Facies model of a mixed clastic–carbonate, wave-dominated open-coast tidal flat
(Tithonian–Berriasian, north-east Spain)

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ABSTRACT

Detailed logging and analysis of the facies architecture of the upper Tithonian to middle
Berriasian Aguilar del Alfambra Formation (Galve sub-basin, north-east Spain) have made it
possible to characterize a wide variety of clastic, mixed clastic–carbonate and carbonate
facies, which were deposited in coastal mudflats to shallow subtidal areas of an open-coast
tidal flat. The sedimentary model proposed improves what is known about mixed coastal
systems, both concerning facies and sedimentary processes. This sedimentary system was
located in an embayed, non-protected area of a wide C-shaped coast that was seasonally
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dominated by wave storms. Clastic and mixed clastic–carbonate muds accumulated in poorly-drained to well-drained, marine-influenced coastal mudflat areas, with local fluvial sandstones (tide-influenced fluvial channels and sheet-flood deposits) and conglomerate tsunami deposits. Carbonate-dominated tidal flat areas were the loci of deposition of fenestral-laminated carbonate muds and grainy (peloidal) sediments with hummocky cross-stratification. Laterally, the tidal flat was clastic-dominated and characterized by heterolithic sediments with hummocky cross-stratification and local tidal sandy bars. Peloidal and heterolithic sediments with hummocky cross-stratification are the key facies for interpreting the wave (storm) dominance in the tidal flat. Subsidence and high rates of sedimentation controlled the rapid burial of the storm features and thus preserved them from reworking by fair-weather waves and tides.

**Keywords:** Galve sub-basin, mixed clastic–carbonate, open-coast tidal flat, storm deposits, Tithonian–Berriasian.

**INTRODUCTION**

The zone where land and sea meet is composed of a variety of complex environments that are subjected to a broad range of sedimentary conditions, including fluvial, wave and tidal influence. As a consequence, in order to interpret coastal environments in the sedimentary record, a thorough understanding of modern counterparts is required (Davis, 1978). For terrigenous clastic coasts, the classic ternary process-based classification scheme of Boyd et al. (1992), comprising deltas (fluvial sediment sources), estuaries and lagoons (mixed fluvial–marine sediment sources), and beaches and tidal flats (marine sediment sources) has been modified in its 'wave-tide coordinate' by Yang et al. (2005) to include wave-influenced and tide-influenced open-coast tidal flats (Fig. 1A). In a broad sense, hybrid wave-influenced and tide-influenced coasts include a wide spectrum of coastal areas,
ranging from moderate wave-energy micro-tidal to meso-tidal sand barriers with frequent tidal inlets, to macro-tidal and mega-tidal beaches (e.g. Anthony & Orford, 2002). A review of the literature on modern examples of clastic coastal environments is offered in Holland & Elmore (2008), including in the ‘wave–tide coordinate’, the classic model of the tide-dominated (sheltered) tidal flat of the Wadden Sea (e.g. Dalrymple, 2010), the open-coast tidal flat seasonally dominated by wave storms of the Yellow Sea coast of Korea (Yang et al., 2005), and many other examples of wave-dominated beaches and tidal beaches (see also Tamura, 2012) (Fig. 1A). Examples in the sedimentary record of terrigenous clastic wave-dominated coasts and tide-dominated tidal flats are frequent (e.g. Longhitano et al., 2012; Vakarelov et al., 2012) but those of wave-influenced and tide-influenced open-coast tidal flats are very scarce (e.g. Basilici et al., 2012).

If a similar wave–tide binary process-based classification is considered for carbonate and mixed clastic–carbonate coastal areas, both modern and ancient examples are included in wave-dominated coasts or in tidal flats (Fig. 1B). For example, modern wave-dominated carbonate coasts located in Australia (e.g. James et al., 2004; Short, 2006) and the Bahamas (e.g. Lloyd et al., 1987; Rankey, 2014) have well-known fossil counterparts (e.g. Kerans & Loucks, 2002; Martin et al., 2004). However, mixed clastic–carbonate wave-dominated coasts are scarcer, and both modern examples and fossil analogues indicate that carbonate sediments tend to concentrate in subtidal areas (e.g. Testa & Bosence, 1998; Gläser & Betzler, 2002; Mata & Bottjer, 2011). Regarding tidal flats, there are good examples of both carbonate tidal flats (for example, modern: Andros Island, Rankey, 2002; ancient: Goldhammer et al., 1990) and mixed clastic–carbonate tidal flats (for example, modern: the barrier island lagoon-tidal system of Abu Dhabi, Alsharhan & Kendall, 2003; ancient: Breda & Preto, 2011; Quijada et al., 2016). However, there are no well-known modern or fossil carbonate and mixed clastic–carbonate coastal areas potentially equivalent to clastic open-coast tidal flats. In addition, for mixed clastic–carbonate successions there is an open debate on whether ancient examples represent ‘true’ mixed clastic–carbonate...
systems or whether, on the contrary the vertical arrangements of clastic and carbonate facies represent episodes of reciprocal sedimentation controlled by external factors (e.g. Schwarz et al., 2016).

The principal aim of this paper is to present a sedimentary model for a mixed terrigenous clastic–carbonate, wave-influenced and tide-influenced coastal environment that developed during the late Tithonian to middle Berriasian in the Galve sub-basin (Maestrazgo Basin, north-east Spain) and can be taken to represent the first fossil example of a ‘true’ mixed coastal area potentially equivalent to the clastic open-coast tidal flat. This example thus fills the gap in sedimentary knowledge of mixed systems, concerning the nature of both the sediments (clastics and carbonates) and the processes (wave and tidal influences) (Fig. 1B). Analysis of the facies at different scales of observation (from bed and outcrop to log correlation scales) in the 240 to 350 m thick successions under study, allows an accurate interpretation of the lateral and vertical relationship of facies and related subenvironments. The integration of the proposed facies model at a broader (basin) scale, the role of internal and external factors controlling sedimentation, and a comparison with modern examples will be discussed.

GEOLOGICAL SETTING

The Maestrazgo Basin (north-east Spain) forms part of the Iberian Rift System and developed during a rifting phase that commenced at the end of the Jurassic and lasted until mid-Albian times (Salas & Casas, 1993; Salas et al., 2001) (Fig. 2A). This basin was divided into several sub-basins bounded by areas with concentrations of major extensional faults, including the Galve sub-basin studied in the present work. In this area, the Cenozoic (Alpine) compression phase produced two orthogonal folding and thrusting structural trends striking around NNW–SSE and WSW–ENE (Fig. 2B) due to the rejuvenation and inversion of normal faults, essentially inherited from the Late Jurassic to Early Cretaceous rifting stage (Liesa et al., 2006).
The Mesozoic stratigraphy of the Galve sub-basin, summarized in Fig. 2C, comprises pre-rift, syn-rift and post-rift successions. The pre-rift succession consists of Upper Triassic mudstones with gypsum (Keuper facies) followed by 400 to 600 m thick Jurassic platform carbonates (e.g. Aurell et al., 2003). The syn-rift series can be divided into two sequences (Liesa et al., 2017): (i) Kimmeridgian to Berriasian syn-rift sequence 1, represented by up to 700 m thick transitional and continental facies (the Villar del Arzobispo, Aguilar del Alfambra and Galve formations); and (ii) Valanginian to mid Albian syn-rift sequence 2 (more than 1000 m thick in the depocentral areas), which includes a predominantly continental to transitional succession (Wealden facies: Salas et al., 2001; Meléndez et al., 2009) and Aptian shallow platform carbonates (Urgonian facies: Vennin & Aurell, 2001). The post-rift sequence encompasses Albian continental sandstones (Utrillas Formation) and Upper Cretaceous shallow marine carbonates.

The Aguilar del Alfambra Formation analysed in this study belongs to syn-rift sequence 1. This unit is of very variable thickness across the Galve sub-basin (0 to 450 m), indicating strong control of the sedimentation by extensional tectonics (Fig. 2C). The unit rests on an erosive unconformity above the Villar del Arzobispo Formation, which has been related to a major tectonic extensional reactivation event occurring around the late Tithonian (Aurell et al., 2016). This age assignment is based on the presence of the larger benthic foraminifer Anchispirocyclina lusitanica in both the Villar del Arzobispo Formation and in some skeletal levels located in the middle and upper parts of the Aguilar del Alfambra Formation and coeval marine units (e.g. Díaz-Molina & Yébenes, 1987; Bádenas et al., 2004). The stratigraphic distribution of this foraminifera ranges from the middle Tithonian to the earliest Berriasian. In addition, the uppermost levels of the Aguilar del Alfambra Formation contain the middle Berriasian charophyte Globator maillardii incrassatus (Canudo et al., 2012). The upper boundary of the unit is also a major erosive unconformity, which has been dated as having developed towards the end of the middle Berriasian (Aurell et al., 2016). Accordingly, the Aguilar del Alfambra Formation was most probably deposited during the late Tithonian to
middle Berriasian timespan. The unit includes a wide variety of lithologies, mainly whitish limestones, red mudstones and sandstones, which represent deposition in transitional environments ranging from muddy coastal plains to lagoons (Aurell et al., 2016). Of particular interest are the laminated limestones with abundant fenestral porosity and the peloid-rich limestones that commonly preserve dinosaur tracks (Alcalá et al., 2014; Aurell et al., 2016).

