Magnetobiochronology of lower Pliocene marine sediments from the lower Guadalquivir Basin: insights into the tectonic evolution of the Strait of Gibraltar area

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ABSTRACT

The Gibraltar Arc is a complex tectonic region, and several competing models have been proposed to explain its evolution. We study the sedimentary fill of the Guadalquivir Basin to identify tectonic processes occurring when the re-opening of the Strait of Gibraltar led to the re-establishment of Mediterranean outflow. We present a chronostratigraphic framework for the lower Pliocene sediments from the lower Guadalquivir Basin (SW Spain). The updated chronology is based on magnetobiostratigraphic data from several boreholes of this basin. Our results show that the studied interval in the La Matilla core is early Pliocene, and further provide better constraints on the sedimentary evolution of the basin during this period. Migrating depositional facies led to younger onset of sandy deposition basinward. In the northwestern passive margin, a 0.7-Ma period of sedimentary bypass related to a sharp decrease in sedimentation rates and lower sea levels resulted from the tectonic uplift of the forebulge. In contrast, high sedimentation rates with continuous deep marine sedimentation are recorded at the basin center due to continuous tectonic subsidence and west-southwestward progradation of axial depositional systems. The marginal forebulge uplift,

continuous tectonic basinal subsidence, and southward progradation of clinoforms in the early Pliocene can be explained by the pull of a lithospheric slab beneath the Gibraltar Arc as the Strait of Gibraltar opened. These findings are, to our knowledge, the first reported sedimentary expression of slab pull beneath the Betics, related to the opening of the Strait of Gibraltar after the Messinian salinity crisis.

INTRODUCTION

The lower Guadalquivir Basin, located in SW Spain, is the westernmost sector of the Guadalquivir foreland basin of the Betic Cordillera (Fig. 1). This area was part of the marine connection between the Atlantic Ocean and Mediterranean Sea during the late Miocene (Martín et al., 2009, 2014; Braga et al., 2010; Pérez-Asensio et al., 2014; Flecker et al., 2015). The complete sedimentary record of the basin encompasses the Miocene-Pliocene boundary, when the Zanclean flood through the Strait of Gibraltar purportedly ended the Messinian salinity crisis (Hsü et al., 1973; García-Castellanos et al., 2009; Roveri et al., 2014). This important geological event was also responsible for the re-establishment of Mediterranean outflow water (MOW), which regulates Atlantic thermohaline circulation and climate in the northern hemisphere (Reid, 1979; Bigg and Wadley, 2001; Özgökmen et al., 2001; Bigg et al., 2003). Weak MOW influence during the earliest Pliocene in the Gulf of Cádiz has been reported (van der Schee et al., 2016; García-Gallardo et al., 2017). However, the consistent influence of MOW on the North Atlantic Ocean started at approximately 4.5 Ma when MOW flowed as an intermediate and deep current (Hernández-Molina et al., 2013, 2014, 2016). These events have been linked to the vertical tectonic motions of the crust underlying the Atlantic-Mediterranean corridors around the Miocene-Pliocene boundary, but competing models proposed to explain the evolution of this region remain controversial (Calvert et al., 2000). During the last decade, discussion has focused on the presence in seismic tomography images of a lithospheric slab inherited from the Cenozoic convergence between Africa and Iberia beneath the Gibraltar Arc (Spakman and Wortel, 2004). While the polarity of the subduction that caused this slab remains debated (e.g., Chertova et al., 2014), its present geometry has been interpreted to reveal ongoing lithospheric tearing beneath the Betics (Garcia-Castellanos and Villaseñor, 2011; Ruiz-Constán et al., 2011), and this activity may have been responsible for the vertical motions that controlled the connectivity between the Mediterranean Sea and the Atlantic Ocean during the late Miocene.

Extensive research on the stratigraphy, sedimentology, paleoclimate, and paleoenvironmental evolution of the lower Guadalquivir Basin during the late Miocene, Pliocene and Quaternary has been performed on both outcrops and cores (e.g., Civis et al., 1987; Sierro et al., 1993, 1996; Ledesma, 2000; Gläser and Betzler, 2002; González-Delgado et al., 2004; Salvany et al., 2011; Aguirre et al., 2015, Jiménez-Moreno et al., 2015). Nonetheless, a complete record of the Miocene-Pliocene transition in the basin has not been investigated until now despite the important paleoceanographic implications. The sedimentary basin fill, which progressively thickens to the south, reaching a thickness of more than 2.6 km near the Betic front (Iribarren et al., 2009), is poorly exposed in outcrops. Therefore, to analyze complete sedimentary sequences, including the Miocene-Pliocene boundary, the only option is to work with sediment cores. To date, the best sedimentary record of this interval in the basin is the Montemayor-1 (MT) core, which has been extensively studied (Larrasoaña et al., 2008, 2014; Pérez-Asensio et al., 2012, 2013, 2014; Jiménez-Moreno et al., 2013; van den Berg et al., 2015). This core is located in a marginal position and includes a complete Messinian record; however, the lower Pliocene record shows discontinuous sedimentation with a sedimentary hiatus of at least 100 kyr at its onset (van den

Berg et al., 2015). Exploratory boreholes in the central portion of the basin have retrieved the complete upper Miocene-lower Pliocene sedimentary sequence (Ledesma, 2000; Sierro et al., 2000). Although biostratigraphy and cyclostratigraphy have been investigated in these boreholes, detailed sedimentary facies and paleoceanographic analysis have not been performed yet. The continuous La Matilla (LM) core was collected from an intermediate position between the northern passive margin of the basin and the central sector. Preliminary data indicate that this core encompasses the uppermost part of the Gibraleón Formation and an expanded Huelva Formation (Salvany et al., 2011). This core is, according to the chronology of these formations (Larrasoaña et al., 2008, 2014; Pérez-Asensio et al., 2012, 2013, 2014; Jiménez-Moreno et al., 2013; van den Berg et al., 2015), a suitable candidate for encompassing a complete record across the Miocene-Pliocene boundary and the early Pliocene. This record could provide valuable information on the vertical tectonic movements and related changes in the sedimentary record (e.g., Soria et al., 1998; Guerra-Merchán et al., 2014) for an important time period for which only scarce information from the Guadalquivir Basin is currently available (e.g., Garcia-Castellanos et al., 2002; Pérez-Asensio et al., 2012, 2013).

In this study, we provide the first detailed chronostratigraphic framework for the LM core based on magnetostratigraphic and biostratigraphic methods, along with information from other cores in the area to assess the tectono-sedimentary evolution of the basin and the neighboring Gibraltar Arc during the late Miocene and early Pliocene.

GEOLOGIC SETTING

The study area is located in the western sector of the Guadalquivir Basin (SW Spain) (Fig. 1). This basin is an ENE-WSW-elongated foreland basin bounded to the N by the passive Iberian

Massif and to the S by the active Betic Cordillera (Sanz de Galdeano and Vera, 1992; Braga et al., 2002; González-Delgado et al., 2004). This region was the Atlantic side of the North Betic Strait until its definitive closure in the earliest Tortonian (Aguirre et al., 2007; Martín et al., 2009, 2014; Braga et al., 2010). Subsequently, it became an open and wide Atlantic-linked embayment (Martín et al., 2009). The sedimentary fill of the Guadalquivir Basin consists of olistostromic deposits from the Betic Cordillera and several marine and continental units, which range in age from earliest Tortonian to Holocene (Sanz de Galdeano and Vera, 1992; Aguirre et al., 1995; Riaza and Martínez del Olmo, 1996; Sierro et al., 1990, 1996; González-Delgado et al., 2004; Salvany et al., 2011; Rodríguez-Ramírez et al., 2014, 2015).

