Deep incision in an Aptian carbonate succession indicates major sea-level fall in the Cretaceous

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ABSTRACT
Long-term relative sea-level cycles (0·5 to 6 Myr) have yet to be fully understood for the Cretaceous. During the Aptian, in the northern Maestrat Basin (Eastern Iberian Peninsula), fault-controlled subsidence created depositional space, but eustasy governed changes in depositional trends. Relative sea-level history was reconstructed by sequence stratigraphic analysis. Two forced regressive stages of relative sea-level were recognized within three depositional sequences. The first stage is late Early Aptian age (intra Dufrenoyia furcata Zone) and is characterized by foreshore to upper shoreface sedimentary wedges, which occur detached from a highstand carbonate platform, and were deposited above basin marls. The amplitude of relative sea-level drop was in the order of tens of metres, with a duration of <1 Myr. The second stage of relative sea-level fall occurred within the Late Aptian and is recorded by an incised valley that, when restored to its pre-contractional attitude, was >2 km wide and cut ≥115 m down into the underlying Aptian succession. With the subsequent transgression, the incision was backfilled with peritidal to shallow subtidal deposits. The changes in depositional trends, lithofacies evolution and geometric relation of the stratigraphic units characterized are similar to those observed in coeval rocks within the Maestrat Basin, as well as in other correlative basins elsewhere. The pace and magnitude of the two relative sea-level drops identified fall within the glacio-eustatic domain. In the Maestrat Basin, terrestrial palynological studies provide evidence that the late Early and Late Aptian climate was cooler than the earliest part of the Early Aptian and the Albian Stage, which were characterized by warmer environmental conditions. The outcrops documented here are significant because they preserve the results of Aptian long-term sea-level trends that are often only recognizable on larger scales (i.e. seismic), such as for the Arabian Plate.

Keywords Aptian, carbonate platforms, Iberian plate, incised valley, sea-level changes, sequence stratigraphy.

INTRODUCTION
Deciphering long-term relative sea-level fluctuations (0·5 to 6 Myr) in Aptian sedimentary successions has palaeoclimatic relevance, given that some of the amplitudes and frequencies of relative sea-level change reported for this stage are attributed to glacio-eustatic controls (e.g.
Aptian relative sea-level history of the north Maestrat Basin

Hillgärtner et al., 2003; Gréselle & Pittet, 2005; Bover-Arnal et al., 2009; van Buchem et al., 2010; Maurer et al., 2010, 2013; Husinec et al., 2012. Glacio-eustasy in the Aptian is controversial because of the classic, widespread idea of stable greenhouse conditions during the Cretaceous (e.g. Wilson & Norris, 2001; Skelton, 2003a; Hay, 2008, 2011). However, palynological data and oxygen-isotope analyses from Aptian strata and fossil biogenic components have revealed the existence of cold episodes (Krassilov, 1973; Weissert & Lini, 1991; Ditchfield et al., 1994; Solé de Porta & Salas, 1994; Stoll & Schrag, 1996; Abreu et al., 1998; Clarke & Jenkyns, 1999; Ferguson et al., 1999; Hochuli et al., 1999; Pucéat et al., 2003; Steuber et al., 2005; Amiot et al., 2011; Cors et al., 2013) which may have increased ice volume on the continents and subsequently lowered global sea-level. Further evidence supporting climate cooling events in the Aptian include ice-rafted debris and glendonites in high latitudes (Picton, 1981; Kemper, 1987; Frakes & Francis, 1988, 1990; De Lucio & Frakes, 1999; Price, 1999; Alley & Frakes, 2003); palaeomigrational patterns of high-latitude calcareous nannofossils towards lower latitudinal sites (Herrle & Mutterlose, 2003; Rückheim et al., 2006; Mutterlose et al., 2009) and, finally, some of the Aptian relative sea-level drops documented seem to be coeval (Hardenbol et al., 1998; Gréselle & Pittet, 2005; Bover-Arnal et al., 2009, 2010; van Buchem et al., 2010; Maurer et al., 2010, 2013; Husinec et al., 2012; Rameil et al., 2012), despite the limitations of chronostratigraphic frameworks mostly based on ammonite, rudist, planktonic foraminiferal and orbitolinid biostratigraphy.

When reconstructing the relative sea-level history of a sedimentary succession, the recognition of emersion surfaces is of key importance since these are commonly taken as the basis of the chronological framework. In the sequence stratigraphic models of Vail et al. (1977), Posamentier et al. (1988), van Wagoner et al. (1988) and Hunt & Tucker (1992), subaerial unconformities, and their marine correlative conformities, are regarded as primary sequence boundaries. In these four models, exposure surfaces are associated with stratigraphic hiatuses resulting from non-deposition and erosion during base-level fall.

In ancient subaerially exposed carbonate systems, sedimentological or diagenetic evidence of emersion is at times absent, or can be very subtle or ambiguous (e.g. Hillgärtner, 1998; Immenhauser et al., 1999; Raven et al., 2010; Rameil et al., 2012). The preservation of a subaerially exposed surface depends on the duration of the exposure time, the magnitude of the relative sea-level fall, the resistance of the rocks exposed to physical and chemical alteration, and the intensity in which the atmospheric agents, biological activity, hydrological processes and pedogenesis act. Furthermore, with subsequent marine onlap, evidence of subaerial exposure are often concealed or removed by marine ravinement and by the activity of organisms. On the other hand, however, seismic-scale incisions down-cutting carbonate platforms, with a peritidal backfill, constitute an unequivocal sign of relative sea-level fall and subaerial exposure of the platform (e.g. Bover-Arnal et al., 2009, 2010, 2011a; Raven et al., 2010; Maurer et al., 2013).

Stratigraphic analysis of the Aptian sedimentary record of the northern Maestrat Basin (Iberian Chain; East Iberian Peninsula) permits the recognition of three differentiated stages of carbonate platform development, terminated by drowning or subaerial exposure. Associated with a Late Aptian emersion phase, a conspicuous seismic-scale incision down-cuts the Aptian succession by 115 m. This incision, which is filled with peritidal to shallow subtidal deposits, provides conclusive evidence of the amplitude of the long-term relative sea-level changes controlling accommodation during the Aptian in the northern part of the Maestrat Basin, and thus deserves special consideration.

The aims of the present study were to: (i) reconstruct the Aptian long-term relative sea-level history of the northern Maestrat Basin; (ii) measure the amplitude and duration of the relative sea-level falls recognized in order to test glacio-eustasy as a potential controlling mechanism; (iii) generate a two-dimensional (2D) profile of the seismic-scale incision affecting the Aptian succession to measure its dimensions and to understand its shape; (iv) analyse the stratal architecture and lithofacies evolution of the carbonate systems to attain a depositional model for these platform carbonates; and (v) compare and contrast the results with other Aptian sedimentary systems, distinguishing regional and potential global patterns. Ultimately, this research documents an important piece of the Aptian relative sea-level history and carbonate platform development from the northern margin of the Tethys. The results are of potential interest for future refinements to the Aptian eustatic sea-level curves.
GEOLOGICAL SETTING

The Maestrat Basin developed as a result of Late Jurassic (Late Oxfordian)–Early Cretaceous (Middle Albian) continental rifting that affected the Iberian plate (Salas & Casas, 1993; Salas et al., 2001). Rifting was related to the spreading of the Central Atlantic domain and opening of the Bay of Biscay. During extension, the Maestrat Basin was compartmentalized into seven sub-basins, where kilometre-thick successions of mixed carbonate–siliciclastic deposits accumulated in continental to hemipelagic depositional environments (Canérot et al., 1982; Salas, 1987).

As a consequence of the Alpine contraction during the Late Eocene–Early Miocene, the Maestrat Basin was tectonically inverted, which gave rise to the eastern part of the Iberian Chain (Fig. 1; Guimerà, 1984; Salas & Casas, 1993; Salas et al., 2001).