DATA SET AND METHODS

The Aguilar del Alfambra Formation was studied in three outcrops near the villages of Aguilar del Alfambra, Ababuj and Allepuz (Los Cerezos), which correspond to the central part of the Galve sub-basin (Fig. 2A and B). The thickness of the unit ranges from 240 m in Aguilar (the uppermost 25 m of the unit are covered) to 350 m in Ababuj, and more than 370 m in Los Cerezos (the lower boundary and lowermost levels of the unit are not exposed). The facies analysis performed in these successions encompassed two main data sets:

1 A detailed logging of three stratigraphic sections was carried out to characterize the features of the sedimentary facies (bed thickness and geometry, lithology, texture, components and sedimentary structures). The field data were complemented with the petrographic study of 55 thin sections of the main facies and the determination of the carbonate content in 40 samples of mudstones and marls by means of a calcimeter that measures the CO2 given off during the reaction of the rock sample with dilute hydrochloric acid.

2 The analysis of the vertical and lateral extension of the facies bodies by physical tracing in the field and on aerial photographs of reference levels (beds or packages of beds of hard lithologies). The aerial photographs included satellite imagery (using the Apple Maps application, with the highest focus at the 1:600 scale allowing observation of metre thick packages) and images taken with a drone flying at low altitude that permitted observation at
a detailed scale (decimetre to metre thick beds or packages). The strata in Aguilar and Ababuj have an almost vertical tectonic dip. Here, the aerial photographs allowed a cross-sectional view of the reference levels along a maximum outcrop length of 500 m (Aguilar) and 900 m (Ababuj). In Los Cerezos, the studied unit has a 60° tectonic dip and crop outs in the cliff faces of a gully; the maximum lateral extension of the outcropping beds is 150 m.

The facies data obtained from the analysis at sample, bed and outcrop scales and the facies architecture reconstructed after correlation of the three stratigraphic reference sections provided the basic arguments for characterizing the dimensions and lateral relationship of the facies and related subenvironments within the sedimentary model proposed.

**FACIES ARCHITECTURE**

**Facies architecture at outcrop scale**

Three groups of facies were differentiated in the Aguilar del Alfambra Formation (Figs 3 to 6): (i) terrigenous clastic facies (T facies), including red mudstones, heterolithic (mudstones-sandstone) alternations, tabular sandstones, cross-bedded sandstones and tabular conglomerates; (ii) mixed terrigenous–carbonate facies (M facies) that encompass grey mudstones-marly limestones and bioturbated silty limestones; and (iii) carbonate facies (C facies), including fenestral-laminated lime mudstones, peloidal packstones–grainstones, bioturbated bioclastic wackestones–packstones and oolitic packstones–grainstones. Type T, M and C facies are vertically stacked, forming eight T/C sequences, named Ag1 to Ag8 in Aguilar (Figs 3 and 4), Ab1 to Ab8 in Ababuj (Fig. 5) and Ce1 to Ce8 in Los Cerezos (Fig. 6). These sequences have a lower interval dominated by T facies passing upwards to dominant C facies; M facies intercalate in both T-dominated and C-dominated intervals. Sequence boundaries are continuous at outcrop scale and correspond to a sharp change from C to T facies, and locally to erosion surfaces.

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Sequences Ag1 to Ag3, Ab1 to Ab3 and Ce1 to Ce3 range from 40 to 95 m in thickness. The T intervals are formed by red mudstones (up to 20 m thick) and intercalations of decimetre to metre thick beds of grey mudstones–marly limestones, decimetre thick tabular sandstones and conglomerates, and lenticular packages of cross-bedded sandstones (up to 5 m thick and 20 to 75 m wide; for example, reference levels Ag1-1, Ag1-3 and Ag2-2 in Fig. 4C, D and G; levels Ab3-1 and Ab3-2 in Fig. 5; level Ce2-1 in Fig. 6). Locally, continuous decimetre to metre thick packages of peloidal packstones–grainstones and fenestral-laminated lime mudstones intercalate in both red and grey mudstones–marly limestones (for example, level Ag2-1 in Fig. 3; level Ab3-3 in Fig. 5). The C intervals are mostly formed by laterally continuous decimetre to metre thick packages of interbedded fenestral-laminated and peloidal facies that intercalate in red mudstones and M facies (Figs 3, 5 and 6). Discontinuous lenticular packages of fenestral-laminated lime mudstones were recognized locally (50 to 200 m wide; for example, levels Ag1-2 and Ag1-4 in Fig. 4C and D; level Ab1-3 in Fig. 5B and C). The Ag1–Ag2 boundary is locally erosive, the erosion cutting down into the uppermost fenestral-laminated lime mudstone level Ag1-8 (Fig. 4B and G).

Sequences Ag4, Ab4 and Ce4 are less than 20 m thick. The T intervals include red and grey mudstones–marly limestones, tabular sandstones, decimetre thick packages of heterolithic facies and lenticular packages of cross-bedded sandstones (up to 30 m wide; for example, level Ag4-1 in Fig. 3; level Ce4-1 in Fig. 6). The C-dominated intervals mainly consist of peloidal facies (for example, level Ab4-2 in Fig. 5; level Ce4-2 in Fig. 6D). In Aguilar, the Ag3–Ag4 boundary is an erosion surface that locally descends into the peloidal level Ag3-4. The cross-bedded sandstones of level Ag4-1 are fitted in this surface (Fig. 4B, E and H). The Ag4–Ag5 boundary is also an erosion surface that partly affects the heterolithic facies and tabular sandstones of level Ag4-3 (Fig. 4E).

Sequences Ag5 to Ag8, Ab5 to Ab8 and Ce5 to Ce8 are 20 to 40 m in thickness and include the first record of bioclastic and oolitic facies. In Aguilar and Ababuj, these sequences have well-differentiated T and C intervals. The T intervals encompass red and
grey mudstones–marly limestones, heterolithic facies and tabular sandstones, in which lenticular packages of cross-bedded sandstones are intercalated (5 m thick and 20 to 50 m wide; for example, levels Ag5-1, Ag6-1 and Ag6-2 in Fig. 4E, F and H; level Ab5-2 in Fig. 5). The C intervals include thin continuous packages of fenestral-laminated lime mudstones and peloidal facies, and also discrete beds of bioturbated bioclastic wackestones–packstones (for example, level Ag5-2 in Fig. 3; level Ab5-1 in Fig. 5). These C facies are continuous, except the laminated-peloidal level Ag7-1, which pinches out into grey mudstones–marly limestones (Fig. 4F). In Los Cerezos, sequences Ce5 to Ce8 have a more complex intercalation of T, M and C facies with an increase in C facies towards the top (Fig. 6). Two particular features in this outcrop are the predominance of heterolithic and mixed facies (for example, sequences Ce5 and Ce6), and the presence of oolitic packstones–grainstones (for example, level Ce7-5 in Fig. 6F). As in Aguilar and Ababuj, bioclastic facies are present in sequence Ce5 (for example, level Ce5-2).

**Facies architecture at the kilometre scale**

Correlation of the three stratigraphic reference sections provides complementary data for refining the vertical and lateral relationship of the facies of the Aguilar del Alfambra Formation along a ca 10 km long, NNW–SSE-oriented transect (Fig. 7). A one-to-one correlation of the eight defined T/C sequences is assumed on the basis of observations in the individual outcrops, in particular the lateral continuity of the boundaries of the T/C sequences. The similar set of facies within the equivalent sequences and the presence of some correlatable beds lend additional support to this best-fit correlation.