In the study area, Neogene sediments have been divided into eight marine and continental lithostratigraphic units (Civis et al., 1987; Salvany et al., 2011). The lowermost unit is the Niebla Formation, which is composed of marine-continental carbonate-siliciclastic sediments deposited during the late Tortonian (Civis et al., 1987; Baceta and Pendón, 1999). This unit unconformably onlaps the Paleozoic-Mesozoic basement of the Iberian Massif. The second unit, deposited during the latest Tortonian-Messinian according to planktonic foraminifera and calcareous nannoplankton (Flores, 1987; Sierro et al., 1993), is the Gibraleón Formation (Civis et al., 1987). This marine unit consists of greenish-bluish clays including glauconitic silts at the base. The contact between the Niebla and Gibraleón formations is interpreted as a condensation level (Sierro et al., 1996; Abad, 2010), associated with a maximum flooding event. In the Seville province, the Gibraleón Formation is overlain by the so-called Transitional Unit, which includes transitional facies such as a sequence of alternating silts and sands that likely encompass the Miocene-Pliocene transition (De Torres, 1975; Mayoral and González, 1986-87; Muñiz and Mayoral, 1996). These facies are missing in more marginal areas of the lower Guadalquivir Basin, such as the Huelva province. The third unit is the Huelva Formation (Civis et al., 1987), which includes marine silts and sands deposited during the early Pliocene. A glauconitic layer and an unconformity are present at the base of the formation in marginal areas (Huelva province). The fourth unit, the Bonares Formation, unconformably overlies the Huelva Formation. Although this formation has been assigned to the upper Pliocene on the basis of its stratigraphic position (Mayoral and Pendón, 1986-87), its correlation with magnetostratigraphically dated sequences indicates a lower Pliocene age (Salvany et al., 2011). There are four continental units lying above these four marine units (Salvany et al., 2010, 2011). The first continental unit is the upper Pliocene-Lower Pleistocene Almonte Formation, which includes gravels and sands interpreted as proximal-alluvial deposits. The proximal sediments of the Almonte Formation are referred to as the High Alluvial Level (Pendón and Rodriguez-Vidal, 1986-87; Salvany et al., 2011). Above the Almonte Formation, the second continental unit, the Lebrija Formation, consists of sands and gravel beds with brown and greenish clays in the lower part and gravels and sand in the upper part. This unit was interpreted as distal-alluvial sediments deposited during the late Pliocene-late Pleistocene. The last two units are the continental Abalario Formation and the continental-estuarine Marismas Formation. These units are composed of eolian sands and alluvial-estuarine dark grey-brown clays, respectively. Their ages range from the uppermost Pleistocene to the Holocene.

TECTONIC SETTING

Post-Cretaceous relative movement between the African and Eurasian plates controlled the tectonic evolution of the study area including the Guadalquivir foreland Basin and Betic Cordillera, which is the active front of the basin (Garcia-Castellanos et al., 2002). Three tectonic domains were differentiated in the Betics-Guadalquivir system by the onset of the Neogene

(Balanyá and García-Dueñas, 1987): 1) the External Zones of the Betic Cordillera; 2) the Flysch Units that consist of allochthonous sediments; and 3) the Internal Zones of the Betic Cordillera.

The Guadalquivir Basin was formed during the NW emplacement of the Betic front, which occurred progressively from the E to the W of the basin due to the oblique convergence of the African and Eurasian plates (Serpelloni et al., 2007). The emplacement of the outermost unit of the Betic Front (the so-called olistostromic or Guadalquivir units, see Pérez-Valera et al., 2017) took place during the middle Tortonian (ca. 9.5 Ma) in the eastern part of the basin (García-García et al., 2014; Pérez-Valera et al., 2017), whereas it occurred during the late Tortonian (ca. 7.5 Ma) in the western part (Fig. 1) (Ledesma, 2000). This is deduced from the age of the sedimentary sequence that postdates major emplacement of the Guadalquivir units, which ranges from <9.5 Ma in the easternmost part of the basin (García-García et al., 2014; Pérez-Valera et al., 2017) to ca. <7.5 Ma (Sierro et al., 1996; Ledesma, 2000; Larrasoaña et al., 2014; van den Berg et al. 2016, 2017) in its westernmost part. The gradual emplacement of the Betic front towards the NW is related to westward rollback and drift of a lithospheric slab under the Betic-Rif orogen (Vergés and Fernàndez, 2012; Crespo-Blanc et al., 2016). The former tectonic process led to progressive subsidence towards the NW along the Guadalquivir Basin during the Tortonian. During the Messinian, subtle uplift of the Betic front promoted progressive northward migration of turbidite depositional systems along the axis of the basin (Sierro et al., 1996; Ledesma, 2000). The Pliocene-Quaternary period is characterized by uplift of the forebulge at the northern basin margin and subsidence in the center of the basin (Salvany et al., 2011). A possible mechanism accounting for the forebulge uplift might be viscous relaxation of the lithosphere after the orogeny (Garcia-Castellanos et al., 2002).

MATERIAL AND METHODS

Lithostratigraphic Methods

The LM core is a 276-m-long continuous core originating from the western lower Guadalquivir Basin (ED50 UTM coordinates zone 29N, X: 702,052, Y: 4,116,679; Z: 47) (Fig. 1; Table 1). It was drilled by the IGME (Geological and Mining Institute of Spain) in 2006 using a direct circulation rotary rig with continuous core sampling. The lithostratigraphy was analyzed to describe the different facies on the basis of grain size, texture, sorting, color, fossil content, and presence of nodules and organic matter debris. Then, each facies was assigned to one of the formations outcropping in the lower Guadalquivir Basin area.

Paleomagnetic Methods

The magnetic stratigraphy of the LM core is based on the study of 136 samples that were drilled perpendicular to the borehole sections using an electric-powered drill. The uppermost 28.7 m of the core were not sampled due to the unconsolidated and noncohesive nature of the sediments. This sampling strategy gives an average resolution of one sample per 1.65 m throughout the studied interval between 28.7 and 278.5 m core depth.

Paleomagnetic analyses were conducted at the Paleomagnetic Laboratory of the Institute of Earth Sciences Jaume Almera (CCiTUB-CSIC) in Barcelona, Spain, using a 2G superconducting rock magnetometer with a noise level of $< 7 \times 10^{-6}$ A/m (significantly lower than the magnetization of the measured samples). Thermal demagnetization of the samples, which has proven more effective than alternating-field methods in previous studies of Miocene and Pliocene sediments from the Guadalquivir Basin (Salvany et al., 2011; Larrasoaña et al., 2008, 2014), was conducted using a MMTD–80 furnace in the same laboratory. Previous paleomagnetic studies were used to optimize the demagnetization steps (e.g., 8 to 12 steps at intervals of 100°C, 50°C, 30°C and 20°C to a maximum temperature of 680°C) and accurately calculate the magnetization directions by minimizing heating and the formation of new magnetic phases in the oven. Stable Characteristic Remanent Magnetization (ChRM) directions were calculated by means of Principal Component Analysis (Kirschvink, 1980) after they were identified through visual inspection of orthogonal demagnetization plots (Zijderveld, 1967) using the VPD software (Ramón et al., 2017). Original demagnetization data and details on the calculation of ChRM directions are available in the Appendix 1.

