INTRODUCTION

The Messinian Erosional Surface (MES, Cita and Ryan, 1978) is a subaerial erosional surface that deeply cuts the continental margins and developed during the acme of the Messinian Salinity Crisis (MSC, Ryan et al., 1973), when the Mediterranean sea-level dropped more than 1000 meters and shallow-water evaporite precipitation was shifted to the deepest parts (Roveri et al., 2006). The rise in global sea level at the beginning of the Early Pliocene (Haq et al., 1987), probably combined with a change in the tectonic activity in the Alboran Sea area, caused the flooding of the desiccated basin (Zanclean transgression). This rise in sea level marked the end of the salinity crisis with the re-establishment of open marine conditions in the Mediterranean. From that time onward, the Early Pliocene was marked by a long period (about 1.7Ma) of high sea level in which the Mediterranean margins were rebuilt by the sedimentary infill of the MES (Clauzon et al., 1987). Many recent works (Rabineau, 2001; Lofi et al., 2003; 2005; 2011; Duvail et al., 2005; Bache, 2008; Bache et al., 2009; 2010; 2012; García et al., 2011; Urgeles et al., 2011; Maillard et al., 2011; Leroux, 2013) have described and quantified the Pliocene–Quaternary Mediterranean sedimentation.
based on a broad database of seismic reflection profiles. The prograding sedimentary prisms migrated rapidly seaward and filled the underlying Messinian topographic lows. This contribution aims to improve our understanding of the Early Pliocene transgression and the re-building of the Mediterranean continental margin after the major Messinian erosion of the Miocene continental shelves by studying the stratigraphy and nannofossil assemblages of onshore well-exposed Lower Pliocene sediments.

**GEOLOGICAL SETTING**

The Betic Cordillera, located in the southeast of the Iberian Peninsula, has an ENE–WSW trend and has classically been divided into two major zones (Fig. 1): the External Zone, adjacent to the Iberian foreland and mainly composed of sedimentary rocks, and the Internal Zone, farther away from the foreland and composed mainly of metamorphic rocks (Fallot, 1948). Before the MSC occurred, two straits, the Northbetic and the Rifian, both representing the foreland of the Betic and Rif cordilleras, formed the Atlantic-Mediterranean passages. The building up of the Betic and Rif cordilleras produced a narrowing and partial closure of the passages (Viseras et al., 2004). The onset of the MSC was a consequence of this seaway restriction, and the opening of the Strait of Gibraltar marked the end of the MSC (Ryan et al., 1973).

Previous works carried out in the Bajo Segura Basin, a Mediterranean marginal Betic basin with evaporite deposits during the MSC (Caracuel et al., 2004; 2011; Soria et al., 2005; 2008a; García-García et al., 2011; Martínez del Olmo, 2011a), have described a major erosional surface separating the Upper Messinian deposits, mostly non-marine, and the overlying Pliocene marine transgressive deposits. The most significant features of this surface, assigned to the MES are broad paleovalleys up to 200m deep (Caracuel et al., 2004; 2011; Soria et al., 2008a; 2008b; García-García et al., 2011; Martínez del Olmo, 2011a).

Recent studies of the MES in seismic profiles (Bache, 2008; Bache et al., 2009; García et al., 2011; Martínez del Olmo 2011b; Bache et al., 2012) have differentiated “rough or badland” morphology and a more basinward “smooth” surface. The latter deepens slightly seaward and extends over 60–70km (Bache et al., 2009). The transition between the two morphologies (rough and smooth) is very clear and lies at a constant two-way-traveltime depth of 1.6 seconds over most of the Provence shelf (Bache, 2008; Bache et al., 2009) and the Valencia trough (Martínez del Olmo, 2011b). Bache et al. (2012) interpreted this surface as the result of a two-step process for the reflooding of the Mediterranean after the MSC, suggesting an initial moderate and relatively slow reflooding accompanied by transgressive ravinement, followed by a second very rapid step that preserved the subaerial MES. The amplitude of these two successive rises in sea level were estimated by Bache et al. (2012) at ≤500m for the first rise and 600–900m for the second rise. According to these authors the start of the second step preceded the Zanclean Global Stratotype Section and Point in Eraclea Minoa (Sicily, Van Couvering et al., 2000; Hilgen et al., 2012), and therefore starts before the end of the Messinian (Bache et al., 2012). The Bajo Segura Basin sections studied here may correspond to the latest Messinian-earliest Pliocene sediments filling the “rough” MES topography.