In sequences 1 to 3 red mudstones, M facies and peloidal and fenestral-laminated facies are arranged in well-defined T-dominated and C-dominated intervals. The T-dominated intervals are characterized by thick red mudstones that intercalate discrete levels of tabular sandstones and conglomerates and lenticular packages of cross-bedded sandstones. These
cross-bedded sandstones are discontinuous at the kilometre scale and are mainly concentrated in Aguilar and Ababuj. In these sections, a correlatable peloidal packstone–grainstone level is recognized in sequence 2 (level Ag2-1/Ab2-2). In the C-dominated intervals, mainly continuous and some discontinuous (up to 200 m in lateral extension) packages of interbedded peloidal packstones–grainstones and fenestral-laminated lime mudstones are recorded at outcrop scale. Some of the packages are correlatable levels along the entire transect, but others are thought to be discontinuous at the kilometre scale, changing laterally to red mudstones and M facies. Sequences 4 to 8 have significant lateral thickness variation, with increasing thickness towards Los Cerezos and reduced thickness in Aguilar, partly due to erosional character of the sequence boundaries in this area (Fig. 7).

The sequences show significant lateral heterogeneity of facies. Heterolithic (mudstone-sandstone) alternations and M facies (grey mudstones–marly limestones and bioturbated silty limestones) dominate in Aguilar and Los Cerezos, and change laterally to red mudstones in Ababuj (for example, sequence 6). The rest of the facies are also discontinuous at the kilometre scale. Lenticular packages of cross-bedded sandstones are concentrated in Aguilar and Ababuj, intercalated both in red mudstones and M facies. Most of the peloidal, fenestral-laminated and oolitic levels are discontinuous; however, the bioclastic level in sequence 5 (level Ag5-2/Ab5-1/Ce5-2) and some peloidal levels in the uppermost part of the sequences are thought to be correlatable along the entire transect.

In summary, the reconstruction of the facies architecture allows the following lateral facies relationships to be deciphered: (i) the fine-grained T facies (red mudstones), M facies and the heterolithic alternations are laterally related (for example, sequence 6); (ii) the tabular conglomerates and sandstones and the lenticular, discontinuous cross-bedded sandstones are mostly interbedded within red mudstones and M facies; (iii) the interbedded fenestral-laminated lime mudstones and peloidal packstones–grainstones form either discontinuous or continuous packages that intercalate in red mudstones and M facies; and (iv) the bioturbated bioclastic wackestones–packstones and the oolitic packstones–grainstones form thin
intercalations within fenestral-laminated, peloidal, heterolithic and M facies. Bioclastic facies are locally continuous at the kilometre scale (see the bioclastic level in sequence 5).

FACIES DESCRIPTION AND INTERPRETATION

Red mudstones and mixed facies: coastal mudflat deposits

The red mudstones and M facies (grey mudstones–marly limestones and bioturbated silty limestones) correspond to intercalated terrigenous and mixed terrigenous-carbonate muds (Fig. 7). The CaCO$_3$ content of the red mudstones ranges from 1 to 15% (mean value = 6%). The grey lutite–marly limestones show variable CaCO$_3$ content, ranging from 5 to 34% (mean value = 22%) in the grey mudstones. Both red mudstones and grey mudstones–marly limestones are massive or parallel-laminated and have local root traces, desiccation cracks, dinosaur tracks and carbonaceous levels enriched in plant debris (Fig. 8A to D). The bioturbated silty limestones have scarce skeletal grains (mainly bivalves, including oysters and small gastropods) and show root traces and burrows including Arenicollites and Thalassinoides and local desiccation cracks and dinosaur tracks preserved as concave epireliefs (Fig. 8E and F).

Subaerial exposure structures, fossil traces and skeletal grains reflecting marine influence indicate sedimentation in a low-lying coastal mudflat. The red mudstones were probably deposited in well-drained emergent and locally vegetated areas (e.g. Breda & Preto, 2011). The grey mudstones–marly limestones and the bioturbated silty limestones represent deposition in poorly-drained areas with a certain marine influence in the coastal mudflat, either as discontinuous ponds (see the discontinuous M facies in sequences 1 to 3 in Fig. 7) or as wider areas lateral to the red lutites of the emergent coastal mudflat (see sequences 4 to 8 in Fig. 7).
Tabular sandstones and conglomerates: sheet-flood and tsunami deposits

Isolated decimetre thick tabular sandstone and conglomerate beds hundreds of metres in lateral extent intercalate within the red mudstones and M facies of the coastal mudflat (see Fig. 7). The sandstones consist of fine and medium quartz grains with locally abundant mica, pebble-sized clasts and trunk and bivalve fragments. They are massive to parallel-laminated, with sporadic cross-lamination, convolute lamination and burrows (Taenidium and Arenicolites; Fig. 9A and B). The tabular (sheet-like) shape of the beds is comparable with deposits from non-confined fluvial sheet floods or flash floods deposited in coastal flood plains (e.g. Nemec & Steel, 1984; Hwang et al., 1995; Breda & Preto, 2011).

The conglomerates are either matrix-supported or grain-supported, and have sandstone matrix and clasts that are very variable in roundness (angular to sub-rounded), size (up to several centimetres) and lithology (red mudstone, sandstone and limestone, mostly fenestral-laminated lime mudstone) (Fig. 9C to H). Sporadic trunk fragments and bivalve fragments (including oysters) are also present. The variable roundness and lithology of the clasts indicate that they were eroded and deposited very close to their source area, i.e. they are rip-up clasts of coastal sediments, probably the products of tsunami or storm flows rather than non-channelized fluvial flows. In particular, a tsunami origin is proposed on the basis of the vertical succession of sedimentary structures within two main types of conglomerate deposits:

1. Conglomerate to sandstone bed formed by grain-supported, normal-graded conglomerate sublayers bounded by scoured surfaces, with a parallel-laminated to wave-laminated sandstone sublayer on top (Fig. 9C to E). This deposit fits the conglomerate-type tsunami deposits described by Fujiwara & Kamataki (2007), because it includes sublayers with scoured and graded structures, each of which indicates deposition from waning sediment flows, with the thick and coarser-grained sublayer resulting from the stronger wave (Fig. 9E). However, the absence of clear palaeocurrent data (i.e. from imbrication) in the
studied conglomerates precludes confirmation of the action of upflow and return flows. Beds formed by a massive or crudely normal-graded grain-supported conglomerate sublayer capped by a laminated sandstone sublayer are also recognized (Fig. 9D). These probably correspond to ‘incomplete’ conglomerate-type tsunami deposits due to lateral variations in tsunami flows.

2 Conglomerate to sandstone bed formed by a massive, matrix-supported conglomerate sublayer, passing upward to sandstone sublayers with rip-up clasts and convolute lamination, and hummocky cross-stratification (HCS) (Fig. 9F and G). The coeval presence of these structures, and in particular of rip-up clasts, internal erosion (scoured) surfaces and syndepositional deformation, is more typical of tsunami deposits than of coastal storm deposits (e.g. Morton et al., 2007). The described deposit has some similarities (rip-up clasts, truncation surfaces, lamination and HCS) with the sandy-type and gravelly-type tsunami deposits of Fujiwara & Kamataki (2007) (Fig. 9H).

Heterolthic alternations: wave-dominated intertidal deposits

The heterolthic (mudstone-sandstone) alternations consist of decimetre to metre thick packages of alternating mudstones and fine-quartz sandstones (with local mica, muddy clasts, glauconite and oysters), which mainly intercalate with and pass laterally to the M facies of the poorly-drained coastal mudflat (Fig. 7). In detail, these alternations include sand-dominated and mud-dominated beds, with thicker sandstone beds usually showing HCS (decimetre wavelength) and/or ripple cross-lamination (current, wave and combined-flow ripples) (Fig. 10A to C). Mudstone-dominated beds mainly consist of lenticular or wavy bedding, including centimetre thick sandstone beds with ripple cross-lamination (Fig. 10D and E). Bioturbation is only locally present at the top of the alternations (Fig. 10C).

The lateral change of the heterolthic facies to poorly-drained coastal mudflat deposits reflects that these sediments probably represent intertidal sediments. The observed
features, in particular the presence of mud-dominated and sand-dominated beds with thick sandstone beds showing hummocky cross-stratification (HCS), have been described in intertidal zones of open-coast tidal flats (e.g. Yang et al., 2005; Basilici et al., 2012; Fan, 2013). In such areas, the interaction of waves and tides produces storm-generated, sand-dominated layers and post-storm, mud-dominated layers (Fan, 2013). In particular, Yang et al. (2005) describe thinning-upward and fining-upward successions in which the thick sandstone deposits with HCS are related to the major winter storms, whereas the finer deposits including thin sand beds correspond to deposition from smaller storms and tides in summer. Thinning-upward and fining-upward successions that are rather similar both in thickness and sedimentary structures are observed in the heterolithic (mudstone–sandstone) alternations studied here (Fig. 10D to F).

