds to the lower evaporites (supposed to be gypsum, Lofi et al., 2011). They are characterized by continuous and low-amplitude seismic reflectors (e.g. Booth-Rea et al., 2007). At its base, is located the Bottom Surface or Bottom Erosion Surface (BS/BES), depending on its erosive character, which marks the base of the deposits related to the Messinian Salinity Crisis (Lofi et al., 2011) (Figs. 4c, 5-7).

Unit III is the Lower Messinian sedimentary unit, formed by the pre-Messinian Salinity Crisis open-marine marls. It is characterized by low-amplitude and low-reflectivity reflections with a maximum thickness of 0.25 s TWTT (Figs. 6, 7).

Unit IV is characterized by highly-reflective continuous reflections, of Late Tortonian age (Fig. 4). It appears infilling the basement troughs, with limited lateral continuity, and its thickness reaches up to 0.4 s TWTT (Figs. 5-7).

The deepest sedimentary unit recognized is Unit V, interpreted as Lower Tortonian deposits. The seismic image of this unit is characterized by strong reflections with low lateral continuity, alternating with transparent, chaotic zones (Fig. 4). It is identified only at the deepest part of the basement depocenter, and it is not continuous along the profiles. It achieves a maximum thickness of 0.45 s TWTT.
Finally, we imaged the basement (B) of the area. In the Palomares margin, the basement has been interpreted as thinned continental crust intruded by magmatism that transitions into the oceanic crust of the Algeo-Balearic Basin (e.g. Booth-Rea et al., 2007). On the seismic images, the top of the basement corresponds to a high-amplitude reflection marking an irregular surface, with no coherent reflections underneath (Figs. 4-7).

4.3. Tectonic structure and stratigraphy of the Palomares margin

The multichannel seismic profiles presented show the main tectonic features of the Palomares margin. They cross some of the main structures, imaging the topographic highs (TM23 and TM24) and the southern flank of the Almanzora-Alias-Garrucha Canyon (TM25) (Fig. 2a). Now, we individually describe, from north to south, each of the seismic sections because they image different structures from three distinct areas.

The northern profile TM25 (Figs. 2a, 5) runs approximately W-E along the southern flank of the Almanzora-Alias-Garrucha Canyon (Fig. 2a), which shows surface incisions associated to a large gully system. The top of the crystalline basement shows an irregular topography, defining some gentle highs and lows mainly filled by the pre-Messinian unconformity sedimentary series (units II, III and IV). The deposition of these units was affected by faults that are now sealed by the Messinian unconformity (M). These deformation processes are evident on Unit II that presents reflections conformable with the limit between Units III and II, but with a current high-dipping angle. However, some of the reflections ending in a toplap geometry against M can also be due to deep erosion during the Messinian, and not only caused by a tectonic deformation process. Above the M reflection, we find a parallel, well-stratified unit Ib, with its characteristic transparent facies. Above lays Unit Ia, represented by parallel and continuous reflections. These reflections are disrupted on their uppermost part by the
incision of gullies and erosive slide scars, such as the lateral scars of a slope failure (i.e. between 15 and 23 km, Fig. 5). In the westernmost part of profile TM25, there is a vertical fault clearly affecting the basement and almost all the sedimentary units. The topmost reflections of Unit Ia are continuous above the uppermost branch of this fault, so we infer that the fault does not reach the seafloor, or that the offset is too small to be resolved (Fig. 5). However, the seafloor seems to be slightly deformed, thus we cannot reject that the fault may be still active.