**THE TRANSGRESSIVE PLIOCENE**

Six stratigraphic sections have been studied (Fig. 2). The Pedrera section (PE), previously studied by Lancis et al. (2004a) and Soria et al. (2008a), is located to the west of the La Pedrera dam. The Pantano de Elche section (PETEL) outcrops in the eastern margin of the Vinalopó river cut (just below the A7-highway bridge to the north of the city of Elche). The Guardamar de Segura section (GUA) is found to the west of the village of the same name; the bottom of the section is at the north of the cemetery and runs following an access road to the water deposit of the village. The San Miguel de Salinas (SM) is the “Canal del Transvase Tajo Segura” section (a water channel) located to the west of the town of San Miguel de Salinas, previously studied by Corbi Sevila (2010). The beds dip 10º to 20º to the east. The Dehesa de Pino Hermoso section (DPH) is located to the southwest of the village of Arneva. As the section bedding tilts to the west it runs across different orange groves in the same direction. The Santa Pola section (SP) corresponds to the northern side of the Santa Pola Sierra, in the Clot de Galvany protected forest area (previously studied by Lancis et al., 2004b). The section runs along the southern side of the “Cabeçó” hill, just west of the La Charca lagoon.

Current knowledge of the Early Pliocene transgression of the Eastern Betic Cordillera basins is based on the work by Montenat (1977), and a later synthesis by Montenat et al. (1990). The uppermost Messinian sedimentation mainly occurred in continental environments (Montenat, 1977; Montenat et al., 1990; Soria et al., 2005; 2008a; 2008b; Caracuel et al., 2011), eroded by incised paleovalleys, filled by high-energy sediments (Caracuel et al., 2004; 2011; García-García et al., 2011), and covered by open marine, white marls (Soria et al., 1996; 2008a; 2008b). These white marls belong to an Early Pliocene major allostratigraphic unit (P unit, according to Caracuel et al., 2006), which forms a shallowing-upward sequence composed of four main terms (Caracuel et al., 2006): i)
The Pliocene infilling of the Messinian Erosional Surface

FIGURE 1. A) Location of the Betic Cordillera in the western Mediterranean. B) Geological map of the eastern end of the Betic Cordillera showing the position of the Bajo Segura Basin. C) Geological map of the Bajo Segura Basin (simplified from Montenat, 1990). The dashed line marks the correlation chart in Figure 2.
the basal high-energy lag, or P0, of the Early Pliocene transgression, and the high-energy paleo-valley fills (Pedrera formation in Soria et al., 2008a); ii) the lower white “Hurchillo marls”, P1 in Montenat et al. (1990); iii) the middle yellow calcareous sandstone, “Rojales sandstone” or P2 (Montenat et al., 1990), interpreted as transitional environments from shoreface to foreshore and ending with backshore and aeolian sand dunes (Soria et al., 2005), and iv) the upper continental “variegated sands and marls”, and the laterally lacustrine equivalent San Pedro
limestone, both included in P3 (Montenat et al., 1990). The paleo-valley fill and the overlying Hurchillo white marls of the Bajo Segura Basin can be the result of the second transgressive step of Bache et al. (2012), and may be interpreted as just the bottom part of the latest Messinian-earliest Pliocene sediments covering the MES.

The classic scheme of the Bajo Segura Basin based on P0, P1, P2 and P3 (see Soria et al., 2005) can be updated. There are white Hurchillo marls above and below the yellow calcareous sandstone in the PE and DPH sections (Fig. 2). Both levels of white marl have the same facies but different nannofossil contents, allowing them to be separated as the lower Hurchillo marls and the upper Hurchillo marls. The latter has an upper yellow calcareous sandstone showing the same transitional facies, ending with aeolian sand dunes, that the lower yellow sandstone. The two yellow calcareous sandstones show several shallowing-upward sequences like the ones described by Soria et al. (2005) for the P2, that is, from bottom to the top: i) sands with bivalves (osteoids and pectinids) and abundant traces of Thalassinoides, in the upper part of which sets of cross stratification caused by the migration of sand waves occur; ii) sands and gravels with bivalves and conglomerates bored by Lithophaga, and iii) well-sorted sands, without fossils, with high-angle cross stratification formed by the migration of aeolian dunes, in which levels of stromatolites and thin channels of gravels are intercalated.