The Aptian strata studied in the northern Maestrat Basin are divided into three formations: Forcall, Villarroya de los Pinares and Benassal (Canérot et al., 1982; Salas, 1987; Salas et al., 2001; Fig. 2). This succession was deposited in tropical to subtropical neritic marine settings, and is characterized by limestone and marl lithologies, which contain orbitolinid benthic foraminifera, rudist bivalves, corals and ammonites. The Forcall and Villarroya de los Pinares formations are Early Aptian in age (Canérot et al., 1982; Salas, 1987; Salas et al., 2001; Bover-Arnal et al., 2010, 2011b; Moreno-Bedmar et al., 2010), while the Benassal Formation spans the latest Early Aptian–Late Aptian interval (Bover-Arnal et al., 2012; Moreno-Bedmar et al., 2012a; García et al., 2014; Fig. 2).

In the study area, the Aptian record is capped by a regional erosional unconformity overlain by Miocene conglomerates (Fig. 1C). For a detailed geological background of the Maestrat Basin and its sub-basins, refer to Canérot et al. (1982), Salas (1987), Salas & Casas (1993), Salas & Guimerà (1996) and Salas et al. (2001).

In the Morella Sub-basin (northern Maestrat Basin; Salas & Guimerà, 1996; Fig. 1B), the Forcall, Villarroya de los Pinares and Benassal formations crop out along five hillocks (the so-called ‘Moles’): Mola de Morella, Mola de la Vila, Mola d’en Camarás, Mola de la Garumba and Mola de la Saranyana (Fig. 1C). The outcrops are located in the surroundings of the Morella, el Forcall, la Todolella and Xiva de Morella towns (Comarca of Els Ports).

MATERIALS AND METHODS

A geological map of the Mola d’en Camaràs (Fig. 1) was made to produce a 2D geological cross-section of the seismic-scale incision that down-cuts Aptian strata. The map was created, in a georeferenced frame, over a colour orthophoto at scale 1 : 5000 (0.5 m per pixel) by the Spanish Centro Nacional de Información Geográfica (sheet 545; available at http://www.cnig.es). Elevations were procured from the 1 : 25 000 topographic map (sheet 545; 10 m contour line interval, also available at http://www.cnig.es). Line drawing and mapping of lithofacies architecture and key stratigraphic surfaces were performed on panoramic photomosaics of the steep cliffs delimiting the five hillocks studied. The different lithofacies identified were sampled to produce thin sections (n = 65) for the analysis of the microfossils. The terminology used for rock textures follows the classification schemes of Dunham (1962) and Embry & Klovan (1971).

Sequence stratigraphic interpretation is based on the ‘four-system-tract’ model of Hunt & Tucker (1992), but follows the proposed standard sequence stratigraphic nomenclature by Catuneanu et al. (2009). The duration of long-term cyclic variations in depositional trends characterized is consistent with the third-order relative sea-level cycles of Vail et al. (1991). Biostratigraphic analysis of ammonites (Moreno-Bedmar et al., 2009, 2010, 2012a; García et al., 2014), rudists (Skelton, 2003b; Skelton & Gili, 2012) and orbitolinids (Schroeder et al., 2010; Cherchi & Schroeder, 2012) was used to calibrate the relative age of the strata studied.

LITHOFACIES CLASSIFICATION

The different lithofacies identified in the field and in thin sections are grouped into five major facies associations, which represent distinct depositional environments and correspond to seismic-scale building blocks of the overall stratigraphic architecture (Fig. 3). Classification is based on the recognition of clinoform surfaces and facies heterogeneity along platform to slope to basin depositional profiles. This large-scale lithofacies evolution pattern repeats throughout the Aptian record studied; only interrupted by the occurrence of platform-detached high-energy wedges and the sedimentary deposits that backfill the incision down-cutting platform carbonates.
Fig. 3. The geographical positions of the outcrop views, facies details and microfacies shown in this paper are indicated in Fig. 4.

Platform lithofacies association

The main lithofacies generated in platform-top settings are decimetre-thick to tens of metres-thick,
light grey, floatstone to rudstone limestones characterized by the presence of wholly preserved rudist bivalves (Fig. 5A). The limestones display massive and tabular bedding. Rudists are commonly preserved in life position. Frequently, some species of elevator rudists such as *Polyconites hadriani* (Skelton et al., 2010) occur grouped in bouquets, or more rarely give rise to decimetre-thick biostromes. Scattered centimetric to decimetric corals in growth position exhibiting branching, sheet-like, platy, domal and irregular massive morphologies are also present. At times, corals display *Gastrochaenolites* borings, which on occasion preserve the in situ valves of the lithophagid bivalve. Skeletal fragments of rudists and corals are abundant. Saddle dolomite within rudist shells and corals occurs. Other common constituents of this lithofacies are peloids, the peyssonneliacean red alga *Polysphaera alba* encrusting skeletal components, nerineid gastropods, other gastropods, orbitolinids, other benthic foraminifera, sessile foraminifera attached to bioclasts, serpulids, *Chondrodonta*, oysters, and angular to sub-angular fragments of molluscs, echinoids and dasycladaceans. The skeletal components frequently appear highly bioeroded.

Also included in this lithofacies assemblage are: (i) mudstones; (ii) orbitolinid-rich limestones; (iii) coral-bearing limestones; and (iv) skeletal–peloidal limestones. The mudstones constitute decimetre-thick to metre-thick massive whitish beds and include scattered orbitolinids, lituolids and other benthic foraminifera. The orbitolinid-rich limestones are light grey-coloured. These limestones have thicknesses ranging from centimetres to decimetres and exhibit floatstone to rudstone textures, which also contain miliolids, other foraminifera, peloids and angular to sub-angular fragments of molluscs. The coral-bearing limestones form decimetre-thick to metre-thick, light grey,
nodular beds. The corals are found in growth position and corresponded to level-bottom communities, which display an unbound growth fabric with irregular massive, platy and domal centimetric to decimetric forms. Scattered rudist shells, as well as peloids, benthic foraminifera and fragments of echinoids and molluscs are also common components of these coral-rich levels. The skeletal–peloidal limestones correspond to light grey tabular beds. Occasionally, these deposits display cross-bedding features. The beds have thicknesses ranging from centimetres to decimetres and exhibit poorly sorted to moderately sorted packstone to grainstone textures. Peloids are abundant. Orbitolinids, miliolids, other benthic foraminifera, *Polystrata alba* encrusting bioclasts, and angular to rounded fragments of oysters, other bivalves, gastropods, bryozoans, corals and echinoids, dominate the skeletal constituents. Rarely, wholly preserved *Toucasia* specimens are found at the top of the beds of this sub-lithofacies.

**Slope lithofacies association**

This lithofacies assemblage is mainly characterized by centimetre-thick to metre-thick, light grey, nodular-bedded deposits shed from the platform top. Ochre marly limestone lithologies are also common. On occasion, the beds display channel-filled sedimentary bodies, slump scars and erosive surfaces. The limestones exhibit very poorly sorted floatstone and rudstone textures, which mainly contain fragments of rudists and corals. The fragmented skeletal parts are frequently bioeroded and mainly show angular edges. Decimetric wholly preserved isolated coral colonies in growth position are also important constituents of this lithofacies. Similar coral populations have been analysed for the same time interval in slope settings from the western Maestrat Basin (Bover-Arnal *et al.*, 2012). The colonies display platy (Fig. 5B), sheet-like, branching, domal and irregular massive morphologies. *Gastrochaenolites* on corals are frequent. Rhynchonellid brachiopods, and elevator rudists such as *Polyconites hadriani* (Skelton *et al.*, 2010) in life position, which sometimes show a clustered growth, are also occasionally present. Other important accompanying components of the lithofacies are the same as those detailed in the platform lithofacies association for the floatstone and rudstone limestones dominated by rudists.

Orbitolinid-rich centimetre-thick to decimetre-thick, light grey, limestones and marly limestones are also found on occasion in slope settings. These levels exhibit floatstone and rudstone textures and, besides orbitolinids, include peloids, miliolids, other foraminifera and angular to sub-rounded fragments of molluscs.