Cross-bedded sandstones: tide-influenced fluvial channels and tidal bars

Lenticular packages of cross-bedded sandstones (up to 5 m thick and 20 to 75 m wide, locally up to 150 m wide) are intercalated both in the red mudstones and M facies of the coastal mudflat and in the intertidal heterolithic facies (Fig. 7). The sandstones are formed by fine-sized to medium-sized quartz grains with sporadic centimetre-sized muddy clasts (red mudstones and mixed facies), oyster fragments and glauconite grains. On the basis of their geometry and internal architecture, two main types of lenticular sandstone packages are recognized:

1 Sandstone packages with a concave-upward base and a convex-upward to planar top, generally embedded in red mudstones. These are exemplified by level Ag1-1 (see also levels Ag1-3 and Ag2-2 in Fig. 4D and G). The packages are isolated; i.e. at outcrop scale they do not have equivalent sandstone packages in the same stratigraphic level. The sandstone package Ag1-1 includes from base to top (Fig. 11): conglomerate beds bearing limestone, sandstone and muddy clasts; low-angle, E-oriented, master bedding (lateral
accretion) surfaces with internal planar (local trough) cross-bedding with south to south-east palaeocurrents (i.e. along the strike of the master surfaces), with local alternating fine-grained and coarse-grained laminae; and bidirectional cross-beds with mud drapes and mud clasts. The lateral termination of the sandstone package is sharp, and the bioturbation is absent (see also the sharp tops of packages Ag1-3 and Ag2-2 in Fig. 4D and G).

The sedimentary features of these sandstone packages indicate that they correspond to tide-influenced fluvial channels. The conglomerate beds are interpreted as an initial channel fill, with clasts mainly derived from the erosion of channel bank deposits (muddy clasts), but they might also be deposited by storms or spring tides from neighbouring coastal areas (for example, limestone clasts) (Fig. 11E). The lateral accretion surfaces reflect the lateral migration of the point bar, and the alternating fine-grained and coarse-grained laminae and the bidirectional cross-beds with mud drapes and mud clasts indicate a tidal influence. The dominance of sand and the absence of bioturbation in tidal and tide-influenced fluvial channels reflect the winnowing of fines during ebb and flood tides in combination with storm events. The sharp lateral termination of the sandstone packages and the absence of crevasse deposits can be related to the diminished effects of long-lasting fluvial floods (e.g. Musial et al., 2012; Brivio et al., 2016).

2 Sandstone packages with a concave-upward base and a convex-upward top that are intercalated in red mudstones and M facies. These deposits are exemplified by levels Ag6-1 and Ag6-2 (see also levels Ag4-1 and Ag5-1 in Fig. 4B). Most of them are encased in the erosional surfaces of sequences 4 and 5 (Ababuj outcrop; see Fig. 7). By contrast with the isolated tide-influenced fluvial channels, these packages are grouped; i.e. at outcrop scale there are two or three equivalent sandstone packages in the same stratigraphic level (see Fig. 4B). These deposits (Fig. 12) are characterized by master bedding surfaces that are horizontal or dip slightly to the west, and show internal cross-beds with palaeocurrents mostly to the south and south-east but also in opposite directions. The top surface of the packages is sharp, and bioturbation is absent.
These sandstone packages are interpreted as tidal bars. The wedge-shaped cross-strata with scarce channelized erosion surfaces (Longhitano et al., 2012) and bidirectional palaeocurrents (Allen, 1980) are criteria that differentiate tidal bars and dunes from cross-bedded fluvial units. The key criterion for identification of tidal bars is that their long axis is almost parallel both to the tidal current direction and to the strike of the lateral accretion master surfaces (Olariu et al., 2012), as is observed in the studied sandstone packages (Fig. 12C). Even taking into account a certain post-depositional compaction, the convex-upward top of the packages (Fig. 11A and B) is likely to reflect the geometry of the bar in oblique section (Fig. 11C), with the palaeocurrent of the cross-beds oriented perpendicular to the master bedding surfaces and reflecting the tidal currents. The presence of several sandstone packages in the same stratigraphic level (for example, two packages in level Ag4-1, two packages in level Ag6-1 in Fig. 4B) reflects the fact that the tidal bars are aligned and separated by a distance of 100 to 200 m.

Like the packages interpreted as tide-influenced fluvial channels, the described tidal bars are sand-dominated. Key features of tidal point bars and tidal bars, such as mud laminae, inclined heterolithic stratification, a fining-upward trend and bioturbation (Dalrymple, 2010; Longhitano et al., 2012), are absent in the studied deposits, probably reflecting the generally high-energy of ebb and flood tides and/or the short duration of slack water periods, and also possible storms contributing to the winnowing of fines.

Fenestral-laminated lime mudstones and peloidal packstones–grainstones: wave-dominated supratidal to shallow subtidal deposits

The fenestral-laminated lime mudstones are arranged in decimetre thick beds and include a set of subfacies (i.e. fenestral lime mudstone, laminated lime mudstone and pure lime mudstone) that are intercalated at the bed and centimetre scale (Fig. 13). The pure lime mudstones have very scarce grains, with scattered thin parallel-laminated and cross-
laminated accumulations of small-sized grains (bivalves, characean algae, gastropods, ostracods, miliolids, lituolids and scattered ooids). Thalassinoides and escape traces of bivalves and fenestral porosity are locally present (Fig. 13A to D). The fenestral lime mudstones are characterized by millimetre-sized and spar-filled fenestral pores that are elongated parallel to the stratification, both as widely spaced pores and forming dense fenestral laminites (Fig. 13A; see also Fig. 14F). The laminated lime mudstones are formed by irregular, submillimetre-thick muddy laminae (probably microbial in origin), with fenestral porosity (Fig. 13D). Locally, the fenestral and laminated subfacies have desiccation cracks, dinosaur tracks preserved as concave epireliefs, centimetre thick flat pebble conglomerates (FPCs) and root traces (Fig. 13E and F).

The presence of structures indicative of subaerial exposure (fenestral and laminated subfacies) indicates that the sediments were probably deposited in supratidal to intertidal areas (e.g. Shinn, 1983; Pratt, 2010). The FPCs were formed after erosion by strong tides or storms. By contrast, the scarce fenestral porosity in the pure lime mudstone subfacies points to very shallow subtidal sediments. The muddy texture and the scarcity of skeletal grains indicate deposition in low-energy and restricted conditions, with local high-energy events generating millimetre thick accumulations of bioclasts reworked from neighbouring areas. Similarly, the laminated subfacies can be interpreted as produced by the trapping and binding by microbial mats of the particulate sediment washed in from subtidal areas (e.g. Riding, 2011). The presence of both continuous and discontinuous levels of fenestral-laminated lime mudstones changing laterally to coastal red mudstones and M facies (Fig. 7) indicates that this facies probably represents deposition both in small, very shallow muddy ponds in the coastal mudflat, which repeatedly emerged into inter-supratidal conditions, and in an intertidal/shallow subtidal belt. In this context, restriction would probably be controlled by the shallow depth of the water and/or fluctuations in salinity and temperature.

The peloidal packstones–grainstones are formed by lithic peloids (i.e. fragments of lithified carbonate mud; e.g. Flügel, 2010), and minor proportions of silt-size quartz grains,
skeletal grains (mainly miliolids and small bivalves, as well as scattered gastropods, ostracods, textulariids, echinoids and characean algae), millimetre sized lime mudstone intraclasts and micritized ooids (Fig. 14F). This grainy facies has frequent tractive structures, including HCS (decimetre wavelength), parallel lamination and ripple cross-lamination (current, wave and combined-flow ripples), and convolute lamination (Fig. 14A to C). Locally, root traces, desiccation cracks and burrows including Thalassinoides are also present (Fig. 14D and E). The peloidal facies can be interbedded at package scale with the fenestral-laminated lime mudstones (Fig. 7), but also at bed scale, forming centimetre thick fining-upward (peloidal to lime mudstone) levels, with a sharp base and bioturbated top (Fig. 14D).

The grainy texture and abundance of tractive structures in the peloidal facies reflect relatively high-energy conditions during deposition. The following evidence indicates that this facies represents wave and storm reworking of very shallow subtidal muddy sediments and their redeposition landward: (i) the nature of the dominating grain (lithic peloids) reflecting generation by breakage and reworking of originally muddy sediments (Flügel, 2010); (ii) the abundance of wave-generated and storm-generated structures; (iii) the interbedding with the fenestral-laminated facies and in particular the observed fining-upwards peloidal to muddy intercalations resembling storm beds (Fig. 14D); and (iv) the interbedding as thin peloidal intercalations in coastal red mudstones and M facies (Fig. 7), with local root traces and desiccation cracks, reflecting the fact that they are storm beds that have accumulated in the coastal mudflat.