Rock magnetic experiments were also performed at the Paleomagnetic Laboratory of the CCiTUB-CSIC to constrain the type, concentration, and grain size of magnetic minerals in the studied sediments. These parameters can be used to help identify the origin and reliability of the ChRM. Selected samples representative of the different sediment types in the LM core were analyzed following the method of Lowrie (1990), which involves the thermal demagnetization of a three-axis isothermal remanent magnetization (IRM) imparted at fields of 1.2, 0.3 and 0.1 T and enables the identification of the main magnetic carriers. Before this, we measured different bulk magnetic properties of the sister specimen of every fourth sample used in the paleomagnetism analyses. We measured the low-field mass-specific magnetic susceptibility (χ), the anhysteretic remanent magnetization (ARM) and two IRMs imparted at 0.3 T (IRM0.3 T) and 1.2 T (SIRM). χ was measured with a Kappabridge KLY-2 (Geofyzica Brno) susceptibility bridge using a field of 0.1 mT at a frequency of 470 Hz. AF demagnetization and ARM experiments were conducted using a D-Tech 2000 (ASC Scientific) AF demagnetizer. The ARM was applied along the Z-axis of the samples with a dc bias field of 0.05 mT parallel to a peak AF of 100 mT. IRM0.3 T and

SIRM were imparted using an IM10–30 pulse magnetizer (ASC Scientific). All magnetic properties were normalized by the weight of the samples. We used different magnetic properties and interparametric ratios to assess variations in the type, concentration, and grain size of magnetic minerals in the studied samples (Evans and Heller, 2003; Liu et al., 2012).

Biostratigraphic Methods

A total of 21 samples, 50 g each, were used for biostratigraphic dating based on planktonic foraminifera (Fig. 2). The samples were washed over a 63 µm sieve and dried at 40°C in an oven. Then, samples were split into equal subsamples using a microsplitter. The subsamples were dry sieved over a 125 µm sieve, and 200-300 planktonic foraminifera were counted and identified at the species level. Relative abundances (%) were calculated based on foraminiferal counts. Only eleven marker species with biostratigraphic significance were selected for the biostratigraphic dating of the core. The presence, disappearance, last common occurrence (HcO = highest common occurrence) and last occurrence (HO = highest occurrence) of these biomarker species were used in order to constrain the age of the core sediments (Fig. 2). The most representative planktonic foraminifer species with biostratigraphic value were imaged with a Scanning Electronic Microscope (SEM) at the Department of Earth Sciences, University of Geneva (Switzerland) (Fig. 3).

LITHOSTRATIGRAPHY, MAGNETOSTRATIGRAPHY AND BIOSTRATIGRAPHY OF THE LA MATILLA CORE

Lithostratigraphy

The studied core consists of 276 m of marine and continental sediments from 4 formations and transitional facies comprising the sedimentary fill of the basin in this area (Fig. 2) (Sierro et al., 1996; González-Delgado et al., 2004).

The lithological changes are very subtle in the lower part of the sequence, which begins with 66 m of massive bluish grey clays and minor brownish grey silt levels with abundant foraminifera, small fragments of bivalve, echinoderm and gastropod shells (Figs. 2, 4), scattered ferruginous levels and, organic matter remains. These clayey facies can be assigned to the Gibraleón Formation (Civis et al., 1987). Upwards, the sequence consists of 75.5 m of brownish to bluish grey silts and bluish grey clays with sporadic shells and 3 centimeter-scale silty sand layers at 177-174.7 m, 164.9-164 m, and 148-144.5 m (Fig. 2). The sandy levels include grey fine-grained silty sands with sparse bioclasts and foraminifera. Based on the facies features, the complete interval (210 to 134.5 m) can be assigned to the transitional facies cropping out in central areas of the basin (Mayoral and González, 1986-87; Muñiz and Mayoral, 1996). These transitional facies are overlain by 71.5 m of fine-grained grey sands with several intercalations of silts (Figs. 2, 4). In general, these facies are composed of grey silty sand with a fine grain size and contain foraminifera and small fragments of shells. The silty intercalations (105.0-95.7 m) consist of grey silt and fine-grained silty sand with foraminifera and small fragments of shells. These facies share similar features with the Huelva Formation (Civis et al., 1987); consequently, they are assigned to this formation in the studied core (Fig. 2). Above, 34.3 m of silty sands with several intercalations of silts and gravels are found. Thin intervals of concentrated shells occur between 63 and 51.9 m (Figs. 2, 4). A level with a mixture of clasts up to 4 cm in size, shell fragments, and grey sand is located from 51.9 to 51.5 m. Another gravel-bearing level (47.0 to 46.2 m) contains well-sorted clasts (up to 2.5 cm in diameter) and grey silty sand (fine to medium grain size) with small fragments of marine shells and has a sharp basal contact (Figs. 2, 4). The section from 46.2 to 30.9 m contains grey silty sands with a fine to medium grain size and rare well-rounded gravels that are less than 1 cm in diameter. The sand contains abundant scattered marine fauna shells that reach several centimeters in diameter, including bivalve, gastropod, and scaphopod shells. The final interval of this section features grey silty sand with a fine grain size and small black organic matter debris (30.9-28.7 m). These sediments between 63 and 28.7 m exhibit the typical facies of the Huelva Formation, but also showing a coarsening upwards trend that is characteristic of the Bonares Formation (Mayoral and Pendón, 1986-87). Overlying this formation, there is a 14.2-m-thick interval of fluvial channel gravels and sands with a sharp surface at the base. In the interval from 28.7 to 24.0 m, loose and well-graded gravels and coarse grey sand dominate (Figs. 2, 4). The gravel clasts are well rounded and up to 6 cm in diameter. Loose grey, medium- to coarse-grained sand bearing some small gravel clasts occurs in the following interval from 24.0 to 18.0 m. The next interval (18.0 to 16.5 m) includes grey clayey sand with a medium to coarse grain size. The interval from 16.5 to 15.7 m is characterized by nodules of reddish ferruginous sand that are 3 to 5 cm of diameter scattered in grey mud. The topmost interval of this section (15.7 to 14.5 m) features dark grey clayey sand. This entire section (28.7 to 14.5 m) can be interpreted as continental facies belonging to the Almonte Formation (Salvany et al., 2011). The core ends with 14.5 m of loose and well-sorted fine- to medium-grained sand that is white, grey, yellow, and/or red in color (Figs. 2, 4). This interval has a sharp basal contact (unconformity) at the base and can be assigned to the Abalario Formation, which consists of eolian deposits (Salvany and Custodio, 1995).

Paleomagnetic Data and Magnetostratigraphy

In most samples, we identified a low-temperature component, labeled A, that is unblocked below 100–225ºC. This component typically shows shallow directions that are broadly perpendicular to the borehole sections (Fig. 5). This component is interpreted to record a present-day field overprint and a magnetization acquired during paleomagnetic drilling. In approximately 73% of the samples, an additional stable component that unblocks between 100–225ºC and up to 500ºC was identified. This component is interpreted as the ChRM and can be divided into three types on the basis of quality. Type 1 ChRMs are well-defined linear trends that enable accurate calculation of their directions (Fig. 6a, b, e, f). Type 2 ChRMs are characterized by either less-developed linear trends or incomplete demagnetization due to the growth of new magnetic minerals in the oven yet provide reliable polarity determinations by fitting incomplete linear trends to the origin in demagnetization plots (Fig. 6c, d). Type 3 ChRMs are characterized by lower intensities and display clustered directions that provide ambiguous polarity determinations (not shown). Quality types 1, 2, and 3 represent approximately 22, 46 and 31% of the ChRM directions, respectively.

The ChRMs show both positive and negative inclinations. The mean of the negative ChRM directions is –47.6º ± 17.1º, slightly steeper than the mean of the ChRM normal directions (+38.4º ± 17.6º). Overall, both positive and negative mean ChRM directions are statistically consistent with the expected inclination for the studied site (approximately ±50º) despite being slightly shallower. Therefore, the ChRMs likely provide a reliable record of the polarity reversals of the geomagnetic field. For the sake of quality and because the azimuth of the borehole is unknown, only type 1 and 2 ChRM inclinations have been used to identify polarity intervals in the studied core (Fig. 7a). Each polarity interval has been determined using at least two consecutive type 1 or 2 samples. The most conspicuous pattern of the LM polarity sequence is a long normal polarity interval in the middle of the core (labeled N2) between two reverse magnetozones (R1 and R2) (Fig. 7a). The paleomagnetic inclinations in R1, N2 and R2 are very similar to the expected inclination at the studied site. In contrast, the upper part of the core is characterized by the rapid alternation of positive and negative inclinations, a spurious shallowing of paleomagnetic inclinations and a significant increase in the sand fraction in the studied sediments (Fig. 7a, e). Thus, the recording fidelity of paleomagnetic inclinations in the upper part of the core is compromised and is therefore considered of uncertain polarity. The lowermost sample of the core has a positive inclination that contrasts with the constant negative inclinations characterizing magnetozone R1 and might represent a basal magnetozone (labeled N1).