Two shallowing sequences seem to develop in the lower yellow calcareous sandstone (PE, DPH, and SM; Fig. 2). However the more basinward sections (GUA and SP) do not record the lower yellow calcareous sandstone. Instead, the lower and upper Hurchillo marls are separated in the GUA section by a glauconitic level, with abundant bivalve fossils, with high-angle cross stratification formed by the migration of aeolian dunes, in which levels of stromatolites and thin channels of gravels are intercalated.

Hurchillo marls capped by the upper yellow sandstone, and then overlaid by the variegated sands and marls and the San Pedro limestone. The lower sequence in the studied sections has neither variegated sands and marls nor the San Pedro lacustrine limestone.

**METHODS**

All sections were sampled by Lancis (1998) to study their nannofossil assemblages quantitatively, except the SP section, studied qualitatively in Lancis et al. (2004b). Twenty-five samples were collected from the PE section, 10 from the SM section, 26 from the PTEL section, 23 from the GUA section, 33 from DPH (Lancis, 1998), and 8 from the SP section (Lancis et al., 2004b). For each sample collected, four different smear slides were prepared, using a method aimed at increasing the nannofossil-to-silt ratio (Lancis, 1998). For the first smear slide, 0.1g of sediment was suspended in 10ml distilled water (buffered at pH 8) spread on a 300mm² surface in one case (direct suspension, without dilution) and in the other smear slide after a 1/3 dilution of the suspension with distilled water (buffered at pH 8). In order to increase smear slide quality, a second procedure was performed as follows. A 10ml suspension of 0.1g sediment in distilled water (buffered at pH 8) was centrifuged at 1800rpm (450g) for 2 minutes at room temperature. After discarding the supernatant, distilled water was added to the pellet to achieve a 10ml suspension and the new mixture was subjected to sonication for 8 seconds. This centrifugation-sonication procedure was repeated five times. Finally, 0.1ml of the suspension was extended directly, covering a 300mm² over the surface of the slide, without dilution for the third smear slide and after 1/3 dilution with distilled water (buffered at pH 8) for the fourth one. The four prepared smear slides of each sample were analysed quantitatively under a 100x objective, scanning the whole slide to detect rare biostratigraphic markers. The percentage of nannoliths >3µm was determined after counting 500 nannoliths larger than 3µm. For counting “small reticulofenestrids” (<3µm), the mean values were calculated using the percentage of those coccoliths found in 10 visual fields, counting around 3000 nannoliths. Finally, the percentages of the different species of Discoaster spp. were calculated, counting 100 asterololiths.

**BIOSTRATIGRAPHY**

The biozonal schemes of Martini (1971) and Okada and Bukry (1980) were adopted for the Pliocene interval in the sections studied. In addition, other bioevents were used to improve nannofossil biostratigraphy based on Driever (1988), Fornaciari et al. (1990, 1996), Rio et al. (1990),
Bukry (1992), Young et al. (1994), Raffi and Flores (1995), Lancis (1998), Marino and Flores (2002), Lancis and Flores (2006), Lourens et al. (2004), and Raffi et al. (2006). Martini’s (1971) Zones NN14 and NN15 were combined by Rio et al. (1990a) because of the rarity of the Zone NN14/NN15 boundary bioevent, the last occurrence (LO) of *Amurolithus tricorniculatus*, in the Mediterranean region. Figure 3 shows the correlation chart based on the scheme by Lourens et al. (2004) and the chronostratigraphy by Hilgen et al. (2012). Our biostratigraphic analysis yielded a number of calcareous nannofossil events that appeared to be useful to improve the biostratigraphic resolution of the Early Pliocene.