Profile TM24 (Fig. 2a, 6) runs in a NW-SE direction from the south flank of the Almanzora-Alias-Garrucha Canyon until the Algero-Balearic Basin, which allows us to image in detail the Abubacer and the Maimonides highs, the sedimentary apron along the southern flank of the Abubacer High and the mouth of the Cabo de Gata Canyon. Along this profile, the basement also exhibits an irregular topography, affected by deeply rooted faults. The roughness of the top of the basement is more evident due to the presence of the two large basement highs: the Abubacer and the Maimonides highs. These highs configure three main sectors of the TM24 profile, in terms of sedimentary characteristics. The northwestern part of the profile (Fig. 6a) has a thin sedimentary cover progressively thickening towards the Abubacer High, where the Abubacer Basin is located. On this part of the section, unit IV is only identified in the deepest local basin, located at the northwestern flank of the Abubacer High. Units III and II are not continuous along this part of the profile, they appear folded and faulted. Above them, units Ib and Ia show similar characteristics as in the previous profile. The top of Unit Ia also displays evidence of gullies. At the eastern part of the Abubacer High there is a sedimentary lobe that covers its southern flank. The maximum sedimentary thickness in this part of the profile reaches 1.5 s in TWTT. Unit V is identified at the deepest part of the basin. Units IV, III and II appear less deformed than in the northwest section of the
profile and are characterized by parallel and continuous reflections (Fig. 6a). The M unconformity is covered by units Ib and Ia, and again some erosive gully-type features are identified at the seafloor. The sedimentary lobe is cut by the lowermost part of the Cabo de Gata Canyon (Fig. 6b). It corresponds to a deep, asymmetric incision that almost reaches down to the M unconformity. The seismic image clearly displays the asymmetry between the two flanks of the canyon, with a steep northwestern flank and a relatively flat southeastern one. Southeast of Cabo de Gata Canyon (Fig. 6b) is the Maimonides High, a basement topographic high with rough topography and chaotic seismic facies that difficult a clear seismic imaging. Some of the roughness seems to be related to cone-shaped and ridge-like structures observed on the bathymetry and also imaged in the seismic data (Figs. 2, 3b, 6b). The southern end of the profile images the western part of the Algero-Balearic Basin. In this area, halokinesis stands out. Two sub-vertical normal faults separate the Maimonides High from the deep basin. Within Unit II, we recognize characteristic deposits of evaporites (Fig. 6b), in which halokinesis processes are active. The flowing of these salt deposits produces faulting in the overlying sediments that affects the seafloor surface, creating long escarpments clearly visible in the bathymetry (Figs. 2, 7). The migration and deformation of the evaporites is responsible of the formation of furrows and swells observed in the relatively smooth and flat seafloor of the deep Algero-Balearic Basin (Figs. 2, 3).

Profile TM23 strikes WNW – ESE (Figs. 2a, 7) and runs from the Cabo de Gata Spur, across the continental slope and south Maimonides High, down to the Algero-Balearic Basin (Fig. 2a). The basement shows an irregular topography as in previous profiles. Worth noting is the section of basement located between the Cabo de Gata Spur and the Maimonides High that is affected by numerous faults defining a normal faulting block structuration (Fig. 7a). Along this profile, the pre-Messinian sedimentary sequences are
relatively thick (almost 0.8 s TWTT) and appear faulted and folded. At the westernmost part of the section, we have identified two high-angle faults with a normal displacement (km 5 – 7.5 in Fig. 7a) and a vertical strike-slip fault (km 10 in Fig. 5a), affecting the basement and sedimentary cover up to the seafloor, suggesting that they are probably active. Taking into account the present-day stress regime, the strike-slip fault will probably have a left-lateral motion. Between 12 - 15 km and 25 - 30 km, we observed the lateral scarps of two submarine landslides (Fig. 2). The southern flank of the Maimonides High appears buried and covered by sediments from units II and I. Towards the eastern part of the TM23 profile, the Algero-Balearic Basin is imaged. Similarly to profile TM24 (Fig. 6b), we recognize the evaporitic deposits within Unit II and halokinesis-related structures (Fig. 7b). The easternmost part of the profile shows gliding structures producing deformation of the uppermost sedimentary layers, and causing the rotation and faulting of units I and IIa-UU above the evaporite layers (IIb-MU) (Fig. 7b). These faults have listric geometry and are rooted at the base of Unit IIb-MU, which acts as a décollement (Fig. 7b). The faults reach up to the surface, creating a step-like relief that can be observed on the bathymetric map (Figs. 2, 3b). Just above the décollement there are salt welds that connect remains of the original evaporite layer (Fig. 7b).