The thermal demagnetization of the composite IRM reveals a strikingly consistent behavior for all the units in the LM core. All the studied samples are dominated by the low-coercivity component, which experiences a progressive decay until being completely unblocked just below 600ºC (Fig. 5). The intermediate- and high-coercivity components show a similar decay and are also completely unblocked below 600ºC. These data indicate that magnetite is the main magnetic carrier in the studied sediments. Below 600ºC, no clear decay is observed, but a slight change in the slope of the IRM occurs at approximately 275-330ºC, possibly indicating the presence of some magnetic iron sulphides in some samples. The homogeneity in the magnetic behavior regardless of lithology is further evidenced by the strikingly constant values of the different magnetic parameters considered proxies for the type (S ratio), concentration (ARM) and grain size (χARM/χ) of magnetic minerals (Fig. 7b, c, d). S ratios of approximately 0.92 are observed throughout the core and confirm that magnetite is the main magnetic carrier. The magnetite exhibits a low and constant concentration, as indicated by the ARM values of <1 x 10-5 Am^2/kg throughout the core. χARM/χ ratios of approximately 2 are observed throughout the core, which excludes the possibility that magnetite magnetofossils (e.g., Suk, 2016), reported in other marine sediments from the lower Guadalquivir Basin (Larrasoaña et al., 2014), are the main magnetic minerals in the LM core sediments. Overall, the rock magnetic data indicate that magnetite of detrital origin is the carrier of the ChRM in the LM core sediments.

Planktonic Foraminiferal Biostratigraphy

The relative abundances of planktonic foraminiferal species with biostratigraphic value are plotted in Fig. 2. The most representative among these are depicted in Fig. 3. Planktonic foraminifera were found in the interval from 275 m to 33 m. In the lower part of the study interval (275-210 m), *Globorotalia puncticulata*, *Globorotalia crassaformis crassaformis*, and *Sphaeroidinellopsis seminulina* show relatively high abundances (Figs. 2 and 3). In this interval, *Globorotalia margaritae* is abundant at approximately 270 m and 220 m, with a minimum between these 2 maxima. *Globorotalia menardii*, *Sphaeroidinellopsis subdehiscens*, and *Dentoglobigerina altispira* show only one peak at 258, 240, and 225 m, respectively. The middle part of the study interval (210-135 m) is characterized by the dominance of *G. margaritae* (Figs. 2 and 3). In this interval, *G. puncticulata*, *G. crassaformis crassaformis*, and *S. seminulina* are also present. In the upper part of the study interval (135-60 m), *G. crassaformis crassaformis* and *Globorotalia crassaformis viola* dominate. *Globorotalia plesiotumida*, *D. altispira*, and *G. puncticulata* also appear. *S. seminulina* and *S. subdehiscens* have a peak at 90 m and 84 m, respectively. Finally, the uppermost interval (60-33 m) is dominated by *G. crassaformis crassaformis*, *Globorotalia crassaformis hessi*, and *G. crassaformis viola*. Additionally, *G. puncticulata*, *D. altispira* and *Globorotalia aemiliana* are also present in this interval (Figs. 2 and 3).

Four planktonic foraminiferal (PF) events are used to develop the age model of the analyzed core (Figs. 2 and 8). The first PF event is the presence of *G. puncticulata* in the lowermost sample of the core (275 m). The second PF event is the last common occurrence (LcO = highest common occurrence) of *G. margaritae* at 126 m. The third PF event is the last occurrence (LO = highest occurrence) of *G. margaritae* at 108 m. The presence of *G. puncticulata* in the uppermost core interval (54-33 m) marks the fourth PF event.

Age Model

Correlation of the polarity sequence established for the LM core with the astronomically tuned Neogene polarity timescale (GTS2012, Hilgen et al., 2012) is straightforward for the interval from N1 to R2 according to the PF events described above (Fig. 8). The first of the PF events is associated with the presence of *G. puncticulata* that first appeared at 4.52 Ma near the top of C3n.2n (Lourens et al., 2004). According to Sierro et al., (2009), *G. puncticulata* appeared synchronously at 4.52 Ma in the mid-latitudes of the North Atlantic and the Mediterranean. In the LM core, *G. puncticulata* is observed in the lowermost sample, which therefore provides a maximum age of 4.52 Ma for the base of the record (Figs. 2 and 8). The second of the PF events is marked by the last common occurrence (highest common occurrence) of *G. margaritae*, which is found at 126 m in the LM core (R2) and represents an age of 3.98 Ma, corresponding to chron C2Ar (Lourens et al., 2004; Sierro et al., 2000). The age of the last common occurrence of *G. margaritae* is derived from astrochronological calibration of mid-latitude Atlantic boreholes (Gulf of Cádiz) correlated to Mediterranean sapropels (Sierro et al., 2000). The third bioevent is the last occurrence (highest occurrence) of *G. margaritae*, which has been dated to 3.81 Ma during chron C2Ar (Lourens et al., 2004) and is located in the LM core at 108 m in R2. The age of this event is based on the isochronous disappearance of *G. margaritae* both in the mid latitude North Atlantic and Mediterranean (Sierro et al., 2000, 2009). With these constraints, the only plausible correlation between the LM record and the geomagnetic polarity time scale (GPTS) implies that R1, N2, and R2 correlate with chrons C3n.1r, C3n.1n and C2Ar, respectively (Fig. 8). Taking into account the presence of *G. puncticulata* in the lowermost sample, which provides a maximum age of 4.52 Ma for the base of the record, we conclude that the interval N1 represents a genuine magnetozone that can be straightforwardly correlated to chron C3n.2n (Figs. 2 and 8).

For the uppermost part of the studied section, the interval of uncertain polarity makes direct correlation with the GPTS unfeasible. Regardless of the poor quality of the paleomagnetic data, the age of the uppermost part of the studied record is constrained by biostratigraphic data. The very low abundances of *G. puncticulata* in the uppermost part of the LM core (in R2, Figs. 2 and 8) suggest an age that is close to, but precedes (i.e., within the upper part of chron C2Ar), the temporary disappearance of this species in the Atlantic Ocean at 3.56 Ma (Sierro et al., 2009). This inference is further supported by the presence of different subspecies from the *G. crassaformis* plexus throughout the studied core (Bylinskaya, 2004; Wade et al., 2011). However, the last common occurrence and last occurrence of *G. margaritae* in LM are reported in the interval where the sand fraction increases significantly. This indicates a shallow marine environment unfavorable for deeper-water dwelling planktonic foraminiferal species (MacLaughlin et al., 1991; Schiebel and Hemleben, 2005; Kucera, 2007), and suggests that these PF events in LM represent artificial last occurrences biased by environmental factors. Therefore, the age of the top of the studied succession can only be estimated extrapolating the sedimentation rate from the middle part of the core (N2), which yields an age of 3.95 Ma for the end of the marine sedimentation (Fig. 8). This age of 3.95 Ma is further supported by the absence of *Globorotalia miocenica* in the LM core, which appears at 3.77 Ma in the Atlantic Ocean (Wade et al., 2011). This allows to tentatively infer the onset of the continental sedimentation at an age not older than 3.95 Ma in the LM core (Fig. 8). The lower Pliocene age (i.e., within chron C2Ar) of the interval of uncertain polarity in the upper part of the core confirms the poor recording fidelity of the sandy sediments from this interval. We attribute this poor fidelity to postdepositional realignment of detrital magnetite during later polarity periods (Fig. 8).