**Calcareous nannofossil assemblages**

The nannofossil assemblage of the lower Hurchillo marls is characterized by an abundance of *Reticulofenestra cisnerosii* (Lancis and Flores, 2006) together with the presence of scarce *Ceratolithus acutus* in the bottom samples (Figs. 4; 5). The most abundant group is the “small reticulofenestrids”, although the species *Coccolithus pelagicus, Reticulofenestra pseudoumbilicus >7µm, Reticulofenestra haqii/minutula, Sphenolithus abies, Sphenolithus neoabies, Scyphosphaera lagena, Helicosphaera carteri, Helicosphaera selli, Calcidiscus macintyreii, Calcidiscus leptopus*, *Umbilicosphaera rotula, U. jafari*, and *Dictyococcites antarcticus* are also abundant. The “small Gephyrocapsa spp.”, and *Pseudoemiliania lacunosa* (Figs. 6; 7; 8; 9; 10; 11) appear in the upper part of the studied interval.

Astero liths are frequent, especially *Discoaster pentaradiatus, D. asymmetricus*, and *D. brouweri*, while *D. surculus, D. quinqueramus*, and *D. berggreni* are absent (Figs. 6; 7; 8; 9; 10; 11).

The top part of the sections shows the first occurrence (FO) of *Discoaster tamalis* and the disappearance of *Reticulofenestra pseudoumbilicus*, and *Sphenolithus* spp.

**Calcareous nannofossil events and biochronology**

In our sections, ten calcareous nannofossil biovents can be recognized from bottom to top (Figs. 3; 4; 5): i) the FO of *Ceratolithus acutus*, ii) the FO of *Reticulofenestra cisnerosii*, iii) the FO of *Ceratolithus rugosus*, iv) the FO of “small Gephyrocapsa spp.”, v) the FO of *Pseudoemiliania lacunosa*, vi) the LO of *Reticulofenestra cisnerosii*, vii) the first common occurrence (FCO) of *Discoaster asymmetricus*, viii) the FO of *Discoaster tamalis*, ix) the LO of *Reticulofenestra pseudoumbilicus*, and x) the LO of *Sphenolithus* spp.

![FIGURE 3. Calcareous nannofossil events in the Bajo Segura Basin. The planktonic foraminifera scale is the Mediterranean eastern scale from Lourens et al. (2004). The NN zonation is from Martini (1971); the CN zonation is from Bukry (1975) and the emended CN is from Okada and Bukry (1980) and Bukry (1991), and the MNN is from Rio et al. (1990). Correlation between the different scales is based in Lourens et al. (2004). The Gulf of Lion column shows the reflectors Pr6 from Duvail et al. (2005); p7 and p9 from Rabineau (2001), and the seismic units: U1-a and U1-b from Lofi et al. (2003). The sequence terminology is based on Snedden and Liu (2011). Cronostratigraphy by Hilgen et al. (2012).](image-url)
FIGURE 4. 1A) *Reticulofenestra cisnerosii* 1500x sample DPH-20, plane light; 1B) *Reticulofenestra cisnerosii* 1500x sample DPH-20, crossed polars; 2A) *Reticulofenestra cisnerosii* sample PTEL-6, Scanning Electronic Microscope (SEM); 2B) *Reticulofenestra cisnerosii* sample PTEL-6, SEM; 3A) *Reticulofenestra cisnerosii* 1500x sample DPH-14, plane light; 3B) *Reticulofenestra cisnerosii* 1500x sample DPH-14, crossed polars; 4A) *Pseudoemiliana lacunosa* sample GUA-1, SEM; 4B) *Pseudoemiliana lacunosa* 1500x sample GUA-20, Parallel plane light; 5A) *Pseudoemiliana lacunosa* 1500x sample GUA-20, Crossed polars; 5B) *Pseudoemiliana lacunosa* 1500x sample GUA-20, Crossed polars; 6) *Pseudoemiliana lacunosa* sample GUA-1 SEM; 7) *Pseudoemiliana lacunosa* sample GUA-1 SEM; 8) *Discoaster tamaris* 1500x sample GUA-16, plane light; 9) *Discoaster tamaris* 1500x sample GUA-14, plane light; 10) *Discoaster tamaris* 1500x sample SP-3, plane light.
FIGURE 5. 1A) Discoaster asymmetricus 1500x sample GUA-16, plane light; 1B) Discoaster asymmetricus 1500x sample SPL-5, plane light; 2A) Ceratolithus acutus 1500x sample PTEL-6, plane light; 2B) Ceratolithus acutus 1500x sample PTEL-6, Crossed polars; 3A) Ceratolithus acutus 1500x sample GUA-1, plane light; 3B) Ceratolithus acutus 1500x sample GUA-1, Crossed polars; 4A) Ceratolithus armatus 1500x sample DPH-13, plane light; 4B) Ceratolithus armatus 1500x sample DPH-13, Crossed polars; 5A) Ceratolithus rugosus 1500x sample DPH-14, plane light; 5B) Ceratolithus rugosus 1500x sample DPH-14, Crossed polars; 6) Reticulofenestra pseudoumbilicus 1500x sample DPH-20, plane light; 7) Reticulofenestra pseudoumbilicus 1500x sample DPH-20 Crossed polars; 8) Reticulofenestra pseudoumbilicus sample GUA-1, SEM; 9) Sphenolithus abies 1500x sample DPH-13, Crossed polars; 10) Sphenolithus abies sample PE-148, SEM.
The FO of *Ceratolitus acutus* marks the CN10a/CN10b Boundary (Okada and Bukry, 1980), and has been considered as the standard event indicating the Messinian/Pliocene (M/P) boundary (Lourens et al., 2004). Shackleton and Crowhurst (1997), and Backman and Raffi (1997) proposed an age of 5.35 Ma for *C. acutus* FO in the equatorial Atlantic. In the Mediterranean, the FO of *C. acutus* was delayed until the beginning of the Pliocene (Cita and Gartner, 1973; Castradori, 1998; Van Couvering et al., 2000) because of the isolation of the basin during the MSC. In the Mediterranean, recordings of the FO of *C. acutus* are rare. Sometimes broken and overgrowth *Amaurolithus primus* (showing birefringence) may be mistaken for *C. acutus*. Lancis (1998) uses these overgrowth *Amaurolithus* forms to mark the uppermost marine Messinian. *C. acutus* has been found sporadically in the PE, DPH, and PTEL sections.

