

Responses of a hybrid carbonate shelf to progressive basin restriction: Pre- and Messinian Salinity Crisis environmental shifts in the Western Mediterranean (Tabernas Basin)[☆]

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ABSTRACT

Before and during Stage 1 of the Messinian Salinity Crisis (MSC), carbonate platforms developed in marginal basins of the Western Mediterranean, recording some of the last manifestation of fully tropical climatic conditions in the region. In the Tabernas Basin (SE Spain), a mixed carbonate-siliciclastic succession was deposited between 6.080 and 5.971 Ma, recording a detailed archive of the environmental, ecological, and sedimentological changes that occurred during an advanced phase of Mediterranean restriction preceding the Messinian Salinity Crisis. Systematic semiquantitative micropaleontological analysis reveals five different benthic foraminiferal assemblages, which respond to a progressive shallowing-upward trend and shifting environmental conditions at the seafloor. In the upper epibathyal environment, the coexistence of planktonic foraminifera with dysoxia-tolerant benthic taxa suggests pronounced stratification within the water column. Enhanced primary productivity in the upper waters fuelled organic matter flux to the bottom waters, driving pyrite precipitation and conditioning benthic ecosystems. The transition to a shelf setting was recorded by an increased diversity of shallower-water benthic fauna and the deposition of bioclastic limestones. Further environmental stress produced by progressive restriction drove a general decline in the marine fauna, culminating in the deposition of barren limestones. The flourishing of microbial mats in the inner-shelf marked the climax phase of restriction preceding evaporite deposition at the onset of the Messinian Salinity Crisis.

1. Introduction

The development and evolution of carbonate platforms in Neogene intramontane basins of the Betic Cordillera (SE-Spain; Fig. 1) were strongly influenced by climatic fluctuations, tectonic reorganization, and progressive restriction of water exchange in the Western Mediterranean during the Late Miocene. These carbonate successions mark the final phase of tropical marine conditions in the area, with their decline and subsequent disappearance closely tied to the environmental disruptions triggered by the onset of the Messinian Salinity Crisis (MSC) (Perrin and Bosellini, 2013; Agiadi et al., 2024).

A notable consequence of the MSC was the deposition of > kilometre-thick halite in deep basins, alongside gypsum and associated mixed carbonate-siliciclastic successions in the marginal basin areas (see

Krijgsman et al., 2024, for a comprehensive review). Another important consequence of the MSC was its profound impact on marine ecosystems, as several groups of marine organisms underwent a marked decline (Agiadi et al., 2024). During the Pliocene, following the crisis, tropical reefs vanished from the Mediterranean (Braga et al., 2009).

During the pre-MSC stage, carbonate deposits were widespread across marginal and intermediate-depth settings, representing a common depositional feature within the broader context of progressive Mediterranean restriction (Rouchy and Monty, 1981; Krijgsman et al., 2002; Orszag-Sperber et al., 2009; Manzi et al., 2016). The progressive restriction of the Mediterranean started during the late Tortonian, driven by enhanced tectonic activity in gateway regions, which ultimately led to the closure of marine connections between the Mediterranean and the Atlantic Ocean (Flecker et al., 2015). These

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oceanographic changes were superimposed on global cooling events (7.8–5.8 Ma, Herbert et al., 2016) and pronounced climate variability, which led to significant sea-level drops (~60 m) (Jiménez-Moreno et al., 2013; Pérez-Asensio et al., 2013; Ohneiser et al., 2015) and decreasing sea-surface temperatures (Lanterna et al., 2026) before the onset of the MSC.

The increasing restriction led to a semi-enclosed configuration of the Mediterranean, amplifying the influence of orbital-driven climatic oscillations, particularly those at the precession scale (Krijgsman et al., 1999), on sedimentation patterns and marine ecosystems, with notable effects on planktonic species (Sierra et al., 2003; Flores et al., 2005). These paleoenvironmental changes are recorded in different Messinian successions across the Mediterranean (Seidenkrantz et al., 2000; Bellanca et al., 2001; Blanc-Valleron et al., 2002; Kouwenhoven et al., 2003; Kouwenhoven et al., 2006; Iaccarino et al., 2008; Orszag-Sperber et al., 2009; Di Stefano et al., 2010; Lozar et al., 2010; Flecker et al., 2015; Corbí et al., 2020). Building on this, Corbí et al. (2020) proposed a nomenclature for a stepwise model for the progressive Mediterranean restriction during the pre-MSC stage and identified three main phases: initiation (7.16–6.76 Ma), amplification (6.75–6.35 Ma), and consolidation phases (6.35–5.97 Ma). An additional phase, named the “restriction climax”, coincides with the MSC onset as defined by Manzi et al. (2013).