We propose an age model for the LM core with three tie points provided by the tops of chrons C3n.2n, C3n.1r, and C3n.1n. Using these tie points, sedimentation rates can be calculated for the different intervals (Table 2). R1 is calculated to have a sedimentation rate of 42.9 cm/kyr. Extrapolating this sedimentation rate downward to the bottom of N1, we infer an age of 4.5 Ma for the bottom of the LM record (Fig. 8). The sedimentation rate for N2 is 49.9 cm/kyr. For the upper part of the record (R2 and uncertain polarity interval), we extrapolate the sedimentation rates of N2 (49.9 cm/kyr) (Fig. 8). In summary, this chronology demonstrates that the LM core represents a continuous 550-kyr early Pliocene marine record (4.5-3.95 Ma) (Fig. 8), which may contain information on important paleoclimatic, paleoceanographic and paleoenvironmental changes following the opening of the Strait of Gibraltar and the re-establishment of the MOW.

LITHOSTRATIGRAPHY, MAGNETOSTRATIGRAPHY AND BIOSTRATIGRAPHY OF THE ADDITIONAL STUDIED CORES AND BOREHOLES

The Montemayor-1 core

The continuous Montemayor-1 (MT) core is 260 m long and includes marine sediments ranging from the late Tortonian to the early Pliocene (Zanclean) (Figs. 1A, 1B; Table 1; Larrasoaña et al., 2008; Pérez-Asensio et al., 2012; van den Berg et al., 2015). This core was recovered in the northwestern margin of the lower Guadalquivir Basin. At the beginning of the core, 1.5 m of reddish clays from the Paleozoic-Mesozoic substrate are found. This basement is unconformably overlaid by 0.5 of sandy calcarenites belonging to the Niebla Formation (late Tortonian). A 198-m-thick unit of silts and clays from the Gibraleón Formation (latest Tortonian-Messinian) with a glauconitic layer at the base overlays the Niebla Formation. Over the Gibraleón Formation clays, 42 m of early Pliocene sands and silts from the Huelva Formation are found. This formation presents a glauconitic level at its base. The core ends with 14.5 m of sands from the Bonares Formation (early Pliocene), which unconformably overlays the Huelva Formation, and a 3.5-m-thick recent soil.

The main PF events identified in the MT core are the PF events 2, 3, 4, and 6 of Sierro et al. (1993) (Larrasoaña et al., 2008, 2014). The PF event 2 is the appearance of abundant *Globorotalia menardii* (dextral coiling); the PF event 3 is the regular appearance of *Globorotalia miotumida* (marking the Tortonian-Messinian boundary); the PF event 4 is the first abundant occurrence of dextral *Neoglobloquadrina acostaensis*; and the PF event 6 is the first abundant occurrence of *Globorotalia margaritae*. Finally, the appearance of *Globorotalia puncticulata* is also used as a biostratigraphic datum.

The paleomagnetic data of the MT core show a magnetozone pattern with 11 magnetozones (6 reversed, R1-R6; and 5 normal, N1-N5) (Larrasoaña et al., 2014). Using the PF events, the magnetozones can be correlated to the astronomically tuned geomagnetic polarity timescale (ATNTS2004) of Lourens et al., (2004). Thus, the MT core ranges from chron C3Br.2r (latest Tortonian, 7.4 Ma) to the C3r/C2Ar boundary (early Pliocene, ca. 4.3-4.2 Ma) (Larrasoaña et al., 2014). Cyclostratigraphic analyses have validated the magnetobiostratigraphic age model for the

Messinian interval of the MT core with a precision of one precession cycle (van den Berg et al., 2015).

The Lebrija core

The continuous Lebrija (LE) core is 336 m long and was drilled close the southwestern active margin of the lower Guadalquivir Basin (Figs. 1A, 1B; Table 1; Salvany et al., 2011). The core has marine and continental sediments from the early Pliocene to the Holocene. Sands and silts from the early Pliocene marine Huelva Formation, 115 m in thickness, occur at the base of the core. Over them, 144 m of sands, gravels and clays from the continental Lebrija Formation (late Pliocene-late Pleistocene) are found. At the top of the core, there are clays and sands, 77-m-thick, from the Holocene continental-estuarine Marismas Formation.

The chronology of the LE core is based on magnetostratigraphic data and radiocarbon (^{14}C) dating (Salvany et al., 2011). A total of 10 magnetozones were identified (5 reversed, R1-R5; and 5 normal, N1-N5). The two available radiocarbon ages in the Marismas Formation (9600 ± 50 years BP and >49000 years BP) and the early Pliocene age of the Huelva Formation allow to correlate the magnetozones to the astronomically tuned geomagnetic polarity timescale (ATNTS2004) of Lourens et al., (2004). Consequently, the LE core spans from chron C2Ar (early Pliocene, ca. 4.2 Ma) to chron C1n (Holocene, 0 Ma) (Salvany et al., 2011).

The Villamanrique-1 borehole

The 1341-m-thick Villamanrique-1 (VM) borehole shows marine and continental sediments that encompass the latest Tortonian, Messinian, early Pliocene and Quaternary (Figs. 1A, 1B; Table 1; Ledesma, 2000). This borehole is located in the central part of the lower Guadalquivir

Basin at the basin's axis. The base of the borehole consists of 24 m of Paleozoic schists from the basement followed by 121 m of calcareous clays with sands from the Triassic. Grey marls with abundant foraminifera, 121-m-thick, from the marine Unit B (latest Tortonian-Messinian) are found overlaying the Triassic sediments. Over the Unit B, 375 m of Messinian clays with intercalated sandy levels from the marine Unit C are found. The early Pliocene marine Unit D consists of 660 m of clays and sands. The borehole finishes with 40 m of continental sands from the Quaternary.

Concerning the biostratigraphy of the VM borehole, the six PF events of Sierro et al., 1993 were identified (Ledesma, 2000): PF event 1 (disappearance *Globorotalia menardii* (sinistral coiling), PF event 2 (appearance of abundant *Globorotalia menardii* (dextral coiling)); PF event 3 was defined by the replacement of the group of *G. menardii* by the group of *Globorotalia miotumida* (marking the Tortonian-Messinian boundary (Sierro, 1985)); PF event 4 is the first abundant occurrence of dextral *Neoglobloquadrina acostaensis*; PF event 5 (disappearance of *Globorotalia miotumida*) and PF event 6 is the first abundant occurrence of *Globorotalia margaritae*. Finally, the presence of abundant *Sphaeroidinellopsis* might indicate the acme of this taxon around the Miocene-Pliocene boundary.

The chronology of the VM borehole is based on the aforementioned biostratigraphic events and a cyclostratigraphic analysis using resistivity logging (Ledesma, 2000). This astrobiochronologic dating allows to date the lower and middle parts of the VM borehole (1341 to 700 m core depth) ranging from the chron C3Br.3r (latest Tortonian, ca. 7.5 Ma) to the Miocene-Pliocene boundary. In the upper part of the core, Unit D (700 to 40 m core depth) is mostly early Pliocene due to the presence of *G. margaritae* and the absence of *Globorotalia puncticulata*, which appears at 4.52 Ma. The uppermost 40 m of the borehole correspond to Quaternary continental sandy sediments.