5. Discussion

5.1. Nature of the basement

Igneous activity in the Alboran Basin was controlled by the convergence between the European and African Plates, and the collisional – subduction system that was created between them. It has been active from the Middle Miocene, when tholeiitic rocks were emplaced, till Lower Pliocene (e.g. Duggen, 2004, Booth-Rea et al., 2007). The Abubacer and Maimonides highs have been dredged (Fernández-Soler et al., 2000), and
the recovery of volcanic and plutonic rocks from the top of the highs confirms its igneous nature (e.g. Duggen et al., 2008). They belong to the Cabo de Gata volcanic province, where volcanic activity occurred between 11.7 and 6.6 Ma. Later, volcanism continued to the north in the Mazarron-Cartagena region, where dated samples indicate an age of 3 Ma (Duggen, 2008). On seismic line TM24 (Fig. 6b) and on the bathymetric maps (Figs. 2, 3b, 8) we recognize volcanic features, such as cones and ridges, along the Abubacer and Maimonides highs.

Although the igneous nature of the Abubacer High is confirmed, the prominent ridge topography is not only generated by volcanic processes but partially generated by thrust and associated anticline, as suggested by Giaconia et al. (2015). The line TM24 across the Abubacer ridge (Fig. 6a) does not show any evidence of volcanic aprons that are usually formed by volcanoclastic deposits during the growth of a volcanic edifice. Moreover, the infill of the Abubacer Basin shows syntectonic deformation, and although the faults responsible of the uplift of the Abubacer ridge are not visible, a kinematic analysis supports thrust faulting in the Abubacer High (Giaconia et al., 2015).

The Abubacer High shows evidence of erosion. It is affected by a system of gullies, especially on its northern part, and also by mass-wasting processes. On the other hand, the Maimonides High has a sharp topography, and does not display any evidence of mass transport deposits. These different erosion degrees may suggest that the Maimonides structures are younger, hypothesis also supported by the recovery of fresh glassy vesicular rocks from the top of the Maimonides High (Fernández-Soler et al., 2000).

5.2. Tectonic evolution
The driving force of the neotectonic activity in the Alboran Basin, and especially at the Palomares margin, is the convergence between the European and African plates. West of the Palomares margin, however, GPS data indicate other tectonic mechanisms (e.g. Koulali et al., 2011; Galindo-Zaldivar et al., 2015; Pérez-Peña et al., 2015) related to the detaching slab that affect the tectonic activity observed in the central and western Betics (Martínez-Martínez et al., 2006; Galindo-Zaldivar et al., 2015; Mancilla et al., 2015). The NW-SE trending convergence (4.5 – 5.6 mm/yr, e.g. Nocquet, 2012) between these two tectonics plates is accommodated along a series of active faults, such as the Eastern Betic Shear Zone (EBSZ) in SE Iberia and East Alboran Basin (e.g. Bousquet, 1979) (Fig. 1). This structure has a clear trace onshore and the PF segment runs along the coastline near Vera (Fig. 2). Although there are no proofs of its offshore continuation, some authors affirm that the PF continues at sea (e.g. Comas et al., 1999; Booth-Rea et al., 2004). Other hypothesis defends that the PF ends at the shoreline (e.g. Gràcia et al., 2006; Moreno et al., 2016), and deformation is instead accommodated along the reverse faults segments located at the southwestern termination of the Palomares Fault, at the northern and southern sides of the Cabrera anticline, and also at the E-W dextral-reverse Polopos Fault zone (Figs. 1, 2) (Giaconia et al., 2012, 2013). The high-resolution bathymetric and slope map of the shelf area as well as the MCS profiles (Figs. 2, 5, 6, 7); do not show any evidence of an offshore continuation of the Palomares Fault. The roughly coast-parallel scarps in the area between the inner and the outer shelf are likely of exogenic origin and related to the Pleistocene glacial-induced regressions and transgressions (Acosta et al., 2013; Lobo et al., 2015). Reverse instrumentally-recorded earthquakes have been recorded in the area (Stich et al., 2003, Giaconia et al., 2015), but there is not enough accuracy on the seismicity location to make a relationship between the earthquakes and these small scarps (Fig. 1). Thus, we suggest that the PF
ends near the shoreline at the westernmost end of Sierra Cabrera, or its slip greatly
decreases offshore, not forming characteristic deformational structures that could be
observed in the bathymetry or in the MCS profiles (Fig. 2, 3a).