**Basin lithofacies association**

The lithofacies association is composed of bluish and ochre marls up to tens of metres-thick with occasional interbedded centimetre-thick to decimetre-thick marly limestones and carbonate nodules. The ochre colour is due to oxidation. The macrofossils are dominated by rhynchonellid and terebratulid brachiopods, ammonites, oysters and other bivalves (Fig. 5C). Other common macrobiota present are serpulids, belemnites, vermetid gastropods, other gastropods, crinoids, irregular echinoids and fragments of decapods. Nannoconids, and benthic and planktonic foraminifera, also occur. Pyritized skeletal components are frequent. The marl lithology together with the fossil content, which is characterized by the presence of ammonoids, indicates deeper neritic environments when compared to the lithofacies assemblages deposited in platform and slope settings.

**Platform-detached wedge lithofacies association**

Ochre and light grey decimetre-thick to metre-thick poorly sorted to moderately sorted grainstones (Fig. 5D) form this lithofacies. The limestones are massive or display low-angle cross-bedding, trough cross-bedding or plane-parallel stratification. Burrows occur. The main constituents of the lithofacies are peloids, coated grains, orbitolinids, other benthic foraminifera, small nerineid gastropods, and sub-rounded to rounded fragments of oysters, other bivalves, gastropods, echinoids, siphonaceous green algae, decapods, serpulids and bryozoans. The orbitolinids frequently show agglutinated silt-sized quartz particles. Saddle dolomite and bioerosional structures within the bioclasts are occasionally present. The presence of grainstone textures and sedimentary structures, such as plane-parallel stratification, and low-angle and trough cross-bedding, indicates a sedimentary environment most probably in the foreshore-upper shoreface (e.g. Chakraborty & Paul, 2008).
Incision backfill lithofacies association

This lithofacies assemblage mainly includes centimetre-thick to tens of metres-thick intervals of white marls, decimetre-thick ochre nodular marly limestones and centimetre-thick to metre-thick limestones. Scattered oyster specimens are occasionally embedded within the marls. In the marly limestones, oysters are abundant and frequently display a clustered growth. Rarely, scleractinian colonies in growth position occur. The limestones exhibit different textures and bedding patterns. These comprise: (i) white nodular mudstones, which occasionally display Thalassinoides, and contain scattered biseriate foraminifera, miliolids and fragments of echinoids; (ii) light grey, occasionally burrowed, wackestone to packstone textures with peloids, orbitolinids, other foraminifera, and fragments of serpulids, gymnocodiacean algae, echinoids, oysters and other molluscs; (iii) light grey and ochre floatstone to rudstone textures dominated by fragments of oysters, but also including rudist shells, fragments of other molluscs and echinoids, benthic foraminifera, peloids, unidentified burrows, lithoclasts (Fig. 6A) and mud pebbles and mud drapes (Fig. 6B); and (iv) ochre and light grey poorly sorted to well-sorted grainstones (Fig. 6C), which exhibit cross-bedding, herringbone cross-stratification (Fig. 6D), tidal bundles (Fig. 6E) and burrows, and contain abundant peloids, miliolids, orbitolinids, other benthic foraminifera and fragments of echinoids, decapods, algae, bryozoans, oysters, gastropods and other molluscs. The grainstones with hydrodynamic structures mainly characterize the upper part of the incision backfill sedimentary succession.

Sedimentary features, such as mud drapes and mud pebbles, the presence of lithoclasts, sedimentary structures such as tidal bundles, bi-directional current indicators and the fossil content, which is dominated by oysters, are in agreement with a peritidal succession. The occurrence of rare stratigraphic levels containing rudist bivalves or coral colonies in life position indicates occasional shallow subtidal conditions.

SEQUENCE STRATIGRAPHIC INTERPRETATION

The Aptian record of the northern Maestrat Basin is subdivided into three depositional sequences (A, B and C) based on the distribution of lithofacies, architecture of the strata and the
identification of sequence stratigraphic surfaces. A 2D continuous platform to basin conceptual profile displaying the sedimentary response for each depositional sequence of the long-term relative sea-level evolution is provided in Fig. 3. Figure 7 displays the position of each hillock (‘Mola’) studied within this conceptual model. A representative schematic log of the Mola de la Vila showing the vertical lithofacies evolution, the stratigraphic position of the age-diagnostic fauna and the sequence stratigraphic interpretation is provided in Fig. 8.

**Depositional Sequence A**

*Transgressive genetic type of deposit*

The transgressive unit comprises the basin lithofacies association (Figs 3A and 5C) and corresponds to the marls of the Forcall Formation (Fig. 2). During rapid creation of accommodation, lime mud and fine-grained siliciclastic particles were transported by suspension from proximal platform and coastal settings to the basin, and mixed to form the marls. The marl interval (Figs 9 and 10) is >100 m thick and contains ammonite taxa belonging to the four Early Aptian Zones: *Deshayesites oglanensis*, *Deshayesites forbesi*, *Deshayesites deshayesi* (Fig. 5C) and *Dufrenoyia furcata* (see Moreno-Bedmar *et al.*, 2010 and Garcia *et al.*, 2014 for species). The presence of the large benthic foraminifer *Palorbitolina lenticularis* distinguishes the lower part of the Forcall Formation, whereas the brachiopods *Burrirhynchia miliani* and *Sellithyris sella*, the bivalve *Plicatula placunea* (Fig. 5C) and the irregular echinoid *Toxaster*...
collegnoi are characteristic of the upper part of these transgressive marls.

**Highstand normal regressive genetic type of deposit**

A decrease in the rate of relative sea-level rise forced the carbonate production zone to prograde into the basin. This prograding carbonate system constitutes a highstand normal regressive genetic unit (Fig. 3A). The boundary between the transgressive and normal regressive deposits corresponds to an interpreted maximum flooding surface, which is placed at the contact between underlying marls of the Forcall Formation and overlying highstand platform carbonates of the Villarroya de los Pinares Formation (Figs 2 and 3A). In seismic stratigraphy, the maximum flooding surface is often interpreted as the surface downlapped by highstand prograding clinoforms (Catuneanu et al., 2011). This highstand platform, which is locally preserved at the Mola de Morella (Figs 1C, 7 and 11A), progrades and...
Fig. 6. Incision backfill lithofacies association. (A) Ochre limestone exhibiting a floatstone to rudstone texture with abundant fragments of oysters. Note the presence of a decimetre-sized lithoclast (upper left). Transgressive unit of Depositional Sequence C, Benassal Formation, Late Aptian, Mola d’en Camaràs. (B) Rudstone texture dominated by fragments of oysters. Note the presence of mud pebbles and mud drapes. Transgressive unit of Depositional Sequence C, Benassal Formation, Late Aptian, Mola d’en Camaràs. Visible part of pen = 7.5 cm. (C) Photomicrograph of a moderately sorted grainstone texture, which includes abundant peloids, orbitolinids and skeletal fragments of oysters, gastropods and echinoids. Transgressive unit of Depositional Sequence C, Benassal Formation, Late Aptian, Mola d’en Camaràs. (D) Herring-bone cross-stratification formed by aggradation during alternate tidal flow. Transgressive unit of Depositional Sequence C, Benassal Formation, Late Aptian, Mola d’en Camaràs. Ruler = 15 cm. (E) Outcrop view of a level with tidal bundles (centre). Transgressive unit of Depositional Sequence C, Benassal Formation, Late Aptian, Mola d’en Camaràs. Hammer length = 32 cm.
downlaps over basin marls (Fig. 3A), indicating that it is indeed a maximum flooding surface. The carbonates of the Mola de Morella correspond to proximal platform settings. These platform-top deposits passed basinwards to slopes, and these pinched out into the basin marls of the Forcall Formation (Figs 3A and 7). Due to the discontinuous nature of the outcrops and the absence of limestone beds as stratigraphic guides within the upper part of the Forcall Formation, the maximum flooding surface cannot be mapped towards the basin (at the other hilllocks; Fig. 7). Therefore, in basin settings, the concept of a maximum flooding zone will be applied (see Strasser et al., 1999). The maximum flooding zone is interpreted to be within the upper part of the Forcall Formation (Figs 3A, 8, 9C and 10C), given that the uppermost part of the marl succession corresponds to basin sediments deposited during the highstand stage of relative sea-level (Fig. 3A).