SUBSIDENCE ANALYSIS OF THE VILLAMANRIQUE BOREHOLE

To estimate the vertical tectonic motions around the M/P boundary at the Villamanrique-1 (VM) borehole, we apply a simplified backstripping analysis in an interval from 6.50 to 5.20 Ma (920 to 650 m core depth) encompassing the Miocene-Pliocene boundary (5.33 Ma, 700 m core depth). The analysis accounts for sediment compaction, isostatic subsidence due to the sediment weight and water column, observed paleodepth change and eustatic sea-level fluctuations (Watts, 1988; Allen and Allen, 1990). For the sediment compaction correction, we used surface porosity of 0.63, porosity-depth coefficient c of 0.51 km^{-1} and sediment grain density of 2720 kg/m^3 for clays; and surface porosity of 0.56, porosity-depth coefficient c of 0.39 km^{-1} and sediment grain density of 2680 kg/m^3 for clayey sandstone. Paleodepth estimations were calculated with a transfer function based on benthic foraminiferal depth ranges and their relative abundances (Hohenegger, 2005, 2008; Báldi and Hohenegger, 2008; Pérez-Asensio et al. 2012, 2017). These paleobathymetric calculations provide a 95% confidence interval as a measure of the accuracy of the estimated paleodepths. A total of 10 samples along the studied interval (920 to 650 m core depth) were analyzed for benthic foraminiferal content and paleodephts were calculated. The eustatic corrections were done using the global sea-level record of Miller et al., (2011).

In the studied interval, the backstripping results show an average tectonic subsidence rate of 166 m/Ma (95% confidence range: 78-252 m/Ma) for the entire 6.50-5.20 Ma interval. Therefore, even considering the uncertainties in the paleodepth calculations (ca. 200 m) and the global sea-level record we can ensure this subsidence values within a 95% confidence interval. Such continuous tectonic subsidence would be consistent with a slab pull across the Miocene-Pliocene boundary as proposed below.

DISCUSSION

Stratigraphic Implications

The magnetobiostratigraphic data and age model presented in this study can be used to redefine some age constraints on the Neogene lithostratigraphic units previously described in field-based studies in the northern part of the Guadalquivir Basin. First, the youngest sediments in the Gibraleón Formation are thought to be uppermost Messinian in Huelva and neighboring areas (Flores, 1987; Sierro et al., 1993; van den Berg et al., 2015, 2017). Nevertheless, the magnetobiostratigraphic data from the LM core presented herein clearly indicate a lower Pliocene (Zanclean) age for the upper part of this formation (Figs. 8, 9). The discrepancy in the age of the top of this formation between field-based studies and borehole-based studies can be explained on the basis of the sedimentary context of the basin: the Gibraleón Formation sediments in the central areas of the basin are interpreted as deeper-water, fine-grained facies that are laterally equivalent to the sandy and silty facies of the Huelva Formation. The vertical transition from blue marls to silty-sandy deposits observed in LM at around 4 Ma occurred in the northern margin of the basin (Montemayor-1 (MT) core) more than 1 Ma earlier, probably because of the southward and westward progradation of shallow-water sandy deposits over the blue marls (Sierro et al., 1996). Upper Miocene coarse sediments (sands) from the active (i.e., southern) margin of the basin also grade into finer sediments (marls) basinward (Aguirre et al., 2015), revealing that the basin was an embayment that opened to the west (Martín et al., 2009, 2014). We propose that the lithologic transitions from fine-grained to coarse-grained sediments should be used with caution when pinpointing the Miocene-Pliocene boundary in boreholes or seismic profiles without performing a detailed dating analysis of the sediments across this boundary.

Another stratigraphic implication concerns the Transitional Unit. According to Mayoral and González (1986-87), this unit encompasses the Miocene-Pliocene transition and features a lower unit composed of Miocene-Pliocene blue marls and an upper unit composed of lower Pliocene silts and yellow sands. The lower unit of the transitional facies described by Mayoral and González (1986-87) can be assigned to the clays of the Gibraleón Formation in the LM core on the basis of the lithologic similarities and have an early Pliocene age in the central areas of the basin as reported in this study. The upper unit of the transitional facies described by Mayoral and González (1986-87) is comparable in age (early Pliocene) and lithology to the transitional facies found in the LM core (Fig. 9). The Transitional Unit therefore represents a diachronous (Miocene to Pliocene) intermediate period between shallower, coarser deposits along the basin margins (i.e., the Huelva Formation and equivalents) and deeper, finer deposits in the basin center (the Gibraleón Formation).

Based on the time-transgressive nature of the Huelva Formation (Civis et al., 1987), deposition of this unit appears to have begun at the Miocene-Pliocene boundary in the marginal positions of the basin (van den Berg et al., 2015, 2017) but during the early Pliocene (4.2 Ma) in more basinal locations (this study).

Implications for the Early Pliocene Tectonic Evolution of the Gibraltar Strait Area

Unraveling the sedimentary evolution of the Guadalquivir Basin during the early Pliocene is important because it can provide novel insights into the tectonic evolution of the Gibraltar Strait area, with implications for Atlantic-Mediterranean water-mass exchange. It is therefore important to integrate the information on sedimentation rates, lithologies, and formation boundaries obtained from the LM core with that from other boreholes along a NW-SE transect from the passive margin towards the active margin of the basin (Fig. 9). In the northwestern part of the basin, the continuous MT core includes the uppermost 198 m of the Gibraleón Formation, a 45-m-thick sequence of the Huelva Formation (including its basal 3-m-thick glauconite layer), and a 14.5-m-thick sequence of the Bonares Formation. Based on an astronomically-tuned age model for the Gibraleón Formation, high sedimentation rates of ~50 cm/kyr initiated at 5.55 Ma and decreased abruptly to nearly 0 cm/kyr within the glauconite level at the base of the overlying Huelva Formation, which is dated to the Miocene-Pliocene boundary (5.33 Ma, van den Berg et al., 2015). These results are similar to those derived from the astronomically tuned age model of the neighboring Huelva (HU) core, which reveals similarly high sedimentation rates starting at 5.55 Ma that decrease abruptly in the glauconite level at the Miocene-Pliocene boundary (van den Berg et al., 2017). Magnetobiostratigraphic data from the MT core indicate that the lower half of the Huelva Formation, including the glauconite layer, is within a normal polarity zone (N5) that is located just below the first occurrence (lowest occurrence) of *G. puncticulata* at 4.52 Ma (Larrasoaña et al., 2014). This situation implies that N5 can be correlated with chron C3n.2n (i.e., 4.493-4.631 Ma, see Larrasoaña et al., 2014) and that a hiatus of approximately 0.7 Ma occurred somewhere between the glauconite layer and the middle part of N5. An alternative to such a hiatus, for which no clear expression is found in the sedimentary record, would be a period of sedimentary bypass and starvation that is more consistent with the origin of the glauconite layer (van den Berg et al., 2017). A similar period of extremely low sedimentation rates can be deduced from outcrops in the Moguer section (Huelva, SW Spain) near the MT core location, where *G. puncticulata* occurs within the Huelva Formation above the glauconite layer (Sierro, 1984). Backstripping analyses performed by Pérez-Asensio et al. (2013) in MT core indicate, when placed in the astronomically-tuned age model of van den Berg et al. (2015), that this period of sedimentary bypass occurred in a context of tectonic uplift.

Magnetobiostratigraphic data in the MT core are not available for the upper part of the Huelva Formation and the Bonares Formation due to the coarse grain size of the sediments, which suggest progressively shallower marine conditions up section (Larrasoaña et al., 2014). Regardless, the linear extrapolation of the low sedimentation rates from the lower part of the Huelva Formation (i.e., <10 cm/kyr) upwards from the first occurrence (lowest occurrence) *G. puncticulata* yields an age for the end of marine sedimentation (i.e., the top of Bonares Formation) within the middle-lower part of chron C2Ar (Fig. 9).