Most of the faults imaged by the MCS profiles are considered as inactive, as they are
either covered by undeformed sediments of Units Ib and Ia (Fig. 5 between 10 – 15
km), or they affect only the basement and oldest sedimentary units (Fig. 6a at 30 km,
Fig. 7a between 13 – 17 km). Only a few faults located in the upper slope appear to
deform Plio-Quaternary deposits. The large majority of these faults were generated
during the basin opening extension, with normal fault kinematics. Units V and IV
appear largely confined to basement relief that was formed by normal faulting. The
strata from Unit V are rotated and locally form wedge-shaped deposits that onlap the
basement. We interpret unit V as syn-rift. Unit IV is also interpreted as a syn-rift unit
due to its thickness variation, increasing at the deepest part of fault-controlled local
depocenters, and with an onlap configuration against the top of Unit V. However, Unit
IV is considerably more widespread than Unit V and filled the fault-controlled
depocenters, possibly indicating a late syn-tectonic character. Unit III presents parallel
reflections, and its thickness variations are probably related with the Messinian Salinity
Crisis erosion. Reflections in Unit II are parallel and the unique deformation observed
in this unit is related to halokinesis processes and not to tectonic structures, suggesting
that the deposition of this unit took place once the rifting phase finished.

The geometry of strata and relative position of the units, support that only a few of the
normal faults formed during rifting may have been later reactivated with reverse or
oblique slip. We interpret that the faults are reactivations of inherited structures because
compressive deformation is especially visible affecting pre-Pliocene sediment, possibly
because those sediments were compartmentalized in narrow bodies and fault
reactivation strengthens the discontinuous character of the deposits of older units. However, the reason why reactivation has occurred only at a few structures from those formed during rifting, and mainly along structures in the middle and upper continental slope is unclear, and perhaps it is related to a priori favorable orientation.

Along line TM25 only one large structure appears active, cutting from depth into the Pliocene and perhaps Quaternary section in the upper slope (km 2.5 – 4.5 in Fig. 5). The deformation is limited and does not seem to cut the seafloor, or it is below the data resolution of our bathymetric and seismic data. However, the seafloor shows some small-scale relief that appears to indicate some recent deformation. The structure is then possibly active but accommodates minor ongoing deformation, or alternatively, the structure corresponds to a growth fault with offset decreasing towards the surface.

Line TM24 images the deformation associated to the relief of the Abubacer Ridge that is a structure with at least part of its relief related to contraction (Giaconia et al., 2015) (Fig. 6a). The relief of the igneous basement is related to a landward dipping thrust fault cutting from deep under the structure (Giaconia et al. 2015). The sediment in the small depocenter overlaying the basement (Abubacer Basin in Fig. 6a), formed in the hanging wall to the fault, shows that the entire sequence seems gently folded, but that folding has probably attenuated since Early Pliocene. Pre-Messinian Unconformity units appear folded. The thickness of Unit III seems to be constant, so we inferred that the deposit of this unit is still previous to the beginning of the thrust. Unit Ib (Early Pliocene) shows evidence of folding, and presents parallel internal reflectors and its thickness increases slightly towards the depocenter. Unit Ia (Late Pliocene – Quaternary) presents almost a constant thickness, onlapping above Unit Ib (km 11 – 22 in Fig. 6a). Thus, it is likely that the thrust initiated in Late Messinian or Earliest Pliocene, and during the Pliocene and Quaternary slip rates decreased. It is also unclear how much ongoing deformation
may currently occur in the Abubacer High, but instrumentally recorded earthquakes showing reverse mechanisms occur near it (Giaconia et al., 2015).