Highstand carbonates of the Mola de Morella (Figs 1C, 7 and 11A) are ca 80 m thick. The top of this platform succession is not preserved due to Neogene and present-day erosion. Therefore, signs of subaerial exposure of Aptian age cannot be recognized. The highstand platform contains abundant *Chondrodonta*, as well as polyconitid, requieniid and caprinid rudists (Fig. 12A), probably including *Caprina parvula*. The presence of Caprinidae indicates an Early Aptian age (Skelton & Gili, 2012).

**Forced regressive genetic type of deposit**

The highstand normal regressive platform of the Mola de Morella was subaerially exposed and shutdown as a result of base-level fall (Figs 3B and 7). During falling relative sea-level, the area available for carbonate production and accumulation was progressively reduced and shifted basinwards, where detached skeletal wedges were deposited on top of basin marls of the Forcall Formation (Figs 3B, 7, 8, 11B and 13). These forced regressive deposits belong to the Villarroya de los Pinares Formation (Fig. 2) and are made up of the platform-detached wedge lithofacies association (Figs 3B, 5D and 11B). Biota identified at species or genus level include *Choffatella decipiens*, *Lenticulina* sp., *Praeorbitolina* sp., *Mesorbitolina lotzei* and *Boueina hochstetteri*. The presence of *Praeorbitolina* and *Mesorbitolina lotzei* implies an Early Aptian age (Schroeder et al., 2010; Cherchi & Schroeder, 2012). Forced regressive deposits are only preserved in the Mola de la Vila, Mola de la Garumba and Mola de la Saranyana (Figs 1C, 4, 7, 8, 11B and 13).

The detached wedges are bound below by regressive surfaces of marine erosion and above by the marine correlative conformity of the subaerial unconformity (Figs 3B, 7, 8, 11B and 13C). The regressive surfaces of marine erosion were formed during base-level fall by wave and sediment scouring of the highstand basin deposits. The correlative conformity marks the lowest point...
of relative sea-level, and thus the most basinwards extension of the subaerial unconformity. The subaerial unconformity and its marine correlative conformity bound depositional sequences A and B, and correspond to Sequence Boundary 1 (Figs 3B, 7, 8, 9C, 10C, 11B and 13C).

Depositional Sequence B

Lowstand normal regressive genetic type of deposit

After the onset of subsequent base-level rise, accommodation increased and a lowstand normal regressive carbonate platform developed. This platform onlaps slopes of the highstand platform of Depositional Sequence A and downlaps onto the correlative conformity; either over basin marls (Figs 3B, 9C and 10C) or over forced regressive units (Figs 3B, 8, 11B and 13C). The lowstand deposits are ca 10 m thick, belong to the Villarroya de los Pinares Formation and are formed by the platform lithofacies, which pass basinwards to slope lithofacies (Figs 3B, 13B, 13C, 14A, and 14B). The corresponding lowstand basin lithofacies are not preserved due to present-day erosion.

The lowstand platform displays prograding clinoforms towards the north (Fig. 14A and B) and channel-fill geometries, which drain northwards as well (Fig. 13B and C). Characteristic fauna of the lowstand platform identified are Praeorbitolina, Mesorbitolina lotzei, Nautiloculina bronnimanni, and requieniid, polyconitid and caprinid (Fig. 12B) rudists. The presence of Praeorbitolina, Mesorbitolina lotzei and Caprinidae indicates an Early Aptian age (Schroeder et al., 2010; Cherchi & Schroeder, 2012; Skelton & Gili, 2012).

Transgressive genetic type of deposit

With the end of the stage of normal regression and the onset of transgression, the lowstand platform was drowned and buried by transgressive marl deposits that correspond to the basin lithofacies association (Figs 3C, 8, 10C, 14A and 14B). A well-developed hardground with widespread Fe–Mn crusts, encrusting oysters (Fig. 14C) and borings of lithophagid bivalves marks the drowning of the platform. This drowning surface corresponds to a maximum
regressive surface, which is observed as both smooth and irregular. In some parts of the Mola de la Garumba, a transgressive lag (Fig. 14D), preserved as a packstone containing abundant fragments of decapods and gastropods, fills the depressions occasionally exhibited by this surface.

The basal part of the transgressive marl interval, which corresponds to the lower part of the Benassal Formation (Fig. 2), is characterized by the presence of *Dufrenovia dufrenoiyi* specimens (Fig. 12C to E). Therefore, the basal part of these transgressive deposits is Early Aptian in age (Figs 2, 3D and 8). Another common organism identified in the marls is *Panopea* sp.

**Highstand normal regressive genetic type of deposit**

With the onset of regression, a large carbonate platform (Figs 3C, 8, 10 and 15) developed (Benassal Formation). The lower part of this carbonate system is distinguished by beds composed of the slope lithofacies association (Fig. 8). The corresponding basin lithofacies is not preserved. The maximum flooding surface of the sequence is interpreted following the same criterion used for the maximum flooding surface of

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*Fig. 9. Sequence stratigraphy. (A) Photomosaic of the north-eastern face of the Mola d’en Camaràs. The area outlined in yellow corresponds to the close-up view provided in (B). Width of image is ca 2 km. (B) Close-up view of (A). (C) Sequence stratigraphic interpretation of (B). Note how the lowstand platform (Villarroya de los Pinares Formation) of Depositional Sequence B is incised and onlapped by the transgressive deposits of Depositional Sequence C. See Fig. 3 for key. A close-up view of the area outlined in black is shown in Fig. 16.*
Fig. 10. Sequence stratigraphy. (A) Photomosaic of the north-western face of the Mola d’en Camaràs. The areas outlined in yellow and red correspond, respectively, to the close-up view provided in panel (B) and to the situation of the highstand normal regressive platform of Depositional Sequence A shown in Fig. 11A. Width of image is ca 1.5 km. (B) Close-up view of (A). (C) Sequence stratigraphic interpretation of (B). Note the significant thickness exhibited by the transgressive unit (marls of the Forcall Formation) of Depositional Sequence A. See Fig. 3 for key.
Depositional Sequence A; at the limit between the transgressive marls and the downlapping slopes of the highstand normal regressive platform (Figs 3C, 8, 10C and 15C). The lower platform-top carbonates (platform lithofacies association) are massive and thick, while the

Fig. 11. Highstand normal regressive platform and forced regressive deposits of Depositional Sequence A (Vilarroya de los Pinares Formation). (A) Panoramic view of the town of Morella with its castle built on top of the preserved highstand normal regressive platform carbonates of Depositional Sequence A. Width of image is ca 700 m. (B) Outcrop view of a platform-detached wedge showing trough cross-bedding. Forced regressive unit of Depositional Sequence A, Vilarroya de los Pinares Formation, Mola de la Garumba. Below the platform-detached wedge, highstand normal regressive marls of Depositional Sequence A (Forcall Formation, Early Aptian) are visible. The downlapping slopes of the lowstand normal regressive platform of Depositional Sequence B (Vilarroya de los Pinares Formation, Early Aptian) are observable at the top of the image. See Fig. 3 for key. Hammer length = 32 cm.
Fig. 12. Age-diagnostic fossils. (A) Detail of mutually attached caprinid rudists. Platform lithofacies association, highstand normal regressive deposits of Depositional Sequence A, Villarroya de los Pinares Formation, Early Aptian, Mola de Morella. Visible part of pen = 3.3 cm. (B) Close-up view of a section of a caprinid rudist (Caprina aff. parvula). Platform lithofacies association, lowstand normal regressive unit of Depositional Sequence B, Villarroya de los Pinares Formation, Early Aptian, Mola de la Garumba. Visible part of pen = 2.3 cm. (C) and (D) Dufrenoyia dufrenoyi, lateral and ventral views of the specimen PUAB (Coleccions de Paleontologia de la Universitat Autònoma de Barcelona) 88390. Basin lithofacies association, transgressive deposits of Depositional Sequence B, Benassal Formation, Early Aptian, Mola de la Vila. Scale bar: 1 cm. (E) Dufrenoyia dufrenoyi, lateral view of the specimen PUAB 88784. Basin lithofacies association, transgressive unit of Depositional Sequence B, Benassal Formation, Early Aptian, Mola de la Vila. Scale bar: 1 cm. (F) Orbitolinopsis simplex. Platform lithofacies association, highstand normal regressive platform of Depositional Sequence B, Benassal Formation, Late Aptian, Mola de la Garumba. (G) Mesorbitolina texana. Platform lithofacies association, highstand deposits of Depositional Sequence B, Benassal Formation, Late Aptian, Mola de la Garumba.
Fig. 13. Sequence stratigraphy. (A) Photomosaic of the north-eastern face of the Mola de la Garumba. The area outlined in yellow corresponds to the close-up view displayed in (B). Width of image is ca 1.9 km. (B) Close-up view of (A). (C) Sequence stratigraphic interpretation of (B). Note the channel-fill deposits exhibited by the lowstand platform at the Mola de la Garumba. The channel drained towards the north. Note also how the lowstand platform of Depositional Sequence B passes laterally from platform to slope settings, and how these lowstand slopes downlap over a forced regressive wedge of Depositional Sequence A. See Fig. 3 for key.