The FO of the abundant *Reticulofenestra cisnerosii* has been proposed by Lancis (1998), and Lancis and Flores (2006) as an alternative event to the FO of *C. acutus* for the M/P boundary in the Mediterranean basins. Although the FO of *R. cisnerosii* was initially identified in chron C3n.4n, Soria et al. (2008a) in a later study recalibrated this event as C3r in the PE section. Also, Di Stefano and Sturiale (2010) have reported this event in the Mediterranean basin as the FO of *Reticulofenestra zancleana*, a junior synonym of *Reticulefenestra cisnerosii* (Lancis, 1998; Lancis and Flores, 2006). This form is frequent in the sections studied (see graphs Fig. 7). Its FO is coincident with the FO of *C. acutus* in the Mediterranean. Its absence in the conglomeratic levels of the bottom of the PE section is due to its particular sedimentary environment. Its LO slightly predates the first occurrence of *Discoaster tamalis* and the FCO of *Discoaster asymmetricus*.

The FO of *Ceratolithus rugosus* marks the base of the CN10c subzone (Okada and Bukry, 1980) and the NN13 zone (Martini, 1971) occurring in the Equatorial Atlantic during the C3n.4n (Thvera) chron at 5.05 Ma, and in the Equatorial Pacific at 5.12 Ma (Backman and Raffi, 1997; Lourens et al., 2004; Raffi et al., 2006). It is also a rare event in the Mediterranean. Thus, Rio et al. (1990) used the drop in abundance of the fairly continuous *Amaurolithus* spp., and the FO of *Helicosphaera sellii* (the increase in the species to a frequency >1% in a count of 100 helicoliths) in their Mediterranean biostratigraphic scheme to mark the bottom of their MNN13 zone. Our data confirm the
The scarcity of *C. rugosus* since we found it only sporadically in the DPH, SM, and GUA sections.