**Bioturbated bioclastic wackestones–packstones and oolitic packstones–grainstones: open shallow-water deposits**

These two facies form very local intercalations within the coastal and tidal facies (mixed, heterolithic, fenestral-laminated and peloidal facies), and locally stack in thick packages (see oolitic level Ce7-5 in Fig. 7). The bioclastic wackestones–packstones are mainly intercalated
in the poorly-drained coastal M facies and form a continuous level in sequence 5 (Fig. 7). This facies has an abundance of whole bivalves (including oysters) and gastropods, and minor proportions of benthic foraminifera (miliolids, textulariids and lituolids), ostracods, serpulid patches, and characean and dasycladacean algae (Fig. 15A, B and D). Micritized ooids, lime mudstone intraclasts and silt-sized quartz grains are also present. The facies is bioturbated (including Thalassinoides traces) and locally has dinosaur tracks preserved as convex hyporeliefs (Fig. 15A). The abundant bioturbation and the predominance of carbonate mud and whole bivalves reflect deposition in low-energy areas. However, compared with the M facies with which it intercalates, the bioclastic facies has a high variety of skeletal remains, including dasycladacean algae (Fig. 15D), thus reflecting more open (non-restricted) marine waters (e.g. Aguirre and Riding, 2005). The discrete levels of bioclastic facies are thus thought to correspond to temporary incursions of open marine waters within the coastal–tidal area.

The oolitic packstones–grainstones include similar skeletal components to the bioclastic facies, but are dominated by ooids and minor proportions of peloids, lime mudstone intraclasts and silt-sized quartz grains. The ooids have bioclastic and peloidal nuclei and cortices with thin sparitic and micritic laminae (type 1/3 ooids following Strasser, 1986; Fig. 15C). Isolated oolitic beds usually have dinosaur casts and intraclastic levels at their base, and burrows (Thalassinoides) and local root traces on top (Figs 6 and 15A). The presence of similar skeletal debris in the oolitic facies, combined with the grainy texture with ooids characteristic of agitated shallow waters (Strasser, 1986), indicates that it was generated in high-energy, open marine conditions. The isolated oolitic levels frequently sealing dinosaur tracks (Fig. 6) are interpreted as storm beds intercalated in coastal–tidal facies, and the thick oolitic package recorded in Los Cerezos (level Ce7-5, Fig. 7) is thought to correspond to local oolitic banks.
SEDIMENTARY MODEL

The Aguilar del Alfambra Formation encompasses a wide variety of terrigenous clastic (T), mixed terrigenous–carbonate (M) and carbonate (C) facies that reflect deposition contexts ranging from a coastal mudflat to a wave-dominated tidal flat. The T and M facies encompass red mudstones and mixed facies of coastal mudflats, with intercalated sheet-flood sandstones and tsunami conglomerates, tide-influenced fluvial channels and tidal sandstone bars (cross-bedded sandstones), and heterolithic facies of wave-dominated intertidal flats. The carbonate (C) facies also include supratidal to shallow subtidal wave-dominated deposits (fenestral-laminated lime mudstones and peloidal packstones–grainstones) and local open shallow-water facies (bioturbated bioclastic wackestones–packstones and oolitic packstones–grainstones).

Figure 16 illustrates the dimensions and lateral relationship of the defined facies and related subenvironments. The facies recorded in the lower part (sequences 1 to 3) and in the upper part of the unit (sequences 4 to 8; Fig. 7) show significant differences. Sequences 1 to 3 are homogeneous in thickness along the studied transect and mainly include red mudstones, mixed facies and peloidal and fenestral-laminated facies. By contrast, sequences 4 to 8 show a large variation in thickness, local erosive bounding surfaces and a greater heterogeneity of facies, including a local but major development of heterolithic facies.

Sequences 1 to 3 represent deposition in a mixed clastic–carbonate coastal mudflat and wave-dominated tidal flat (Fig. 16A and B). Red mudstones became concentrated in the well-drained coastal mudflat, whereas the proportion of carbonate sediments increased seawards. The coastal mudflat included poorly-drained depression areas or small ponds flooded during major tides and storms, in which mixed and carbonate muddy sediment accumulated (i.e. M facies and fenestral-laminated lime mudstones). Discrete tide-influenced fluvial channels (cross-bedded sandstones) and sheet-flood deposits (tabular sandstones)
developed within this coastal mudflat. Tide and storm currents explain the predominance of sand in the tide-influenced fluvial channels. Local conglomerate tsunami deposits are also recorded in the coastal mudflat. The intertidal to shallow subtidal areas were also the loci of deposition of mixed and carbonate muddy sediments, with mixed clastic–carbonate muds concentrated around the main high-tide level and fenestral-laminated carbonate muds preferentially located in intertidal to shallow subtidal areas. The key facies for interpreting the episodic wave (storm)-dominance on this tidal flat is the peloidal facies, which originated from the episodic reworking of the shallow subtidal carbonate muds; the lithic peloids generated in this way accumulated as storm beds in the intertidal and shallow subtidal area and even over the coastal mudflat. The intertidal to shallow subtidal belt is thought to be no more than 3 km in width (Fig. 16B).

The facies in sequences 4 to 8 allow two laterally related areas to be defined (Fig. 16C and D): (i) a mixed clastic–carbonate coastal mudflat and wave-dominated tidal flat recording oolitic storm beds and local subtidal oolitic banks; and (ii) an embayed area generated by local extensional tectonic activity in which clastic sediments were trapped. In the northern part of this embayed area (Aguilar, where the successions are reduced in thickness and have significant erosion surfaces; Figs 7 and 16D), the tectonic uplift of the coastal mudflat generated small incised valleys (see the irregular erosion surfaces in Fig. 4 and 7) in which sandy sediment accumulated forming ‘encased’ tidal bars. To the south (Los Cerezos), muddy and heterolithic sediments accumulated in a relatively (a few kilometres) wide and subsident intertidal zone. The key facies for interpreting the episodic wave (storm)-dominance on this tidal flat is the intertidal heterolithic facies with HCS and ripple structures, which recorded the episodic storm reworking of muddy and sandy sediment. Local incursions of open-marine waters from southern and eastern sectors lead to the generation of bioclastic layers with shallow (non-restricted) marine fauna (Fig. 16D). The intertidal clastic belt is thought to be more than 5 km wide (Fig. 16D).

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The Aguilar-Los Cerezos successions in the Galve sub-basin thus reflect coastal environments with coeval carbonate and clastic sedimentation controlled by both tidal and wave (storm) action. The tidal influence is recorded in several of the defined facies (for example, tide-influenced fluvial channels, tidal bars, fenestral-laminated lime mudstones and heterolithic facies), and key carbonate and clastic facies for interpreting the wave (storm) influence in the tidal flat are the peloidal packstones–grainstones and the heterolithic facies, respectively.

DISCUSSION

Facies distribution and sedimentary evolution at basin scale

The sedimentary model proposed for the Aguilar del Alfambra Formation in the central area of the Galve sub-basin reflects the fact that terrigenous clastic and carbonate sediments accumulated coevally in the coastal mudflat and wave (storm)-dominated tidal flat (Fig. 16). However, stages with preferential accumulation of clastic vs. carbonate sediments are present (i.e. the T-dominated and C-dominated intervals in T/C sequences 1 to 8: see Fig. 7). Integration of the proposed sedimentary model at basin scale (i.e. Maestrazgo Basin) and comparison with coeval reference sections may prove useful in discussing the significance of the predominance of carbonate versus terrigenous sediments and the factors controlling sedimentation.

During the late Tithonian to middle Berriasian interval under study, there was significant clastic input from the exposed areas of the Teruel–Valencia High across the south-western margin of the Maestrazgo Basin (Fig. 17A), including the Galve sub-basin (the Aguilar–Los Cerezos area and the north-western Galve and Abeja sections; Aurell et al., 2016) and the Penyagolosa sub-basin (for example, the Montanejos section; Bádenas et al., 2004). By contrast, in the northern marginal areas (i.e. the less subsident El Perelló and Morella sub-basins; for example, the Luco de Bordón and Las Parras sections), carbonate tidal flat
sedimentation dominated. In the depocentral areas of the basin (i.e. the Salzedella sub-basin) sedimentation occurred in a wide shallow carbonate platform (e.g. Aurell et al., 2003, 2010).