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upper deposits are characterized by thinner beds (Figs 10 and 15) arranged in parasequences of up to few metres thick, reflecting a loss in accommodation as a consequence of the slowing of the relative sea-level rise to a near standstill.

Fossils identified at species level are Orbitolinopsis simplex (Fig. 12F), Mesorbitolina texana (Fig. 12G), Everticyclammina hedbergi and Charantia cuvilieri. The presence of Mesorbitolina texana and of Orbitolinopsis simplex indicates a Late Aptian age (Schroeder et al., 2010; Cherchi & Schroeder, 2012). Given that the first appearance of Mesorbitolina texana occurs at the base of the highstand platform, the boundary between the Early and the Late Aptian may be located within the transgressive marls, somewhere above the basal level with Dufrenoyia furcata ammonites (Figs 2, 3D and 8).

Forced regression
The subsequent relative sea-level fall subaerially exposed the platform of Depositional Sequence B (Figs 3D, 15 and 16). The resulting subaerial unconformity corresponds to Sequence Boundary 2 (Fig. 17A to C). This surface limits depositional sequences B and C, but also C and A at the Mola d’en Camaràs where the Aptian succession was locally incised as a result of base-level fall (Fig. 7). Consequently, Depositional Sequence B was locally eroded (Figs 3D, 7, 15 and 16). The subaerial unconformity exhibits a stepped morphology (Figs 15 and 16) and locally, truncated external moulds of massive and branching colonial corals (Fig. 17C). The voids left by the dissolution of the aragonite coral skeletons are partially to completely filled with whitish, yellowish and brownish fine-grained material (Fig. 17C). Given that the most basinward part of the highstand platform of Depositional Sequence B is not preserved (Figs 1C, 4, 7, 10, 13A and 15A), forced regressive deposits cannot be recognized.

The incision documented (Figs 15 and 16) is a unique feature and has not been recognized at the other hillocks within the study area (Fig. 1). However, at the Mola de la Saranyana (Figs 1C and 4) a metre-sized sedimentary succession of tabular-bedded completely dolomitized peritidal deposits with very limited lateral extent (few tens of metres) (Fig. 17D) crops out above transgressive basinal marls of Depositional Sequence B. Such proximal lithofacies could be relics of the peritidal infill of an incised valley. However, highstand deposits of Depositional Sequence B are not preserved in this hillock (Fig. 7). Therefore, it is not possible to confirm that the Mola de la Saranyana was incised during the Late Aptian forced regressive stage of Depositional Sequence B.

Depositional Sequence C
Lowstand normal regression
At the lowest point of relative sea-level and with the succeeding relative sea-level rise, available depositional space shifted basinwards. Normal regressive deposits may have accumulated in these distal settings, while the proximal areas of the highstand platform of Depositional Sequence B stayed emerged. However, the most distal parts of this carbonate platform are not preserved (Figs 1C, 4, 7, 10, 13A and 15A).

Transgressive genetic type of deposit
Coinciding with transgression, a maximum regressive surface characterized by a hardground, was superposed onto the subaerial unconformity corresponding to Sequence Boundary 2 and, thus, a composite surface formed (Figs 3D, 9, 15, 16 and 17A to C). Hardground features displayed by the maximum regressive surface are borings of lithophagid bivalves, iron stains and encrusting oysters (Fig. 17B). The incision was backfilled with transgressive deposits (Benassal Formation), which onlap Sequence Boundary 2 (Figs 3D, 9, 15 and 16). Biota identified at genus level in the transgressive rocks are Praeorbitolina sp. and Choffatella sp. The incision cuts the Aptian succession from the highstand carbonates of Depositional Sequence B down to the transgressive marls of Depositional Sequence A (Figs 3D, 9C, 15C and 16B). The uppermost part of the incision backfill (Figs 9, 15, 16 and 17A) is characterized by cross-stratified grainstones with herring-bone cross-bedding (Fig. 6D) and tidal bundles (Fig. 6E). Where these high-energy deposits onlap the transgressive marls of Depositional Sequence B, the subaerial unconformity would have been eroded and the sequence boundary would then correspond to a transgressive ravinement surface. Fluvial or coarse-grained siliciclastic deposits backfilling the incision were not observed.

INCISION CHARACTERIZATION
The seismic-scale erosional incision located at the Mola d’en Camaràs (Figs 1, 4 and 18A) is
Fig. 14. Stratal geometries and drowning of the lowstand normal regressive platform of Depositional Sequence B. (A) Outcrop view of the south-western part of the Mola de la Garumba. Width of image is ca 50 m. (B) Sequence stratigraphic interpretation of (A). Note the prograding geometries displayed by the lowstand platform, and how the in situ platform lithofacies pass basinwards to slope settings. On top of the platform, the maximum regressive surface marks the drowning of the platform, which is buried by transgressive basin marls. (C) Close-up view of three oysters encrusting the maximum regressive surface, which exhibits an iron crust. Mola d’en Camaras. (D) Detail of the transgressive lag present in the Mola de la Garumba. Note how the transgressive lag drapes over an irregular maximum regressive surface. See Fig. 3 for key. Hammer length = 32 cm.

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characterized with a geological map and a cross-section (Fig. 18B). In its present state, the difference in elevation between the lowest and the highest part of the incision is 100 m on the SW slope (950 to 850 m) and 140 m on the NE slope (970 to 830 m) (Figs 18A and B). The cross-section was restored to the pre-contractional geometry (prior to the Alpine orogeny) (Fig. 18C) to define the original shape and dimensions of the incision. The restored cross-section displays a preserved portion of the incision that is ca 2 km wide, with clear down-cutting of 115 m into the Aptian succession. This difference in elevation was obtained by means of a linear interpolation between the heights measured for the two aforementioned slopes. However, if the uppermost transgressive deposits of Depositional Sequence C preserved were deposited within the incision, the depth of the incision would then be 140 m (Fig. 18C). The upper part of the transgressive unit and the original lateral extent of the subaerial unconformity are not preserved due to Neogene and present-day erosion (Figs 18A, 18B and 19). Therefore, the original incision probably would have been larger in both width (>2 km) and depth (>140 m).