In the center of the basin, the Villamanrique-1 (VM) borehole includes uppermost Tortonian to lower Pliocene marine sediments (Figs. 1A, 1B; Table 1; Ledesma, 2000) and a continental sequence that corresponds to the upper Pliocene to Quaternary Almonte-Lebrija formations of Salvany et al. (2011). Astronomically tuned age models for the marine sequence of the VM and other neighboring boreholes indicate a significant change in the sedimentation setting across the Miocene-Pliocene boundary (Ledesma, 2000), where a marked increase in sedimentation rates from approximately 25 to 80 cm/kyr is observed within a sequence of fine-grained sediments (clays and silty clays, Unit C) similar to those of the Gibraleón Formation (Fig. 9). Higher up in the sequence, the absence of *G. puncticulata* in the uppermost 300 m of the lower Pliocene marine sequence may suggest an age older than 4.52 Ma (Ledesma, 2000). Nevertheless, this age constraint has to be treated with caution due to the small number of studied samples and their coarse-grained nature (Unit D, equivalent to the Huelva Formation). These sediments are indicative of progressively shallowing marine conditions, which led to very low percentages of planktonic foraminifera (Ledesma, 2000). The onset of the continental sedimentation in the VM core should be cautiously placed between 4.52 and 3.7 Ma (age of continentalization in Lebrija core) (Fig. 9).

Farther to the SE in the active margin of the basin, the Lebrija core includes the magnetostratigraphically dated uppermost sediments of the lower Pliocene Huelva Formation and the upper Pliocene to Quaternary continental sediments of the Lebrija Formation (Salvany et al., 2011). Although the age model does not allow calculation of sedimentation rates for the Huelva Formation sediments, it enables dating of the continentalization of the basin to 3.7 Ma (Salvany et al., 2011).

Overall, these data have important implications for assessing the sedimentary evolution of the Guadalquivir basin during the latest Messinian and early Pliocene. The latest Messinian was characterized by sedimentation in relatively deep marine conditions throughout the basin (Ledesma, 2000; Pérez-Asensio et al., 2012). The sedimentation rates were broadly homogeneous at 10 to 25 cm/kyr but rapidly increased to peaks of 90 cm/kyr only in along the passive margin (MT and HU cores, van den Berg et al., 2015, 2017). Right at the Miocene-Pliocene boundary, the sedimentation rates along the passive margin decreased abruptly (MT and HU cores, van den Berg et al., 2015, 2017) in association with a drop in the water depth conditions (MT core, Pérez-Asensio et al., 2012). Continued deep marine sedimentation with high sedimentation rates occurred simultaneously across the Miocene-Pliocene boundary in the basin center (VM borehole, Ledesma, 2000). During the earliest Pliocene, a 0.7 Ma period of sedimentary bypass occurred in the uplifting passive margin of the basin under shallow marine conditions (MT core, Pérez-Asensio et al., 2012, 2013, this study). Enhanced uplift towards more distal positions in the passive margin of the basin is indicated by the erosional hiatus reported in the present-day off-shore extension of the basin in the Gulf of Cádiz associated with the Miocene-Pliocene boundary (Hernández-Molina et al., 2013, 2014, 2016). The fact that no major sea-level change is associated with the Miocene-Pliocene boundary (Miller et al., 2005; 2011) clearly points to tectonic uplift as the underlying cause of this erosional hiatus at the most distal fringe of the basin and of coeval sedimentary bypass at the location of core MT. Simultaneously, deep marine sedimentation with high sedimentation rates was recorded in the central part of the basin (VM borehole, Ledesma, 2000). A backstripping analysis of the vertical tectonic motions at the center of the basin (VM borehole) indicate continuous tectonic subsidence across the Miocene-Pliocene boundary consistent with the increase in sedimentation rates during the early Pliocene (Fig. 9). The presence of relatively deep marine sedimentary facies with high sedimentation rates (i.e., the Gibraleón Formation) in the LM core (this study) suggests that sedimentary bypass and uplift were restricted to the passive margin of the basin, with the rest of the basin undergoing continued tectonic subsidence (Figs. 9 and 10). In the later early Pliocene, low (<10 cm/kyr) sedimentation rates on the passive margin (MT core, Pérez-Asensio et al., 2012; Larrasoaña et al., 2014) contrast with much higher sedimentation rates in the central part of the basin (>40 and up to 80 cm/kyr in LM and VM, respectively) (Ledesma, 2000 and this study) (Fig. 9). These high sedimentation rates in the central part of the basin are linked to the west-southwestward progradation of large depositional systems along the axis of the basin (Sierro et al. 1996), a process that accelerated after the Miocene-Pliocene boundary and that also involved the southward progradation of clinoforms towards the basin axis (Ledesma, 2000). Continuous tectonic subsidence across the Miocene-Pliocene boundary surely contributed to create accommodation space for sediment bodies progradation towards the south.

There are 3 putative mechanisms that might have led to the early Pliocene tectonic subsidence in the basin center and simultaneous forebulge uplift on the basin passive margin: 1) compression in the external western Betics leading to thrusting and loading (Ruiz-Constán et al., 2009; González-Castillo et al., 2015); 2) post-tectonic viscous relaxation of the lithosphere below the basin (Garcia-Castellanos et al., 2002); and 3) slab pull from a subducted or delaminated lithosphere below the Gibraltar Arc region (Garcia-Castellanos and Villaseñor, 2011). Modern compression in the external western Betics is related to the favorable orientation of the mountain front with respect to ongoing Eurasia-Africa plate convergence and appears to be accommodated by a blind thrust system at depths of 8-12 km that forms a tectonic wedge with top-to-the NW tectonic transport (Ruiz-Constán et al., 2009; González-Castillo et al., 2015). We acknowledge that ongoing active compression along this tectonic wedge during the Miocene-Pliocene boundary might explain forebulge uplift and basin subsidence, as has been recently proposed by van den Berg et al. (2017). However, this mechanism also implies tectonic uplift of the wedge top that is consistent with neither the southward-prograding clinoforms reported in the area starting at the Miocene-Pliocene boundary nor the dominantly northern provenance of the sediments (Ledesma, 2000; Sierro et al., 2008). Post-tectonic viscous relaxation of the lithosphere can also be excluded because it would imply a protracted period of forebulge uplift (see Garcia-Castellanos et al., 2002) that does not fit the seemingly short-lived (~0.7 Ma) period of margin uplift reported in this study and in Hernández-Molina et al. (2013, 2014, 2016) immediately after the Miocene-Pliocene boundary. In contrast to these mechanisms, the pull of a sinking slab below the Gibraltar Arc region (Garcia-Castellanos and Villaseñor, 2011) can simultaneously explain the temporary forebulge uplift, the continuous basin subsidence, and the southward migration of clinoforms during the earliest Pliocene (Fig. 10). Based on seismic stratigraphy, wells, and hydrodynamic numerical modeling in the Alborán Basin and the Gulf of Cadiz, it has been proposed that the pull of a sinking slab led to the opening of the Strait of Gibraltar and the Zanclean Deluge at the

Miocene-Pliocene boundary (Garcia-Castellanos et al., 2009; Garcia-Castellanos and Villaseñor, 2011). This hypothesis has been supported by seismic tomography data and by the uplift of the internal basins within the Betic Cordillera (Garcia-Castellanos and Villaseñor, 2011). The sedimentary evolution of the Guadalquivir Basin during the early Pliocene provides the first sedimentary evidence for the subsidence associated to the westward shift in sub-lithospheric loading and provides independent support for the hypothesized lithospheric tearing beneath the Betics.

The different ages of the beginning of continental sedimentation, with gradually younger ages towards the southeastern margin of the basin (Fig. 9), point to a NW to SE diachronous continentalization as consequence of the progradation of deltaic systems (Bonares Formation). During the late Pliocene to early Pleistocene, continental progradation of alluvial systems of the Almonte and Lebrija formations also occurred from NW to SE (Salvany et al., 2011).