Line TM23 images also two major structures that seem currently active in the upper slope (km 5 - 8 in Fig. 7a). The narrow anticline fold deforming up to Unit Ia (km 6 in Fig. 7a) may have been formed by a steep blind thrust or may alternatively correspond to a positive flower structure of a strike-slip fault, similar to faults onshore in the neighboring region. The high dip angle of the fault trace and the stratigraphic relations between the oldest units (like the fan shape of Unit III, km 6 – 10 in Fig. 6a), drive us to think that these faults were normal faults. The compressive deformation associated with the younger units, like the anticline previously described, supports that these faults have been reactivated as compressive structures. Here again, deformation is higher in Units II-IV, still fairly clear in Unit Ib and subdued in Unit Ia. The deformation structures support an initiation of the contraction in the Late Messinian – Early Pliocene with the largest slip at the time and a much slower slip in Pleistocene time. This decrease in slip rate is supported by the relatively constant sedimentation rates averaged across the Plio-Quaternary time (calculated sedimentation rates from ODP Site 978 are 154 m/Ma for the Upper Miocene, 120 m/Ma for the Lower Pliocene, 111 m/Ma for the Upper Pliocene and 127 m/Ma for the Pleistocene) (Braga and Comas, 1999). These ages are coherent with the ages and rifting phases described for the opening of the western Mediterranean and its later compressive reorganization, which is proposed to have started at around 8 Ma (late Tortonian) (e.g., Mauffret et al., 2007; Billi et al., 2011; Medauri et al., 2012, 2014; Giaconia et al., 2015).

The seafloor mapping is of sufficient resolution to detect structures ~10 m high in the slope and a few meters high in the shelf so that we interpret that the relief data does not support the presence of significantly deformed structures and thus, current deformation
is very limited in the study region. This is also supported by the images of the subsurface structures described above from the three seismic profiles. So, we infer that ongoing deformation due to the Eurasian - African plate convergence has shifted from the older offshore structures and it is currently mainly accommodated by structures onshore and few uppermost slope faults offshore along this margin. Instead, we proposed that the ongoing deformation in the deep basin offshore the Palomares margin should be accommodated at the evaporites layer, due to gravity driven salt-tectonics.

5.3. Halokinesis

During Messinian times, the connection between the Atlantic Ocean and Mediterranean Sea through the Straits of Gibraltar was closed, resulting in what has been referred as the “Messinian Salinity Crisis” (MSC), which occurred about 5.96-5.33 Ma ago (e.g. Krijgsman et al., 1999; Duggen et al., 2003; García-Castellanos et al., 2009, 2011; Roveri et al., 2014). The ductile deformation processes affecting evaporite deposits are responsible for the sediment deformation observed in some of the MCS profiles. This thin-skinned deformation is mainly driven by gravity (Sage et al., 2005; Hudec and Jackson, 2007; Lofi et al., 2011). Halokinesis is also the responsible of the swell and furrow topography present across the Algero-Balearic Basin (Figs. 2, 3 and 8). Furthermore, this topography has been imaged in the nearby areas of the Palomares continental margin (Camerlenghi et al., 2009; Acosta et al., 2013). This characteristic deformation of the sediments is due to the displacement of the evaporite layer in the deeper part of the basin, where the salt flow produces salt anticlines and pillows (Camerlenghi et al., 2009). Other typical structures derived from halokinesis have been identified. On profile TM24, we imaged a salt dome. The upward flow of this evaporite accumulation has generated faulting and block rotation in the above sediments that mainly affects Units Ib and Ia (Fig. 6b). We interpret that this is an active process, as
evidenced by seafloor deformation (Figs. 6b and 8). In the eastern section of the TM23 profile (Fig. 7b) we observe structures related to gliding processes. Similar situations have been described across the whole Mediterranean Sea (e.g., Sage et al., 2005; Lofi et al., 2011). Sedimentary Units Ib and Ia have been gravitationally sliding over a thin salt layer, which base acts as a weak décollement surface. This seismic reflector shows strong inverse polarity (Fig. 7b), which defines the base of the evaporites. Gliding is probably due to their low coefficient of friction and the flow of evaporites from the shallower part of the continental slope into the deeper basin. The migration of salt has generated normal faulting and block rotation in the uppermost sediments corresponding to Units Ib and Ia (Fig. 7b). The different sedimentary load of the sediment wedge at the continental slope have been proposed as a possible triggering mechanism (Gaullier and Vendeville, 2005; Camerlenghi et al., 2009). We interpret that this process is also active, as faulting is affecting the seafloor. On both profiles, where Subunit IIb-MU is identified, Unit Ib is composed of parallel reflectors without internal deformation, while Unit Ia is onlapping above Ib where halokinesis phenomena occurred, and its thickness increases toward the deeper part of the basin at the top of Unit Ib (Fig. 6b, 67-70 km). On the basis of this internal structuration, we propose that the salt movement at the Palomares continental margin began after the deposition of Unit Ib, and that Unit Ia is coetaneous with this activity (Late Pliocene – Quaternary).