Fig. 15. Sequence stratigraphy. (A) Photomosaic of the south-western face of the Mola d’en Camaràs. The area outlined in white corresponds to the close-up view displayed in (B). The area outlined in red is enlarged in Fig. 17A. Width of image is ca 3.3 km. (B) Close-up view of (A). (C) Sequence stratigraphic interpretation of (B). In this image, the incision down-cuts the entire Depositional Sequence B. Note also the stepped shape (strath terraces) displayed by the unconformable surface, which corresponds to Sequence Boundary 2, and how the transgressive strata of Depositional Sequence C onlap this surface. See Fig. 3 for key.
Fig. 16. Sequence stratigraphy. (A) Close-up view of Fig. 9C. North-eastern face of the Mola d’en Camaràs. Cliff is ca 50 m height. (B) Sequence stratigraphic interpretation of (A). In this view, the erosional surface cuts the Aptian deposits down to the lowstand carbonates of Depositional Sequence B. Note how the unconformable surface (Sequence Boundary 2) displays a stepped morphology (strath terraces) and how the transgressive deposits of Depositional Sequence C onlap this surface. See Fig. 3 for key.
The shape and dimensions of the depression indicate that the incision was a valley (Fig. 18C), which had a south-west to north-east trend (Fig. 18). The north-western and south-eastern sides of the palaeovalley exhibit a stepped geometry (Figs 9C, 15B, 16B, 18B, 18C and 19).

**DISCUSSION**

**Genesis of the palaeovalley**

A submarine origin for the seismic-scale incision discovered in the northern Maestrat Basin has been ruled out. Tidal currents are not effective in eroding competent strata such as limestone bedrock (Mitchell *et al.*, 2013). In addition, oceanic streams capable of scouring bedrock are normally generated in the basin seaward from the carbonate shelves (Pinet *et al.*, 1981; Pinet & Popenoe, 1985). Moreover, the channels produced by these oceanic currents do not exhibit stepped sides or peritidal infills (Pinet & Popenoe, 1985), and therefore are not comparable with the Late Aptian valley under study (Figs 6, 15, 18 and 19).

On the other hand, the palaeovalley was incised owing to base-level fall. The stepped
geometry exhibited by its sides indicates that the incision originated under subaerial conditions as a result of alternating periods of stream-cutting erosion and lateral planation (Figs 15, 16, 19C and 19D). The morphology of the palaeovalley, but not the depth, could have been accentuated by wave abrasion during the transgressive stage of Depositional Sequence C. However, it is highly unlikely that the stepped sides correspond to wave-cut benches, given that both sides of the incision show this geometric feature (Figs 15, 16, 19C and 19D).

Sequential development of the sedimentary succession

In this study, four genetic types of deposits separated by key sequence stratigraphic surfaces...
were recognized: forced regressive, lowstand normal regressive, transgressive and highstand normal regressive (Fig. 3). Results provide a valuable example of the four genetic unit-based sequence stratigraphic methods for interpreting marine carbonate successions, which is useful for academic and industrial applications.

However, the application of sequence stratigraphy to the interpretation of incision backfill deposits is often problematic. Within the valley, due to spatial limitation and confinement, the geometric arrangement of the sedimentary fill does not provide any evidence of deposition during a specific stage of relative sea-level (Figs 15, 16, 17A and 19). Examples of transgressive valley-fill successions are numerous in the literature (e.g. Dalrymple et al., 1992; Ardies et al., 2002; Zaitlin et al., 2002; Chaumillon

Fig. 19. Incision characterization. (A) Panoramic view of the south-western face of the Mola d’en Camaràs showing the Late Aptian incised valley documented. This photomosaic corresponds to the outcrop expression of Fig. 18B. See Figs 1C and 4 for location of the Mola d’en Camaràs. Width of image is ca 2·9 km. (B) Interpretation of (A). The area outlined in yellow is enlarged in (C) and (D). (C) Outcrop expression of the area indicated with a segment in the geological cross-section displayed in Fig. 18B. (D) Interpretation of (C). See Fig. 18 for key.
et al., 2008; Raven et al., 2010). However, the possibility that the lowermost and uppermost strata of the incision backfill could also correspond to lowstand and highstand normal regressive deposits, respectively, should not be ruled out. Although the infill succession exhibits the shallowest lithofacies association reported in the study, in order to accumulate ≥115 m thick peritidal to shallow subtidal carbonates (Fig. 18C), rapid creation of accommodation is needed; this is typical of transgressive sequences. The shift from subaerially exposed stream-cut bedrock to peritidal backfill deposits is also indicative of marine onlap. Therefore, the incised valley is interpreted to have been filled during the transgressive stage of Depositional Sequence C (Figs 2, 3, 15 and 16).

In addition, multiple superimposed cycles of incision and sedimentation were not observed within the sedimentary fill. Therefore, following the classification by Zaitlin et al. (1994), the incised valley-fill succession corresponds to a single fill, i.e. a single depositional sequence.

Age control of depositional sequences

Using the absolute ages of Gradstein et al. (2004), the duration of Depositional Sequence A, which contains all of the Early Aptian ammonoid zones (Fig. 2), would be between 3 Myr and 4 Myr. The base of Depositional Sequence B, which corresponds to the basal part of the Benassal Formation, correlates to the uppermost *Dufrenovia furcata* Zone of Early Aptian age (Moreno-Bedmar et al., 2012a; Figs 2, 8 and 12C to E). Therefore, the limit between the Early and Late Aptian would be located within the transgressive marls of Depositional Sequence B (Figs 2, 3 and 8). The upper boundary of this sequence and the sedimentary fill of Depositional Sequence C lack precise biostratigraphic control. However, these deposits cannot be of Albian age because the Escucha Formation, which regionally unconformably overlies the Benassal Formation, contains the first Albian ammonoid Zone (García et al., 2014). The Escucha Formation is not preserved in the study area, but is present overlying the Benassal Formation a few kilometres to the north, in the environs of the village of Herbers (Querol et al., 1992), still within the Morella Sub-basin. In addition, the occurrence of *Toucasia* sp. and *Polyconites* sp. rudist shells in life position at the upper part of the sedimentary backfill succession also rules out a Late Cretaceous or Cenozoic age (Masse, 1995) for these rocks and, hence, the possibility that the formation of the incised valley studied could have been related to the Alpine orogeny (Late Eocene–Early Miocene; Salas et al., 2001). Therefore, given that the Late Aptian spanned 9 Myr (Gradstein et al., 2004), Depositional Sequence B, together with the transgressive phase of Depositional Sequence C, would have had a duration of <9 Myr.

The transgressive backfill deposits of Depositional Sequence C contain *Praeorbitolina* specimens, indicating an Early Aptian age (Schroeder et al., 2010; Cherchi & Schroeder, 2012). However, *Mesorbitolina texana* (Fig. 12G) and *Orbitolinopsis simplex* (Fig. 12F), which are indicative of the Late Aptian (Schroeder et al., 2010; Cherchi & Schroeder, 2012), are found in the incised highstand platform carbonates of Depositional Sequence B. The incised highstand platform of Depositional Sequence B cannot be younger than its backfill. Therefore, the *Praeorbitolina* specimens present in the transgressive peritidal deposits of backfill of Depositional Sequence C were reworked during transgression into the Late Aptian backfill.

Duration and magnitude of the long-term changes in accommodation

The late Early Aptian forced regressive stage recognized occurred within the *Dufrenovia furcata* ammonite biozone (Fig. 2), which according to Gradstein et al. (2004) spanned 1 Myr. The *Dufrenovia furcata* Biozone is contemporaneous with sedimentation of the uppermost transgressive and the highstand deposits of Depositional Sequence A, and to the lowstand and basal transgressive stages of Depositional Sequence B (Fig. 2). Therefore, given the time constraints, this fall of relative sea-level would have had a duration of less than 1 Myr.

The amplitude of relative sea-level change during the late Early Aptian forced regression can also be roughly estimated. The preserved highstand platform carbonates of Depositional Sequence A (Fig. 11A) are ca 80 m thick, whereas the lowstand platform of Depositional Sequence B (Figs 10, 13, 14A, 14B, 15 and 16), which is situated in a basinward position with respect to the highstand platform (Fig. 3), locally overlying forced regressive wedges (Figs 3, 8, 11B and 13), is only ca 10 m thick. Both platforms exhibit the same lithofacies distribution.
(Figs 3, 12A and 12B) and thus, are interpreted to have developed at similar depths. To accommodate a 10 m thick lowstand carbonate succession at the toe of the slope of a highstand platform (Fig. 3), which originally was ≥80 m thick, a relative sea-level drop of at least tens of metres is needed.