Rio (1982) reported an increase in “small *Gephyrocapsa* spp.” in the Mediterranean Sea around 3.5 to 3.6 Ma. Dermitzakis and Theodoridis (1978) observed the first forms at the bottom of the NN13 of Martini (1971). Driever (1988), Lourens et al. (1996), and Lourens et al. (2004) established an age of 4.33 Ma for the FO of “small *Gephyrocapsa* spp.” for the Mediterranean, and found it slightly below the FCO of *D. asymmetricus*. In our sections, this event is found to almost coincide with the FO of *C. rugosus*, at the bottom of the NN13 zone (Martini, 1971).

The appearance of *Pseudoemiliania lacunosa* in the stratigraphic record is not normally used in current calcareous nannofossil biostratigraphic schemes because of discrepancies regarding its FO among the different investigators. Gartner (1969), Raffi and Rio (1979), Martini (1979), and Young et al. (1994) place the FO of *Pseudoemiliania lacunosa* close to the LO of *Reticulofenestra pseudoumbilicus* in the NN15 zone (Martini, 1971). For Rio et al. (1990) and Raffi et al. (2006) *P. lacunosa* appears with low frequency between the FCO of *D. asymmetricus*, and the LO of *R. pseudoumbilicus*. By contrast, Driever (1988) and Dermitzakis and Theodoridis (1978) observed it from the base of the NN13 zone (Martini, 1971). In our samples the FO of *Pseudoemiliania lacunosa* was found slightly above the FO of *C. rugosus* and can thus be included in the NN13 zone (Martini, 1971), this FO being an easily recognizable and useful event in the Eastern Betic basins for correlation purposes.

The FCO of *D. asymmetricus* marks the boundary between the NN13 and the NN14 zones (Martini, 1971), and also between the CN11a and the CN11b zones (Okada and Bukry, 1980). Rio et al. (1990) have defined this event as the point at which *D. asymmetricus* reached a >5% frequency in a count of 100 discoasters.

**FIGURE 7.** Quantitative distribution patterns of selected nannofossils in the Pantano de Elche (PTEL) section. Percentages are relative to 500 specimens >3 μm counted (500%). The percentage of “small *Gephyrocapsa*” is calculated relative to all the nanoliths (%). The *Discoaster tamalis* and *Discoaster asymmetricus* percentage is calculated relative to all the asteroliths (asteroliths %). See section Methods. Note differences in scaling.
and they used it as the boundary between their MNN13 and the MNN14 zones. In the Mediterranean, Rio et al. (1990) observed this event just above the C3n.1n subchron. Recently, in the Equatorial Pacific FCO of D. asymmetricus has been dated at 4.13Ma (Raffi and Flores, 1995; Shackleton et al., 1995) and in the Mediterranean at 4.12Ma (Lourens et al., 1996; Driever, 1988; Lourens et al., 2004). The DPH and PTEL sections exhibit this event, and the GUA and SP sections record D. asymmetricus from the bottom.

It is hard to establish the FO of Discoaster tamalis owing to the scarcity of this form in the samples. For the Mediterranean it is calibrated at 3.97Ma (Lourens et al., 2004), slightly younger than the FCO of D. asymmetricus. The FO of D. tamalis is found in the GUA and SP sections.

Raffi and Rio (1979) and Rio et al. (1990) placed the LO of R. pseudoumbilicus in the sample in which frequency of the species, for forms larger than 7µm, was below 2% of the total assemblage in a count of 500 nannofossil specimens. However, as mentioned by Rio et al. (1990) its detection in the sections is complicated owing to reworking. This event appears in the GUA section below the LO of Spenolithus spp. It also marks the boundary between the NN15 and the NN16 zones of Martini (1971), and the CN11b and the CN12aA zones of Okada and Bukry (1980) and has been
calibrated in the Mediterranean at 3.84 Ma (Lourens et al., 1996; Lourens et al., 2004; Raffi et al., 2006).

The last event is the LO of *Sphenolithus* spp., occurring above the LO of *R. pseudoumbilicus*. It has been used to split the lower CN12a into CN12aA/CN12aB (Okada and Bukry, 1980 emend. Bukry, 1991). The LO of *Sphenolithus* spp. has been calibrated at 3.65 Ma in the Equatorial Pacific (Raffi and Flores, 1995; Shackleton et al., 1995; Lourens et al., 2004; Raffi et al., 2006), and at 3.52–3.56 Ma in the Equatorial Atlantic (Raffi et al., 2006). It has been found in the uppermost part of the GUA section.