In the Galve sub-basin, the correlation of the eight T/C sequences recorded in Aguilar–Los Cerezos with the north-western marginal sections (the Galve and Abeja sections; Fig. 17B) indicates that these sequences are not continuous at sub-basin scale. The C-dominated intervals of the sequences (i.e. supratidal to shallow subtidal fenestral-laminated and peloidal limestones) pinch out towards the north-west to clastic coastal mudflat deposits, with a decreasing number of C-dominated intervals towards Abeja (five intervals) and Galve (one interval). This confirms that there is a lateral relationship of terrigenous and carbonate sediments also at sub-basin scale and indicates that the observed T/C sequences are not the product of reciprocal sedimentation (e.g. Schwarz et al., 2016). According to the alternative model of coeval sedimentation, the C-dominated intervals would represent transgressive deposits (i.e. relative deeper facies; see Schwarz et al., 2016, fig. 1), but this does not fit with the coexistence of carbonates and siliciclastics recorded in the intertidal belt (Fig. 16). As pointed out by Labaj & Pratt (2016), in mixed carbonate–siliciclastic systems the interaction of sediment production/supply and/or accommodation space would lead to a complex facies architecture comprising clastic and carbonate sediments. In the studied case, the observed thickness and facies variation and the evidence of synsedimentary tectonic activity described in marginal areas of the Galve sub-basin (Aurell et al., 2016) suggest strong tectonic control on accommodation. In particular, in the uppermost sequences 4 to 8, the erosional sequence boundaries seem to reflect local tectonic uplift of the coastal mudflat area that generated small incised valleys in which sandy sediment accumulated in ‘encased’ tidal bars, whereas heterolithic sediments concentrated in a subsident intertidal zone (Fig. 16). Despite this evidence of sedimentary tectonic activity, climatic control on the supply of terrigenous sediment cannot be ruled out.
The sequential organization is quite different in the reference sections of the Morella and Penyagolosa sub-basins (Fig. 17). In the northern Morella sub-basin, peritidal carbonates are stacked in metre thick muddy and grainy shallowing-upward sequences, with supratidal caps (Ipas et al., 2005; Aurell et al., 2010). Intervals dominated by shallow subtidal (high-energy) grainy carbonates are dominant in the middle part of the succession, reflecting open marine influence similar to the observations in Aguilar–Los Cerezos (Fig. 7). By contrast, in the Montanejos section of the Penyagolosa sub-basin, sedimentation mainly took place in a mixed carbonate–clastic platform setting, with third-order sequences and parasequences dominated either by carbonates (grainy and muddy), clastics or both (Bádenas et al., 2004). The difference in sedimentary systems and stratigraphic architecture recorded in the Galve and Penyagolosa sub-basins (mixed clastic–carbonate successions) and the Morella sub-basin (pure carbonate successions) lends support to the idea of the complex interaction of tectonics and climate in controlling accommodation changes and sediment production and redistribution.

Controls on the mixed clastic–carbonate, wave-dominated tidal flat: an open-coast tidal flat

The development of a mixed clastic–carbonate, wave-dominated tidal flat during the late Tithonian to middle Berriasian in the Galve sub-basin was mainly controlled by the local palaeogeographic, tectonic and palaeoclimatic context (Fig. 17A):

1 The Galve sub-basin was located in the north-west marginal (coastal) area of a wide C-shaped embayment open towards the Tethys Ocean, which gradually passed over to the shallow platform located to the east (i.e. the Salzadella and eastern Penyagolosa sub-basins; e.g. Bádenas et al., 2004). According to palaeoclimatic reconstructions for the Berriasian (Golonka & Krobicki, 2001, and references therein), the study area would have been subjected to summer winds blowing from the north-west and mainly parallel to the
coastline, but during winter the winds would have come from the east in a mainly onshore direction. The mixed clastic–carbonate, wave-dominated tidal flat described here was thus an open-coast tidal flat seasonally dominated by winter storm waves, similar to the clastic tidal flat of the present-day Yellow Sea coast in Korea (Yang et al., 2005).

2 Within the Maestrazgo Basin, lateral variation of coastal environments occurs. The mixed clastic–carbonate, open-coast tidal flat of the Galve sub-basin is located in the transition area between a pure carbonate tidal flat to the north (Morella sub-basin) and a shallow mixed platform to the south (southern Penyagolosa sub-basin). In spite of the location of the basin within a low-latitude arid climate (following Fölmi, 2012; Hay & Floegel, 2012), no evaporite-bearing sediments were deposited, by contrast with the coeval coastal embayment in north-east Iberia (Cameros Basin; Quijada et al., 2013). This reflects the fact that sedimentation was not primarily linked to the generally arid climate conditions, but most probably depended on its palaeogeographic location on an open coast that was influenced by the local climate and tectonics, which controlled the source and accumulation of carbonates and siliciclastics. Clastics were mainly sourced from the uplifted areas of the Teruel–Valencia High and mainly accumulated in the Penyagolosa sub-basin, forming part of the terrigenous muds and sand-size grains reworked by tides and storms to be deposited in the mixed clastic–carbonate, open-coast tidal flat. Carbonate mud and sand-sized carbonate grains were probably provided by tides and storms from the nearby shallow subtidal areas (peloidal storm-generated facies) and from the open marine platform located to the east (for example, carbonate mud, ooids and open marine bioclasts). Moreover, it is interesting to note that most of the exposed northern and western areas consisted of uplifted Jurassic rocks (marls and limestones), and spring flows could thus have provided a significant additional source of carbonates, especially in the poorly-drained coastal ponds.

3 Differential tectonic uplift and subsidence in the Galve sub-basin were key factors controlling the accumulation of carbonate and clastic sediments in tidal areas. As a whole, the sedimentation compensated for the subsidence, so a general low-gradient sedimentary

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surface was maintained throughout deposition. As with the ancient clastic wave-dominated open-coast tidal flat studied by Basilici et al. (2012), rapid burial would have been a key factor preventing the reworking of the storm features by fair-weather waves and tides, as well as favouring the preservation of dinosaur tracks on coastal mudflats and intertidal muds.

**Mixed processes and sediments: comparison with similar present-day settings**

An improved understanding of the described mixed coastal system in question, regarding both processes and the resulting deposits, can be gained by drawing comparisons with some key modern examples.

1 **Mixed clastic–carbonate systems.** Knowledge of the factors that influence the mixing of carbonate and clastic sediments is necessary when it comes to generating accurate geological models that provide these spatially and temporally complex mixed systems with a certain degree of predictability (McNeill et al., 2004). Some publications dealing with modern mixed clastic–carbonate systems have emphasized that the depth and mode of mixing depend largely on the hydrodynamic regime and the type of carbonate factory, due to the variations in the light-dependence and response to suspended siliciclastics and turbidity in the water column (e.g. Halfar et al., 2004). Nevertheless, other controlling factors can cause the depth, proportion and scale of mixing to vary within a given carbonate factory. For example, in the northern Bay of Safaga (Red Sea), which is a small, shallow mixed system with photozoan carbonates, siliciclastics are concentrated on the low-energy coast, and variable mixed sediments occur in nearby subtidal areas (a few kilometres away), due to variable energy conditions and sources of clastic sediments (flash floods, aeolian transport, reworking of underlying rocks and relict sediments; Piller & Mansour, 1990, 1994). By contrast, in the Abrolhos Shelf (eastern Brazilian coast), which is one of the largest modern examples of a tropical mixed carbonate–siliciclastic system, clastic and mixed facies occur nearshore and on the mid-shelf, respectively, but intrabasinal variation also occurs due to
the presence of palaeotopographic depressions where clastic sediments accumulate (D’agostini et al., 2015).

In the mixed clastic–carbonate coastal system described here, clastics and carbonates interfinger and mix in the tidal area at scales of a few tens to hundreds of metres, due to a suite of interrelated tectonic, palaeogeographic and hydrodynamic factors. Particularly significant among these factors is its location in a tectonically controlled, open-coast embayment in the linking area between clastic and carbonate (photozoan) shallow marine environments, where clastic and carbonate sediments accumulated through the interaction of tidal and storm processes.