The forced regressive stage of Depositional Sequence B is of Late Aptian age (Fig. 2). Other than the occurrence of Dufrenoyia dufrenoiyi specimens within the basal transgressive deposits of this depositional sequence (Figs 8 and 12C to E), no ammonites have been identified and so an approximate estimation of the duration of the relative sea-level fall cannot be made. Depositional Sequence B and the transgressive stage of Depositional Sequence C occurred in <9 Myr. Thus, the Late Aptian forced regressive phase would have had a duration of much less than 9 Myr.

The incised valley characterized permits an accurate determination of the amplitude of this drop in relative sea-level that was at least 115 m, but may have been >140 m (Fig. 18C). Given that the incision was completely backfilled by peritidal to shallow subtidal deposits during the subsequent transgression (Figs 3, 6, 15, 16 and 19), the magnitude of the following relative sea-level rise within Depositional Sequence C would also have been ≥115 m (Fig. 18C).

The depth at which the transgressive marls of Depositional Sequences A and B were deposited is unknown, making it difficult to accurately estimate the amplitude of these transgressions. However, the transgressive deposits of Depositional Sequence A (Forcall Formation) contain the entire Deshayesites forbesi and Deshayesites deshayesi ammonoid biozones, the lower part of the Dufrenoyia furcata Zone and at least the upper part of the Deshayesites oglanensis Zone (Fig. 2; Moreno-Bedmar et al., 2010; Garcia et al., 2014). Therefore, according to the absolute ages of Gradstein et al. (2004), this transgression would have spanned between 2 Myr and 4 Myr.

The basal transgressive unit of Depositional Sequence B (Benassal Formation) recorded the uppermost part of the Dufrenoyia furcata Zone (Figs 2, 8 and 12C to E). The rest of the transgressive marls of Depositional Sequence B were thus most likely to have been deposited during the earliest Late Aptian (Figs 2, 3 and 8). Age-diagnostic ammonites of Late Aptian age were not found in the outcrops studied. Therefore, a more exact estimation of the duration of the transgressive stage of Depositional Sequence B is not possible. However, according to the absolute ages provided in Gradstein et al. (2004), it would have been much less than 9 Myr, given that the Late Aptian substage would have spanned 9 Myr.

Regional significance of the long-term changes in accommodation

The transgressive unit of Depositional Sequence A (Figs 2 and 3), which corresponds to the marls with ammonites of the Forcall Formation (Figs 5C and 10), is observed throughout the Maestrat Basin (Fig. 1A and B) with similar lithofacies and a thickness ranging from tens to hundreds of metres (Salas, 1987; Salas et al., 2001; Vennin & Aurell, 2001; Moreno-Bedmar et al., 2009, 2010; Bover-Arnal et al., 2010, 2011a,b; Embry et al., 2010; Garcia et al., 2014). Thickness variations are related to differential accommodation due to pronounced compartmentalization into fault blocks of the basin during the Early Aptian (Salas et al., 2001; Bover-Arnal et al., 2010; Moreno-Bedmar et al., 2010; Martin-Martín et al., 2013).

The late Early Aptian (intra Dufrenoyia furcata Zone) forced regression of Depositional Sequence A (Figs 2 and 3) has been also recognized in the central Galve Sub-basin in the western Maestrat Basin (Fig. 1B; Bover-Arnal et al., 2009, 2010, 2011a). There, as in the Morella Sub-basin, the occurrence of platform-detached wedges deposited above the basal marls of the Forcall Formation, and topographically below a highstand platform, provides evidence of base-level fall during the late Early Aptian (Bover-Arnal et al., 2009, 2011a). In the central Galve Sub-basin, the rate of this relative sea-level drop was at least 60 m in <1 Myr (Bover-Arnal et al., 2009). Further, palaeokarst features located at the top of the preserved highstand carbonates of Depositional Sequence A and a kilometre-wide late Early Aptian incision down-cutting the sedimentary succession to tens of metres deep (Peropadre et al., 2007; Bover-Arnal et al., 2009, 2010, 2011a) are preserved in the Galve Sub-basin. This erosional surface also displays strath terraces (fig. 10 in Bover-Arnal et al., 2011a), as observed in this study.

The transgressive stage of Depositional Sequence B (Fig. 3), which was coeval with deposition of the marly unit of the lower part of the Benassal Formation (Figs 2, 8, 14A and 14B), can also be traced throughout the Maestrat
Global significance of the long-term changes in accommodation

The global Aptian sedimentary record is calibrated with biostratigraphic data (e.g. Moreno-Bedmar et al., 2009, 2010, 2012a,b; Schroeder et al., 2010; Cherchi & Schroeder, 2012) and/or analyses of stable isotopes such as C or Sr (e.g. Burla et al., 2008, 2009; Huck et al., 2011, 2012). However, the duration of changes in depositional trends recognized in the stratigraphic record is less than the resolution of these calibration methods, which can have errors of up to millions of years. Therefore, relative ages that are used to calibrate Aptian strata cannot be used as absolute age correlations between stratigraphic units and/or changes in accommodation recorded in different basins. Consequently, the synchronicity of Aptian sedimentary events is always difficult to test. Despite these limitations, when reviewing the literature, striking similarities exist between the changes in depositional trends, lithofacies evolution and geometric relation of sedimentary successions characterized in this study and those reported from rocks of the same age from different basins worldwide.

In this regard, deposition of the Early Aptian transgressive marls of Depositional Sequence A (Forcall Formation; Figs 2, 3, 10B and 10C) coincide with a large-scale transgression that has been recorded along the margin of the Tethys (e.g. Föllmi et al., 1994; Sahagian et al., 1996; Hardenbol et al., 1998; Wissler et al., 2003; Husinec & Jelaska, 2006). This Early Aptian transgression was accompanied by mass occurrences of orbitolinids in the Tethyan neritic realm (Villas et al., 1995; Pittet et al., 2002; Burla et al., 2008; Bover-Arnal et al., 2010; Embry et al., 2010; Stein et al., 2012).

The late Early Aptian fall in relative sea-level of Depositional Sequence A (Figs 2 and 3) also has potential counterparts elsewhere in the Tethys (Haq et al., 1988; Hardenbol et al., 1998). Arnaud & Arnaud-Vanneau (1989) reported the existence of a late Early Aptian subaerial unconformity associated with incised valleys several hundred metres wide, which down-cut platform carbonates in south-eastern France. Subaerial exposure surfaces above platform carbonates of late Early Aptian age have been also recognized in the Lusitanian Basin in Portugal (Burla et al., 2008), in the Basque-Cantabrian Basin in northern Spain (Rosales, 1999; García-Mondéjar et al., 2009; Fernández-Mendiola et al., 2013), in the Organyà Basin in the Catalan Pre-Pyrenees (Bernaus et al., 2003), in the Adriatic platform in Croatia (Husinec & Jelaska, 2006), in Oman (van Buchem et al., 2010; Rameil et al., 2012), in onshore Abu Dhabi in the United Arab Emirates (Yose et al., 2010), in the Levant Basin in southern Lebanon and northern Egypt (Bachmann & Hirsch, 2006), in the Apennines in Italy (Ruberti et al., 2013) and in the western Taurides in Turkey (Yilmaz & Altiner, 2006). In addition, late Early Aptian regressive deposits have been
identified in the north-western Pacific guyots (Röhl & Ogg, 1998) and in south-eastern Ethiopia (Bosellini et al., 1999).

Transgressive deposits correlative with the transgressive phase of Depositional Sequence B (Figs 2 and 3), have also been reported in other basins of the Tethys and the Atlantic extension of it (Haq et al., 1988; Hardenbol et al., 1998). Latest Early Aptian drowning of carbonate platforms occurred in Mexico (Moreno-Bedmar et al., 2012a), Venezuela (Jacquin et al., 1993) and south-eastern France (Masse & Fenerci-Masse, 2013). Wilmsen et al. (2013) also placed the time of drowning of the Shah Kuh Platform in Central Iran near the boundary between the Early and the Late Aptian. Similarly, Röhl & Ogg (1998) identified a transgressive phase in the western Pacific guyots that began in the uppermost Early Aptian.