5.4. Sedimentation and mass wasting processes

In the area, sedimentary fluxes are conditioned by two main processes: a relatively constant hemipelagic sedimentation, and a high-energy sedimentation driven by slope failures and turbidite currents. Regarding the occasional sedimentation related to bottom currents and mass-transport processes, the role of the canyons as sediment pathways feeding sediments from the shelf to the abyssal plains is fundamental. These canyons
are highly erosive structures (Fig. 2), which location in origin may be related to a fault in depth (Fig. 6). In the cases of the Cabo de Gata Canyon (Fig. 6b) and the Almanzora-Alias-Garrucha Canyon (Fig. 6a), it is clear that they are related to a fault trace at depth (Fig. 6b). These faults are most likely inactive, because they are not affecting the Pliocene – Quaternary sedimentary sequence. We suggest that the other canyons of this area may also follow fault traces as weakness areas for preferential erosional pathways, at least at their early stages. Due to the complex topography of the margin, the present pathways are also determined by topographic factors, such as topographic highs. The mass-wasting processes affecting the canyon flanks, the narrow crests between them, and the depositional sedimentary bedforms recognized in some of the canyons, such as the “cyclic steps” (e.g. Cartigny et al., 2011) identified at the mouth of the Almanzora-Alias-Garrucha canyon (Figs. 2b, 8), are indicative of current activity along these three canyons, which is consistent with recent studies (Perez-Hernandez et al., 2014).

The inner shelf corresponds to an erosional platform covered by a thin layer of Pleistocene sediments. Along this part of the shelf, sedimentary bars with parallel to sub-parallel orientation relative to the shoreline are recognized (Fig. 8). Sedimentary bars with similar orientation have been recognized at the northern adjacent area, the Mazarrón margin (Acosta et al., 2013). In this area, bars parallel to the shoreline have been interpreted as relict structures constructed during the Holocene transgression by alongshore currents (Acosta et al., 2013). At the Mazarrón margin active sediment bars have been identified. They are related to southwest flowing (the Liguro-Provençal-Catalan or North Current), and have a SW-NE orientation (Acosta et al., 2013). Due to the proximity between the two areas and the similarity between the processes described, we inferred that the sediment bars recognized at the Palomares margin with a parallel-subparallel orientation relative to the shoreline are relict. The relief of the outer-shelf is
mainly dominated by erosive processes, such as slope failures, tributary valleys, gullies, and canyon erosion (Fig. 8). It is still unclear how fast these erosional processes are and whether they can obliterate relief formed by slow-slip active faulting common in the region. Recent sedimentation at the Southeastern Iberian shelf is poorly known (Lobo et al., 2014). The suspended fine sediments travel with the dominant current (Liguro-Provençal-Catalan or North Current) from North to South. At the southernmost area, the abrupt shelf configuration favors the sediment capture by submarine canyons, and they are exported to deeper parts of the basin (Lobo et al., 2014). Sedimentary processes have been identified at the deeper part of the continental slope and the Albero-Balearic Basin, such as the sedimentary lobe covering the southeastern flank of the Polacra Ridge (Figs. 2a, 8). A large number of mass transport deposits have been recognized (Fig. 8), and possibly different triggering mechanisms should be considered. Seismic activity may be a plausible one, although is scarce along the margin. Sediment loading has also been proposed as a trigger mechanism (Acosta et al., 2013), although our data does not provide evidences to support it.