**Sequence dating**

From bottom to top, the events recognized in the lower sequence, including the paleo-valley fill (Pedrera Formation), the lower Hurchillo marls and the lower
yellow calcareous sandstone are (Figs. 3; 4; 5): i) the FO of *Ceratolithus acutus*; ii) the FO of *Reticulofenestra cisnerosi*; iii) the FO of *Ceratolithus rugosus*; iv) the FO of "small Gephyrocapsa spp."; v) the FO of *Pseudoemiliana lacunosa*; vi) the LO of *R. cisnerosi* and vii) the FCO of *Discoaster asymmetricus*. These allow it to be situated in the NN12-NN14 biozones of Martini (1971), the CN10b-CN11b zones of Okada and Bukry (1980), and the MNN12b-MNN14 zones of Rio et al. (1990).

The upper sequence, including the upper Hurchillo marls and the upper yellow calcareous sandstone, yields: viii) the FO of *Discoaster tamalis*; iv) the LO of *Reticulofenestra pseudoumbilicus* and x) the LO of *Sphenolithus spp.* This sequence can be dated as the upper part of NN15-NN16 of Martini (1971), CN11b-CN12a of Okada and Bukry (1980), and MNN15 to MNN16 of Rio et al. (1990).

Both sequences can be considered as Highstand System Tracts and the boundary between them a relative sea-level fall. The top of the lower sequence has at least two emersion episodes (see the Transgressive Pliocene header description of PE, SM, and DPH). The data on the Bajo Segura indicate that the discontinuity between both

---

**FIGURE 10.** Quantitative distribution pattern of selected nannofossils in the San Miguel (SM) section. Percentages are relative to 500 specimens >3µm counted (500%). The percentage of "small Gephyrocapsa" is calculated relative to all the nanoliths (%). See section Methods. Note differences in scaling.
sequences could be calibrated as the upper part of the NN14 to the middle part of the NN15 of Martini (1971), within the CN11b of Okada and Bukry (1980), and also the upper part of the MNN14 to the middle part of the MNN15 of Rio et al. (1990) (Fig. 3). The sequence boundary occur after the FCO of Discoaster asymmetricus found in the uppermost sediments of the lower sequence and before the FO of Discoaster tamalis in the lowermost part of the upper sequence. Thus the age of this sequence boundary can be estimated between 4.1 and 4.0 Ma ago.

**PALEOECOLOGICAL CONSIDERATIONS**

Coccolithophore abundances (Figs. 6; 7; 8; 9; 10; 11) were used to reconstruct the paleoenvironmental conditions of the Early Pliocene in the basin. The abundance of *R. pseudoumbilicus* and asteroliths has been related to warm and relatively deep water (Bukry, 1981; Rio and Sproveri, 1986; Driever, 1988; Rio et al., 1990; Lancis, 1998). Additionally, the intermediate and small reticulofoenestrid forms indicate the proximity of the coast. The lower parts of both sequences point to these relatively open marine conditions and an upward-trending restriction.

**CORRELATION WITH OFFSHORE DATA**

Many authors (Rabineau, 2001; Lofi et al., 2003; 2005; Gorini et al., 2005; Duvail et al., 2005; Garcia et al., 2011; Urgeles et al., 2011; Martinez del Olmo, 2011b; Leroux, 2013), using the seismic lines around the Western Mediterranean Basin, have described the post-Messinian sediments as multi- hectometric prograding prisms. The offshore study of the thick Early Pliocene sediments filling the MES in the Gulf of Lion by Lofi et al. (2003) differentiated a major prograding seismic unit U1, divided into two seismic subunits (U1-a, the lower, and U1-b) deposited beneath the modern inner to middle shelf, showing a shallowing-upward trend with a major upper boundary interpreted as a major erosional unconformity formed during a relative fall in sea-level. Leroux (2013) correlates the sequence boundary between both seismic subunits with the p7 reflector of Rabineau (2001), and the Pr6 reflector of Duvail et al. (2005) of the Gulf of Lion (Fig. 3). The sediments included in the U1-a seismic unit of Lofi et al. (2003) have been mapped by Leroux (2013, Messinian top to p7 sediments), showing that they are restricted to depths between 100 to 2000 m below today sea-level in the Gulf of Lion.