Evidence of a pronounced Late Aptian relative sea-level drop potentially coeval to the forced regressive stage of Depositional Sequence B (Figs 2 and 3), is also widespread (Haq et al., 1988; Hardenbol et al., 1998). Deposits equivalent to this regressive stage are the Upper Aptian lowstand clinoforms on the Arabian Plate (Morrison et al., 1997; Maurer et al., 2010; Pierson et al., 2010); these are interpreted as the lowstand of the Late Aptian to early Albian Supersequence (van Buchem et al., 2010). Incised valleys of that age have been described from the Arabian Plate (Raven et al., 2010; Maurer et al., 2013), where incisions are up to 65 m deep and 8 km wide and infilled by estuarine deposits; western Siberia (Medvedev et al., 2011), where incised valleys are up to 90 m deep and also infilled with estuarine materials; and in western Canada (Zaitlin et al., 2002), where incisions of up to 40 m deep are infilled with siliciclastic tidalites. The transgressive infill of Depositional Sequence C is of latest Aptian age, such as the siliciclastic backfills observed in Qatar (Raven et al., 2010) and western Siberia (Medvedev et al., 2011). In addition, Late Aptian sea-level falls of up to tens of metres in magnitude have been recognized in Venezuela (Jacquin et al., 1993), the western Pacific guyots (Röhl & Ogg, 1998), the Adriatic Platform in Croatia (Husinec & Jelaska, 2006), the North Sea (Crittenden et al., 1997), Yemen (Morrison et al., 1997) and the Russian Platform (Sahagian et al., 1996; Zorian & Ruban, 2012). Palaeogeographic maps showing the location of sites with evidence of significant relative sea-level falls of Aptian age are found in Rameil et al. (2012) and Maurer et al. (2013).

In summary, long-term changes in relative sea-level that are observed in the northern Maestrad Basin have also been described in numerous Aptian successions worldwide. Similar sedimentary expressions of these relative sea-level changes, such as a comparable lithofacies evolution, drowning of carbonate platforms during the latest Early Aptian or the occurrence of deeply incised valleys during the Late Aptian, can be recognized in distinct tectonic plates. Therefore, all these different examples seem to indicate that a global mechanism is responsible for controlling base level during the Aptian in the Morella Sub-basin.

On the triggering mechanism of the long-term changes in accommodation

Long-term relative sea-level fluctuations (0.5 to 6 Myr) such as the ones recognized in the northern Maestrad Basin are commonly interpreted to have a tectonic and/or climatic origin (e.g. Vail et al., 1977, 1991; Cloetingh, 1986, 1991; Bouilla et al., 2011). The main mechanisms for changes in accommodation are: glacio-eustasy, tectono-eustasy and thermo-eustasy (global); and sediment supply, uplift, and thermal and tectonic subsidence (regional) (e.g. Cronin, 1999; Immenhauser, 2005). These mechanisms can act in concert or in isolation. Consequently, reconstructing the relative contribution of each of these factors is a challenge.

Falls in relative sea-level of tens of metres in <1 Myr or of ≥115 m in much less than 9 Myr, such as those recognized in the northern Maestad Basin (Figs 2 and 3), are considered ‘rapid’ in geological terms (e.g. Cronin, 1999; Immenhauser & Matthews, 2004; Gréselle & Pittet, 2005; Miller et al., 2005a). Along the same lines, the incised valley documented was completely filled within the Aptian Stage. Therefore, the rate of the latest Late Aptian transgressive phase of Depositional Sequence C (Figs 2, 3, 15, 16, 18 and 19) would have been similar to one of the Late Aptian forced regressions of Depositional Sequence B. According to Immenhauser (2005), the only known mechanisms of sea-level fluctuations that would produce rates of change matching the above are regional tectonism, glacio-eustasy and ocean-volume changes.

Changes in the volume of the ocean basins are commonly related to volcanic activity at the mid-oceanic ridges and to subduction of one continent beneath another (e.g. Kominz, 2001). However, neither intense magmatic activity at
oceanic ridges, nor subduction processes can explain the rates of sea-level change documented. Thermal anomalies causing rapid spreading of oceanic ridges would be expected to trigger rapid transgressions and much slower forced regressive phases (Immenhauser, 2005), which is the opposite phenomenon to that observed in the northern Maestrat Basin (Fig. 2). Rates of sea-level change due to plate subduction differ by at least one order of magnitude from those estimated in the Morella Sub-basin and so cannot account for the sea-level falls recognized. As an example, the subduction of the Indian Plate under the Eurasian Plate has generated a sea-level drop of 70 m during the last 50 Myr (Kominz, 2001).

During the Aptian, the northern Maestrat Basin was bounded by a major normal fault with a west–east trend (Herbers fault). Due to hangingwall accommodation, the depocentre of the Morella Sub-basin, and thus the basinward direction of the depositional system studied, was towards the north-west and north-east (Fig. 1B). This trend is demonstrated in the platform top to slope clinoforms that prograde NW–NE in the lowstand and highstand platforms of Depositional Sequence B (Figs 13B, 13C, 14A and 14B); this supports fault-induced subsidence as a factor in creating accommodation.

The southern area of the Morella Sub-basin, where the carbonate succession studied crops out (Fig. 1B), corresponds to the upthrown part of the tilted fault block and, thus, to proximal settings of the depositional system. Therefore, tectonic uplift linked to the tilting of the fault block could have accounted, at least partially, for the long-term late Early and Late Aptian forced regressions identified.

However, the nature, amplitude and duration of the relative sea-level falls documented cannot be explained solely as the result of regional tectonic processes. Coeval drops of tens of metres in relative sea-level have been recognized worldwide (e.g. Hardenbol et al., 1998; Bover-Arnal et al., 2009; Maurer et al., 2010, 2013; Husinec et al., 2012; Rameil et al., 2012), so there must be a eustatic component. In addition, Immenhauser (2005) estimated rates of ca 0.65 to 0.88 m Myr\(^{-1}\) for Jurassic and Cretaceous sea-level changes associated with regional to plate-scale tectonism. These estimated rates are considerably lower than the changes of tens of metres in <1 Myr or of ≥115 m in <9 Myr observed. Therefore, the most plausible mechanism capable of producing such rapid fluctua-

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Aptian relative sea-level history of the north Maestrat Basin

1587
abundances of *Araucaria* pollen is interpreted to reflect a shift towards cooler and probably less arid climatic conditions. This trend has been recognized within the marls of the Forcall Formation and is most pronounced in post-Early Aptian Oceanic Anoxic Event strata (Bover-Arnal et al., 2011b). In summary, the Aptian sedimentary succession of the Maestrat Basin provides evidence of late Early and Late Aptian cooling trends that precede or are coeval with stratigraphic relations and sedimentary features that resulted from base-level fall.

**CONCLUSIONS**

Although the amplitude and duration of the two relative sea-level falls reported are within the glacio-eustatic domain (respectively, of tens of metres in <1 Myr and of ≥115 m in <9 Myr), results do not unambiguously prove that the growth and melting of large continental ice-sheets were the main processes controlling long-term accommodation changes in the northern Maestra Basin during the Aptian. However, the relative sea-level trends recognized, and their sediment expression (for example, the presence of a Late Aptian incised valley with a peritidal to shallow subtidal backfill), are comparable to those documented in coeval sedimentary successions from other basins worldwide, and thus have an important eustatic component. If these eustatic variations in sea-level are taken in context with the existing bibliography of sedimentological, mineralogical, palaeontological and geochemical studies reporting the occurrence of cooling events in the Aptian, glacio-eustasy becomes a plausible mechanism for explaining the sea-level changes identified.

Finally, the wider importance of the study lies in the documentation and interpretation of the outcrops analysed in the northern Maestra Basin. These rock exposures allow investigation of the stratigraphic architecture and lithofacies relations of global long-term relative sea-level fluctuations of Aptian age at the outcrop scale, where they are often only recognizable at much larger (seismic) scales.

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