6. Conclusions

The Palomares margin is characterized by a complex relief mainly generated by the presence of three large highs, and three deeply eroded submarine canyons in between them. We have recognized the following areas: (i) the continental shelf, (ii) the continental slope and (iii) the Albero-Balearic Basin. Active processes and tectonic structures present on each area are different, and determine their specific seafloor morphology.

The shelf may be divided into the inner and outer shelf. These two levels are separated by an escarpment that is probably related to a previous cliff cut during the Pleistocene.
glacially induced regressions and transgressions (Acosta et al., 2013, Lobo et al., 2015).

The continental slope is characterized by the presence of highs and canyons. Three highs are identified: (i) the Aguilas High, (ii) the Abubacer High and (iii) the Maimonides High. They are affected by erosive and mass-wasting processes, although we infer that the Maimonides High is probably younger than the remaining two due to its sharp relief. Three canyons are also mapped: (i) the Aguilas canyon, (ii) the Almanzora-Alias-Garrucha canyon and (iii) the Cabo de Gata canyon. At least the Cabo de Gata and Almanzora-Alias-Garrucha Canyon seem to follow possibly inactive fault traces as preferential erosion direction. The complex topography of this area has probably conditioned the later evolution of canyon pathways. They show evidence of erosional processes, with gully systems on their flanks, slope failures and erosive slide scars, and mass transport deposits. Finally, the geomorphology of the Algero-Balearic Basin at the Palomares margin area is mainly controlled by halokinesis processes related to the Messinian evaporites that formed a series of troughs and swells.

The bathymetry of the Palomares margin and the imaged structures with the MCS profiles reveal that the geomorphology of the area is controlled mainly by erosional and halokinesis processes. Faults identified in the MCS profiles do not seem to affect the seafloor, and deformation apparently decreases from the Messinian till present. We suggest that the ongoing deformation in the area is then accommodated by onshore structures. Offshore, there are only few active faults located at the uppermost slope.

The Palomares margin formed during the Early Tortonian. The rift stage continued in the margin till the Late Tortonian, when compressive deformation started to control the evolution of the area. This change in the dominant tectonic mechanism produced the reactivation, as thrusts and/or strike-slip faults, of the few previously extensional structures, mainly located at the continental slope.
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**Figure Caption**

**Figure 1.** Regional bathymetric map of the Alboran Sea constructed from digital grids released by SRTM-3, IEO bathymetry, GEBCO compilation and data acquired during our own cruises (Ballesteros et al., 2008, Gràcia et al., 2012). The location of the earthquakes that occurred in this area since 1916 are shown (grey dots). The maximum magnitude registered is Mw 5.2 (*Instituto Geográfico Nacional* catalog, available online at [http://www.ign.es/](http://www.ign.es/)). Historical events occurred in the area are also located. Main tectonic structures, as the faults belonging to the EBSZ (from north to south, Bajo Segura, Alhama de Murcia, Palomares and Carboneras faults) are displayed (Gràcia et al., 2012, Giaconia et al., 2015). Orange-filled dots correspond to wells location, including the scientific ODP-161 sites 977 and 988, and the industry well Andalucia-A1. The area bounded by the dotted light purple polygon shows the coverage of the new bathymetry acquired. The yellow rectangle depicts the area presented in Figure 2. Inset: Location of the Palomares margin (red rectangle) in SE Iberia.

**Figure 2.** (a) Colour shaded-relief bathymetric map of the Palomares margin (illumination from the NW) (see Figure 1 for location). Main geomorphological features and structures are depicted. MCS profiles presented in this work (TM23, TM24, and TM25) are located. Location of the close up figures (Figs. 3a, 3b) is depicted. (b) Slope map of the Palomares margin, same area as Figure 2a. The boundaries between the geomorphologic zones are located: black dashed line corresponds to the inner - outer shelf boundary, red dashed line corresponds to the outer shelf - continental slope boundary, grey dashed line corresponds to the eastern boundary of the continental slope, and pink dashed line corresponds to the East Alboran Basin and Algero-Balearic Basin boundary. MCS profiles are located.
Figure 3. Zooms of the colour shaded-relief bathymetric map of the northern and southern part of the Palomares margin, with isobaths every 50 m. Main structures described in the text are located. See Figure 2a for location. (a) Close up of Aguilas Canyon, Aguilas High and Almanzora-Alias-Garrucha Canyon area. (b) Close up of Abubacer High, Cabo de Gata Canyon and Maimonides High area.