CONCLUSIONS

Magnetobiostratigraphic analysis of the La Matilla (LM) core, located in the lower Guadalquivir Basin (SW Spain), provides an age model that yields lower Pliocene ages (from 4.5 to 3.95 Ma) for the studied interval. This chronological framework provides age constraints for a continuous 550-kyr lower Pliocene marine sedimentary sequence that might have recorded significant paleoclimatic, paleoceanographic and paleoenvironmental changes when the MOW was reactivated due to the opening of the Strait of Gibraltar.

Based on the updated chronology of the LM core, several stratigraphic implications for the Neogene lithostratigraphic units from the lower Guadalquivir Basin arise. Firstly, the youngest sediments from the Gibraleón Formation, assigned to the Messinian on the basin margin, were deposited during the early Pliocene (Zanclean) in the central areas of the basin. Therefore, the Gibraleón-like facies in the LM core might represent the lateral facies change of the sandy and silty facies of the Huelva Formation on the basin margin. This implies that locating the Miocene-Pliocene boundary in the Guadalquivir basin based exclusively on lithological changes without performing detailed dating is not valid. Secondly, the transitional facies is a diachronous (Miocene to Pliocene) intermediate unit between marginal coarser sediments and basinal finer sediments. As expected, gradually finer facies basinward is the result of migrating depositional facies following the Walther's Law of Facies (Walther, 1894).

The sedimentary evolution of the Guadalquivir Basin during the late Miocene-early Pliocene provides new insights into the tectonic activity in the Gibraltar Arc area. In the northwestern passive margin, tectonic forebulge uplift produced a sharp relative sea-level fall and an abrupt decrease in sedimentation rates, causing a 0.7-Ma period of sedimentary bypass. In contrast, the central areas of the basin experienced continuous deep marine sedimentation with high sedimentation rates, suggesting continuous tectonic subsidence and the WSW-directed progradation of depositional systems. Among the three possible mechanisms that could explain the contemporaneous forebulge uplift in the basin passive margin and the continuous tectonic subsidence and southward-prograding clinoforms in the basin center, the most plausible is the westward migration of the pull of a subducted lithospheric slab beneath the Gibraltar Arc area. The opening of the Strait of Gibraltar, causing the re-establishment of the MOW and the Zanclean flood at the Miocene-Pliocene boundary, has been suggested to have been a consequence of this geodynamic process and the resulting vertical motions. The sediments of the lower Guadalquivir Basin are the first geological evidence of these significant tectonic subsidence and paleoceanographic events.

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Figure Captions

Figure 1. (a) Simplified geological map of the study area and locations of the studied La Matilla, Montemayor-1, Villamanrique-1, and Lebrija cores. The locations of other boreholes (Casanieves-1, Isla Mayor-1, and B14-1, see Fernàndez et al., 1998) has been also indicated. The regional cross section in Figure 1B is indicated. (b) A NW-SE cross section of the lower Guadalquivir Basin showing the main lithostratigraphic units and boreholes locations. The section has been constructed using information from boreholes shown in Figure 1A and the seismic profile shown in Figure 1C (see also Ledesma, 2000 and Fernàndez et al., 1998). (c) NW-SE seismic section showing the lithostratigraphy of the lower Gualdalquivir Basin and Late Tortonian emplacement of the Betic Front (accretionary wedge). The location of Villamanrique-1, Casanieves-1 and Isla Mayor-1 boreholes is also shown.

Figure 2. Lithology, stratigraphy and biostratigraphy of the studied interval in the La Matilla core. The position of the analyzed biostratigraphic samples is indicated. Relative abundance (in percentage) of planktonic foraminiferal species used as biostratigraphic markers. Four planktonic foraminiferal (PF) events are indicated by the horizontal lines.

Figure 3. Representative planktonic foraminiferal species with biostratigraphic value from the La Matilla core. (a) *Globorotalia puncticulata*; spiral view. (b) *Globorotalia puncticulata*; peripheral

view. (c) *Globorotalia puncticulata*; umbilical view. (d) *Globorotalia crassaformis crassaformis*; spiral view. (e) *Globorotalia crassaformis crassaformis*; peripheral view. (f) *Globorotalia crassaformis crassaformis*; umbilical view. (g) *Sphaeroidinellopsis subdehiscens*; umbilical view. (h) *Sphaeroidinellopsis seminulina*; umbilical view. (i) *Globorotalia margaritae*; spiral view. (j) *Globorotalia margaritae*; peripheral view. (k) *Globorotalia margaritae*; umbilical view. (l) *Globorotalia plesiotumida*; spiral view. (m) *Globorotalia plesiotumida*; umbilical view. (n) *Dentoglobigerina altispira*; umbilical view. Scale bars = 100 μm.

Figure 4. Representative facies of the La Matilla core: (a) Yellow sand of the Abalario Formation. Interval between 3 and 4.2 m core depth. (b) Coarse and loose sand of siliceous composition in the Almonte Formation. Interval between 27.9 and 28.1 m core depth. (c) Well-rounded and well-sorted gravel clasts of siliceous composition in the Huelva Formation. Interval between 46.5 and 46.7 m core depth. (d) Mixture of shell fragments, well-rounded gravel clasts and coarse sand of the Huelva Formation. Interval between 45.5 and 45.7 m core depth. (e, f) Coquina levels (bivalve fragments and brachiopods (arrow)) found at 51.8 m and 66 m core depth in the Huelva Formation. (g) Silty sand with scattered scaphopod fragments (arrows) in the Huelva Formation at 54.7 m core depth. (h) Uniform grey clays of the Gibraleón Formation. Interval between 243.4 and 244.4 m core depth.

Figure 5. Examples of representative thermal demagnetization diagrams for the different lithologies in the LM core. Black and white dots indicate projections onto the horizontal and vertical planes, respectively.

Figure 6. Thermal demagnetization diagrams of a composite IRM for representative samples of the different stratigraphic units in the LM core.

Figure 7. Downcore variations in the inclination of the ChRM and the associated pattern of polarity intervals identified in the LM core (a), which are shown along with variations in the magnetic parameters indicative of the type (S ratio (b)), concentration (ARM (c)), and grain size (χ_{ARM}/χ (d)) of magnetic minerals. The sand content (>63 μm) is also shown.

Figure 8. Age model and sedimentation rates (cm/kyr) for the LM core based on the magnetobiostratigraphic results. Planktonic foraminiferal events are indicated. FO = First occurrence (lowest occurrence); LO = Last occurrence (highest occurrence); LcO = Last common occurrence (highest common occurrence).

Figure 9. Chronostratigraphic framework of the Montemayor-1 (passive margin), La Matilla, Villamanrique-1 (basin center) and Lebrija (active margin) cores. Main tectonic movements, glauconite layer, sediment bypass interval and sedimentation rates in cm/kyr are shown.

Figure 10. Cartoon showing the geodynamic processes related to the Miocene-Pliocene sedimentary bypass in the marginal areas and tectonic subsidence in the basinal areas. The basin experienced vertical tectonic movement (indicated by solid arrows) related to 1) lithospheric bending in response to tectonic thrusting, 2) horizontal stresses associated with Africa-Iberia convergence, and 3) gravitational pull from a lithospheric slab beneath the Western Betics. The ellipse indicates the location of the NW-SE cross section in Figure 1B.

Table Captions

TABLE 1. STUDIED CORES IN THE LOWER GUADALQUIVIR BASIN INDICATING LENGTH, UTM COORDINATES (X, Y), HEIGHT (Z), INTERVALS OF CONTINUOUS CORE SAMPLING (CCS) AND AVERAGE OF CSS (%).

TABLE 2. AGE DATUMS USED FOR THE AGE MODEL AND SEDIMENTATION RATES (cm/kyr) OF THE LA MATILLA CORE (SW SPAIN).