2 Interaction of waves and tides. Whereas tides show a regular, predictable cycle, waves exhibit great variability in energy and period (from days and weeks to seasonal variations); accordingly they can interact via mutual muting, modulation or amplification (Anthony & Orford, 2002). In modern open-coast intertidal mudflats, wind waves and tides interact to produce increased suspended sediment concentrations and settling rates mainly during spring tides and weak offshore winds (Zhu et al., 2017). Also, the storm-wave processes of offshore swells propagating onshore are greatly modulated by tides, as fluctuating tidal levels control the depth of the wave-breaking zone and the entrainment capacity of onshore currents (Fan et al., 2006). It is likely, therefore, that in the open coastal area analysed in present study, the interaction of tides and wind waves in fair-weather conditions would favour the deposition of muddy sediments (mixed, heterolithic and fenestral-laminated muds); during storms, tidal fluctuation would control the upslope and downslope shift of the wave-breaking and erosion zone and the variation in the onshore transport of sand-size sediment of the heterolithic and peloidal facies, with minor onshore transport probably during weak neap flood currents and during ebb currents (Fig. 18A).

As regards the sedimentary structures generated by storm waves, observations of present-day open-coast tidal flats reveal there to be a systematic cross-shore change in the
nature of storm deposits due to the frictional attenuation of storm waves as they propagate across the gently sloping tidal flat, with ripples dominating in the inner intertidal area, and parallel lamination and HCS in the outer intertidal area (Yang et al., 2005). A similar distribution of sedimentary structures can be proposed for the sand-size storm deposits of intertidal heterolithic and peloidal facies (Fig. 18A), with parallel lamination and HCS in the outer intertidal area, and wave, combined-flow and current-ripple structures in the inner part depending on variable orbital velocities and/or unidirectional flows associated with waves and tides (see the phase diagrams of Perillo et al., 2014). In addition, it is likely that during a storm event, these zones with preferential structures would shift upslope and downslope in turn with tidal cycles. The vertical distribution of the sedimentary structures observed in some storm beds is possible evidence of the tide-modulated storm deposits that are generated (Fig. 18B).

CONCLUSIONS: KEY FEATURES OF THE MODEL

The sedimentary model of the late Tithonian to middle Berriasian succession of the Galve sub-basin (north-east Spain) is the first documented fossil example of a mixed (clastic and carbonate), wave (storm)-dominated open-coast tidal flat. It encompasses a wide variety of terrigenous clastic, mixed terrigenous clastic–carbonate and carbonate facies deposited in areas ranging from a coastal mudflat area to the shallow portion of a subtidal area. The proposed model improves current knowledge of facies and processes in mixed coastal systems and may also be applicable to the analysis of both carbonate and clastic coastal successions in the sedimentary record. Its key features are as follows:

1 Extension and sub-environments. The mixed clastic–carbonate, open-coast tidal flat was located in an embayed area of a wide, C-shaped coast, in the linking area between a carbonate coastal (tidal) system to the north and a mixed shallow carbonate platform to the south. It is estimated that it was ca 100 km² in area, with a wide and low-gradient coastal
mudflat and a narrow intertidal to shallow subtidal belt (3 to 5 km wide). The coastal mudflat area included well-drained, locally vegetated areas represented by red mudstones, poorly-drained areas with marine influence represented by grey mudstones–marly limestones and bioturbated silty limestones, as well as scattered tide-influenced channelized fluvial sandstones and fluvial sheet-flood sandstones and tsunami conglomerates. The intertidal to shallow subtidal belt was carbonate-dominated (fenestral-laminated lime mudstones, peloidal packstones–grainstones) with local records of an open marine influence (bioturbated bioclastic wackestones–packstones, oolitic packstones–grainstones). Laterally, clastic-dominated tidal areas were developed, characterized by sandy tidal bars and intertidal heterolithic (mudstone–sandstone) alternations.

2 Processes and key facies. The non-restricted open-coast tidal flat faced the winter winds and was thus seasonally dominated by wave storms. A significant part of the clastic and carbonate sediment was sourced from tidal currents and storms from lateral clastic-dominated and carbonate-dominated coastal and shallow subtidal areas. Despite the arid climate, evaporite-bearing sediments were absent, mainly because of the lack of restriction. The tidal influence is recorded in several facies (for example, tide-influenced fluvial channels, tidal bars, fenestral-laminated lime mudstones and heterolithic facies), and wave-storm action is mainly indicated by: (i) sand-dominance in the tide-influenced fluvial channels and tidal bars; (ii) storm-related sedimentary structures (i.e. hummocky cross-stratification) in the intertidal heterolithic (mudstone–sandstone) alternations and peloidal packstones–grainstones; (iii) discrete peloidal storm beds in the coastal mudflat and oolitic storm beds in the tidal flat. Key facies for interpreting the wave (storm)-dominance in the tidal flat are the peloidal packstones–grainstones and the heterolithic facies. Tidal fluctuation would control the upslope and downslope shift of the breaking and erosion zone of the storm waves and the variation in the onshore transport of sand-size sediment, as well as the lateral variation in sedimentary structures generated by storm waves in the intertidal areas.
3 Preservation in the sedimentary record. Subsidence and high rates of sedimentation were key factors controlling the accumulation of carbonate and clastic sediments in tidal areas. Sedimentation kept pace with subsidence, so that a general low-gradient sedimentary surface was maintained throughout deposition. Rapid burial preserved the storm features in tidal sediments from reworking by fair-weather waves and tides.

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FIGURE CAPTIONS

**Fig. 1.** (A) Wave-tide binary process-based classification of terrigenous clastic coats by Yang et al. (2005), after modifying the scheme of Boyd et al. (1992) to include wave-influenced and tide-influenced open coast tidal flats. Reference works for modern and ancient examples are included. (B) Similar wave-tide binary process-based classification for carbonate and mixed terrigenous clastic–carbonate coats. Modern and ancient examples are included in wave-dominated coasts or in tidal flats. There are no modern or fossil carbonate and mixed terrigenous clastic–carbonate coastal areas potentially equivalent to clastic open coast tidal flats.

**Fig. 2.** (A) Location of the three studied outcrops of the Aguilar del Alfambra Formation in the Galve sub-basin in the western part of the Maestrazgo Basin (north-east Spain). (B) Geological mapping around the Galve sub-basin of the Mesozoic-Cenozoic units; see legend in (C). (C) Synthetic stratigraphy of the Mesozoic succession recorded in the depocentral areas of the Galve sub-basin. Adapted from Aurell et al. (2016) and Liesa et al. (2017).

**Fig. 3.** Synthetic stratigraphic section of the Aguilar del Alfambra Formation in the Aguilar outcrop, indicating the distribution of: terrigenous (T), mixed (M) and carbonate (C) facies; T/C sequences formed by a lower T-dominated interval and an upper C-dominated interval; and terrigenous and carbonate reference levels. Note that the uppermost 25 m of the unit are mostly covered.

**Fig. 4.** Satellite, drone and field images of the Aguilar del Alfambra Formation in the Aguilar outcrop, indicating the distribution of T/C sequences and terrigenous and carbonate reference levels (see their stratigraphic position in Fig. 3). (A) General view of the outcrop indicating the location of the stratigraphic section. The lateral extension of reference levels has been traced with confidence along a maximum distance of 500 m, especially in sector A; westward (sector B), some faults and covered intervals have prevented the confident

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physical tracing of some levels. **(B) to (F)** Distribution of T/C sequences and terrigenous and carbonate reference levels in sector A. Note the erosion surfaces at the Ag1–Ag2, Ag3–Ag4 and Ag4–Ag5 boundaries (see B and E), and some terrigenous and carbonate reference levels that pinch out laterally – for example, levels Ag1-1, Ag1-2 and Ag1-4 in C and D; level Ag4-1 that is encased in the depression area of the Ag3–Ag4 erosive surface in (B) and (E); levels Ag5-1, Ag6-1, Ag6-2 and Ag7-1 in (B), (E) and (F). **(G and H)** Distribution of T/C sequences and terrigenous and carbonate reference levels in sector B. Note in G the erosion surface at the Ag1–Ag2 boundary (here level Ag1-8 is absent by erosion) and terrigenous and carbonate levels Ag1-3, Ag2-1 and Ag2-2 that pinch out laterally. Note in (H) the erosion surface at the Ag3-Ag4 boundary (here level Ag3-5 is absent by erosion). **(I)** Summary of the lateral continuity of the reference levels (colours refer to facies; see legend in Fig. 3).

**Fig. 5.** The Aguilar del Alfambra Formation in the Ababuj outcrop. **(A)** Synthetic stratigraphic section, indicating the distribution of: terrigenous (T), mixed (M) and carbonate (C) facies; T/C sequences formed by a lower T-dominated interval and an upper C-dominated interval; and terrigenous and carbonate reference levels. See legend in Fig. 3. **(B)** Satellite image of the Ababuj outcrop, indicating the distribution of T/C sequences and terrigenous and carbonate reference levels that have been traced laterally along a maximum distance of 900 m; see their stratigraphic position in (A). Note that some reference levels pinch out laterally (for example, Ab1-3, Ab3-1 and Ab3-2). **(C)** Summary of the lateral continuity of the reference levels – colours refer to facies; see legend in (A).