From well data by Lofi et al. (2003), U1 was dated as Early Pliocene, but not ruling out the earliest Late Pliocene. The age boundary between both seismic subunits is not sufficiently well constrained, being estimated by Lofi et al. (2003) to lie between the MPL2 to MPL3 planktonic foraminifera biozones (Lourens et al., 2004). Considering both, the Gulf of Lion seismic subunits as well as the Bajo Segura sedimentary units filling up the MES, a correlation between them could be proposed. The lower Hurchillo sedimentary cycle may be the onland equivalent of the U1-a seismic unit and the upper Hurchillo sedimentary cycle that of U1-b seismic unit (Fig. 3).

The age of the U1 seismic unit mainly coincides with the set of the two Bajo Segura sedimentary units but the age boundary between the two seismic subunits is not sufficiently well constrained, being estimated by Lofi et al. (2003) to lie between the two Bajo Segura sedimentary units. This could be explained in terms of the intrinsic errors arising from the broad correlation between the seismic units and the well data used in the dating of the U1 subunits (Lofi et al., 2003). The sequence boundary between the two Bajo Segura sedimentary units could be correlated with the Za1-Za2 sequence boundary (Snedden and Liu, 2011). Since the U1 has been dated as Lower Pliocene, the fall in sea-level that occurred around its upper boundary could have been close to the Lower to Upper Pliocene boundary (Lofi et al., 2003). However, in the Bajo Segura Basin above the upper Hurchillo marls further marine sedimentation is only recorded in the high sea-levels during the Quaternary, such as the Holocene (Soria et al., 1999).

**DISCUSSION**

The lowermost Pliocene sediments correspond to a high-energy paleo-valley fill (García-García et al., 2011). Outside...
the paleo-valley, where the MES is exposed as a gentle basin-scale surface, a wide hiatus separates the transgressive Pliocene deposits from the Messinian ones. The most seaward section, the GUA, and the seaward and elevated portions of the paleorelief, seen at SP, show a glauconitic level (Lancis et al., 2004b), that may be interpreted as condensed sections with a low sedimentation rate. These levels mark areas of the basin where the oceanic currents winnowed the coccoliths. This could explain the absence of the earliest Pliocene nannofossil bioevents in the GUA and SP sections. The PE and DPH are the most proximal sections studied in this work and they show the upper sequence to be more transgressive over the lower coastline, as seen in the seismic profiles (see Lofi et al., 2003). At this point, no quantification of the landward migration of the coastline in the upper transgression can be estimated because of the uncertainties about the rate of sea-level increase in an anomalous environment with high subsidence (after extensive flooding of the platforms), and with a highly available sediment supply (after the increased erosion of the Mediterranean landscapes during the Messinian). Additionally, the SM section shows thinner upper Huchillo marls, probably due to a local uplift of this section, located in a hinge of an anticline (Soria et al., 2008a), which would have compensated the second marine transgression.

CONCLUSIONS

Six stratigraphic sections located in key points of the Bajo Segura Basin show that the base of the Lower Pliocene sediments resting on the MES are diachronous. These sediments correspond to two transgressive-regressive sequences, the lower Huchillo sedimentary cycle and the upper Huchillo sedimentary cycle. These two transgressive-regressive sequences can be correlated with the two lower Pliocene seismic sequences infilling the inner to middle Gulf of Lion shelf on the Western Mediterranean (Lofi et al., 2003; Leroux, 2013). The upper transgression arrived farther inland than the lower one. Future work should focus on explaining the cause of this younger Early Pliocene transgression and quantifying the coastline migration.

ACKNOWLEDGMENTS

D. Violanti is acknowledged for comments on an early version of this paper. We would like to acknowledge also the editor M. Garcés and two anonymous referees for some interesting suggestions. This work has been supported by projects: CGL2007-65832/BTE Ministerio de Educación y Ciencia, CGL2009-07830/BTE Ministerio de Ciencia e Innovación, and PASUR.CGL2009-08651 Ministerio de Ciencia e Innovación Projects and BEST/2010/068 Generalitat Valenciana.

REFERENCES


Manuscript received January 2014; revision accepted February 2015; published Online July 2015.