Figure 4. (a) Ages and seismostratigraphic units identified in the TOPOMED MCS profiles, on (b) the continental slope and (c) the Algero-Balearic Basin. Units are as follows: Ia: Quaternary + Upper Pliocene, Ib: Lower Pliocene, II: Upper Messinian, III: Lower Messinian, IV: Upper Tortonian, V: Lower Tortonian, B: top of the basement. Unit II (Upper Messinian) has been subdivided in three sub-units (IIa, IIb and IIc) in the Algero-Balearic Basin, where evaporites are present. These units have been correlated with the ones proposed by Lofi et al. (2010). The M points out the Messinian unconformity, the boundary between Units II and Ib. This boundary is correlated with the TS/TES (Top Surface/Top Erosion Surface) in the areas where the salt layers exist. In these areas, the boundary between Units III and II is the BS/BES (Bottom Surface/Bottom Erosion Surface). See Figure 7 for location of MCS sections.

Figure 5. Time migration of profile TM25. Main structures and seismostratigraphic units are identified. Age of the units is defined in the caption of Figure 4. Vertical exaggeration is 5:1 taking into account the sound velocity in water (1500 m/s). Profile is located on Figure 2.

Figure 6. Time migration of profile TM24, northwestern (I) and southeastern (II) sections. Main structures and seismostratigraphic units are identified. The dashed line corresponds to the inferred fault trace responsible to uplifting the Abubacer High. Age
of the units is defined in the caption of Figure 4. Vertical exaggeration is 5:1 taking into account the sound velocity in water (1500 m/s). Profile is located on Figure 2.

Figure 7. Time migration of profile TM23, western (I) and eastern (II). Main structures and seismostratigraphic units are identified. Age of the units is defined in caption of Figure 4. Vertical exaggeration is 5:1 taking into account the sound velocity in water (1500 m/s). Profile is located on Figure 2. MCS sections showed at Figure 4 are located.

Figure 8. Geomorphological, tectonic and sedimentary map of the Palomares margin. Map is located on Figure 1.
Figure 1. Regional bathymetric map of the Alboran Sea constructed from digital grids released by SRTM-3, IEO bathymetry, GEBCO compilation and data acquired during our own cruises (Ballesteros et al., 2008, Gracia et al., 2012). The location of the earthquakes that occurred in this area since 1916 are shown (grey dots). The maximum magnitude registered is Mw 5.2 (Instituto Geográfico Nacional catalog, available online at http://www.ign.es/). Historical events occurred in the area are also located. Main tectonic structures, as the faults belonging to the EBSZ (from north to south, Bajo Segura, Alhama de Murcia, Palomares and Carboneras faults) are displayed (Gracia et al., 2012, Giaconia et al., 2015). Orange-filled dots correspond to wells location, including the scientific ODP-161 sites 977 and 988, and the industry well Andalucia-A1. The area bounded by the dotted light purple polygon shows the coverage of the new bathymetry acquired. The yellow rectangle depicts the area presented in Figure 2. Inset: Location of the Palomares margin (red rectangle) in SE Iberia.
Figure 2. (a) Colour shaded-relief bathymetric map of the Palomares margin (illumination from the NW) (see Figure 1 for location). Main geomorphological features and structures are depicted. MCS profiles presented in this work (TM23, TM24, and TM25) are located. Location of the close up figures (Figs. 3a, 3b) is depicted. (b) Slope map of the Palomares margin, same area as Figure 2a. The boundaries between the geomorphologic zones are located: black dashed line corresponds to the inner - outer shelf boundary, red dashed line corresponds to the outer shelf - continental slope boundary, grey dashed line corresponds to the eastern boundary of the continental slope, and pink dashed line corresponds to the East Alboran Basin and Algero-Balearic Basin boundary. MCS profiles are located.
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