**Fig. 6.** The Aguilar del Alfambra Formation in Los Cerezos outcrop. **(A)** Synthetic stratigraphic section, indicating the distribution of: terrigenous (T), mixed (M) and carbonate (C) facies; T/C sequences formed by a lower T-dominated interval and an upper C-dominated interval, and terrigenous and carbonate reference levels. See legend in Fig. 3. **(B) to (E)** Satellite and field images of Los Cerezos outcrop, indicating the distribution of T/C sequences and terrigenous and carbonate reference levels that have been traced laterally.

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along a maximum distance of 150 m; see their stratigraphic position in (A). Note that some terrigenous reference levels pinch out laterally (for example, Ce2-1, Ce4-1 and Ce5-3). (F) Summary of the lateral continuity of the reference levels – colours refer to facies; see legend in (A).

Fig. 7. Correlation of facies of the Aguilar del Alfambra Formation between the three studied stratigraphic sections. The main facies involved within these sequences are also indicated (see right).

Fig. 8. Field images of red mudstones and mixed facies (grey mudstones–marly limestones and bioturbated silty limestones). (A) Red mudstones. (B) Grey mudstones. (C) Root traces in red mudstones. (D) Root traces in marly limestones. (E) and (F) Thalassinoides traces (E) and dinosaur tracks preserved as concave epireliefs (F) in bioturbated silty limestones.

Fig. 9. Field images of tabular sandstones and conglomerates. (A) and (B) Sandstone beds with Taenidium (A) and Arenicolites (B) traces. (C) to (E) Grain-supported conglomerate–sandstone bed with muddy clasts of red mudstone (Mc), limestone clasts (Lc) and sandstone clasts (Sc). The bed is formed by normal-graded sublayers and is interpreted as a conglomerate-dominated tsunami deposit – see (E), similar to that described by Fujiwara & Kamataki (2007). The key for sedimentary structures is indicated in (E). (F) to (H) Matrix-supported conglomerate–sandstone bed, with muddy clasts of red mudstone (Mc), limestone clasts (Lc) and sandstone clasts (Sc), interpreted as a tsunami deposit – see (H), with some similarities with sandy-type and gravelly-type tsunami deposits of Fujiwara & Kamataki (2007). Key for sedimentary structures is indicated in (H).

Fig. 10. Field images of heterolithic (mudstone-sandstone) alternations interpreted as wave (storm)-dominated intertidal deposits. (A) to (C) Sand-dominated and mud-dominated beds. Thicker sandstone beds usually have HCS. (D) to (F) Sand-dominated and mud-dominated beds form thinning-upward and fining-upward successions with thicker sandstone
beds having HCS, which resemble those described by Yang et al. (2005) in present-day intertidal areas of open coast tidal flats.

**Fig. 11.** Sandstone package Ag1-1 (A) to (D), interpreted as a tide-influenced fluvial channel deposit (E), with conglomerate beds (1: initial channel fill); low-angle, east-oriented, master bedding surfaces (2 to 4: lateral accretion of the point bar); and bidirectional cross-beds with mud drapes and mud clasts (5: channel fill with tidal influence).

**Fig. 12.** Sandstone packages Ag6-1 and Ag6-2 (A) and (B), interpreted as tidal bars (C). The lenticular shape with a convex-upward top is likely to reflect the geometry of the bar in oblique view. Palaeocurrents of internal cross-beds are oriented perpendicular to the master bedding surfaces.

**Fig. 13.** Fenestral-laminated lime mudstones, including fenestral lime mudstone, laminated lime mudstone and pure lime mudstone subfacies. (A) Bed of pure lime mudstones (pM) and fenestral lime mudstone with dense fenestral laminites (dF) and widely spaced fenestral pores (wF). (B) Pure mudstone (pM) with escape traces of bivalves (B) and local fenestral porosity (F), and a peloidal packstone–grainstone level on top (P). (C) Pure lime mudstone (pM) with cross-laminated grainy accumulations with small skeletal grains (gl). (D) Laminated (microbial) lime mudstone (lM) intercalated with pure lime mudstone (pM) with bioclastic laminae (gL) and fenestral porosity (F). (E) and (F) Dinosaur track preserved as a concave epirelief (E) and flat pebble conglomerates (F) on fenestral-laminated lime mudstones.

**Fig. 14.** Peloidal packstones–grainstones. (A) to (C) Tractive structures, including hummocky cross-stratification (HCS) and wave ripples (WR). (D) Bed with centimetre thick fining-upward, peloidal to pure lime mudstone levels, with sharp base and bioturbated top (B). (E) Desiccation cracks in a peloidal level. (F) Thin-section image of peloidal grainstone laminae with fenestral porosity (P) on top of fenestral lime mudstones (dF: dense fenestral laminites; wF: widely spaced fenestral pores). Note the erosion surface (dashed line).

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**Fig. 15.** Bioturbated bioclastic wackestones–packstones and oolitic packstones–

grainstones. (A) and (B) Field images of these facies in level Ce7-2 (see Fig. 6 for

stratigraphic location), including casts of dinosaur tracks (convex hyporeliefs). Bioclastic

wackestones–packstones with abundant whole and disarticulate bivalves in level 2b; see

(A). (C) Thin-section image of oolitic packstones–grainstones with abundant small lituolids.

(D) Thin-section image of bioturbated bioclastic wackestones–packstones, with

dasycladacean algae – see (D) – lituolids *Anchispirocyclina lusitanica* – see (A) –
gastropods and bivalves.

**Fig. 16.** Sedimentary models for the Aguilar del Alfambra Formation in the studied area of

the Galve sub-basin, and integrated palaeogeographic maps. Sedimentary model 1 (A) and

(B) corresponds to sequences 1 to 3. Sedimentary model 2 (C) and (D) represents

deposition during sequences 4 to 8. Both models encompass mixed clastic–carbonate

coastal mudflats and wave (storm)-dominated tidal flats. The key facies for interpreting the

episodic wave (storm)-dominance on the tidal flat are the peloidal facies in model 1 and the

heterolithic facies in model 2.

**Fig. 17.** Distribution of main coastal and shallow marine sedimentary environments (A)

and main facies of some reference sections (B) in the different sub-basins of the Maestrazgo

Basin. The mixed clastic–carbonate, wave-dominated tidal flat in the Galve sub-basin

located in the eastern part of the Iberian Plate was an open-coast tidal flat seasonally

dominated by wave storms (winter and summer wind vectors after Golonka & Krobicki,

2001).

**Fig. 18.** Possible interaction of tides and storm flows in the mixed clastic–carbonate,

wave (storm)-dominated tidal flat in the Galve sub-basin (A) and resulting tide-modulated

storm deposits formed in the intertidal areas (B).
<table>
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<tr>
<th>T/G sequences</th>
<th>Terrigenous reference levels (Ab)</th>
<th>Carbonate reference levels (Ab)</th>
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Conglomerate-dominated tsunami deposit

Sandstone-dominated tsunami deposit

(see Fig. 9C of Fujiwara & Kamataki, 2007, for comparison)

(see Fig. 9A & B of Fujiwara & Kamataki, 2007, for comparison)
Wave-dominated intertidal flat deposits

1 thinning- and filling-upward succession
2 thinning- and filling-upward successions

(see Fig. 10 of Yang et al., 2005, for comparison)
A  Interaction of tides and storm flows

1. Up-slope and down-slope shift of wave-breaking and erosion zone and variation in onshore sediment transport in turn with tidal cycles
   - Wave erosion (terrigenous sand / lithic pebbles)
   - Mainly within the wave-breaking zone
   - Storm waves + spring tides or flood tides

2. Wave energy decrease / wave attenuation increase
   - Wave breaking
   - Wave damping
   - Reduced onshore sediment transport (due to weak neap flood currents or wave propagation)

- Offshore sediment transport by flood currents + wave propagation

See B

Ripples and ripple cross-lamination
(wave, combined-flow & current ripples)

HCS-hummocky cross-stratification and parallel lamination

B  Tide-modulated storm deposit

- HCS-hummocky cross-stratification and parallel lamination
- Ripple cross-lamination

- cRR current ripples
- cR combined-flow ripples

- Storm beds
  - Heterolithic (mudstone-sandstone) alternations
  - Peloidal packstones-grainstones

- FLOOD
- LOW TIDE (starvation)
- EBB