During this interval, significant paleoenvironmental and ecological changes were recorded in pre-MSC successions across several Mediterranean basins, including Sicily (Bellanca et al., 2001), Northern Italy (Kouwenhoven et al., 1999; Lozar et al., 2010; Dela Pierre et al., 2011), Greece (Gennari et al., 2024; Cyprus (Kouwenhoven et al., 2006; Wade and Bown, 2006; Gennari et al., 2018), and Southeastern Spain (Sierra et al., 1999, 2001, 2003; Flores et al., 2005; Corbí et al., 2016, 2020). Many of these studies highlight common sedimentological and paleontological responses to the progressive marine restriction, reflecting synchronous abiotic and biotic changes across the Mediterranean. Among the abiotic responses, intensified water-column stratification is frequently recorded, often resulting in oxygen-depleted bottom waters (Gennari et al., 2024). Biotic responses include the disappearance of both planktonic and benthic foraminifera, a relative increase in benthic “stress-marker” taxa, and an overall decline in calcareous microfossils and marine biodiversity (Agiadi et al., 2024; Bertini et al., 2024). However, recent research highlighted the potential taxonomic bias in the pre-MSC fossil record, as early diagenetic processes may have

selectively dissolved calcareous microfossils, therefore distorting palaeoecological interpretations (Mancini et al., 2020).

In the Betic Cordillera, carbonate platforms flourished along the margins of narrow and elongated intramontane basins (Esteban, 1979; Dabrio et al., 1981; Riding et al., 1991; Martín et al., 1993; Martín et al., 1997; Martín et al., 1999; Braga and Martín, 1996; Braga and Aguirre, 2001; Bourillot et al., 2010). A well-preserved example of carbonate platform development is found at the northern margin of the Sorbas Basin (Fig. 1B), considered an onshore reference site for MSC-related studies due to its continuous sedimentary record (Esteban, 1979; Riding et al., 1991; Braga and Martín, 1996; Braga and Aguirre, 2001; Braga et al., 2006, 2009; Bourillot et al., 2010; Martín et al., 1993; Roveri et al., 2009). These shallow-water systems were highly sensitive to sea-level fluctuations, constituting valuable sedimentary archives for reconstructing palaeoceanographic dynamics prior to and during Stage 1 of the MSC (Roveri et al., 2009). Their proximity to important palaeogateways (Flecker et al., 2015) further underscores their importance for investigating environmental responses to restriction in the Western Mediterranean.

This study presents an integrated analysis of a pre-MSC carbonate-shale succession from the northern margin of the Tabernas Basin (Figs. 1B and 2), spanning the interval between 6.164 and 5.97 Ma, and forming part of a Messinian composite succession recently studied for sedimentology, bio- and magnetostratigraphy, and evaporite isotope geochemistry (Lacerda-Orita et al., 2025). This succession recorded the transition from the “consolidation” to the “climax” phase of the Mediterranean restriction, as defined by Corbí et al. (2020).

To reconstruct the paleoenvironmental, depositional, and diagenetic constraints of this interval, we employed a multi-proxy approach integrating litho- and microfacies analysis, semiquantitative micropaleontology, and bulk rock mineralogy. The main objectives of this study are to: (1) Characterize carbonate facies through systematic petrographic analysis, and reconstruct the depositional evolution of the mixed carbonate-siliciclastic shelf studied; (2) Analyse microfossil assemblages, with a focus on benthic foraminifera, to identify faunal responses to increasing restriction and bathymetric changes; and (3) Investigate early diagenetic (eogenetic) processes and products as indicators of redox conditions and water-column stratification, evaluating their implications for fossil preservation/dissolution and broader paleoenvironmental changes.

This study broadens our understanding of marginal carbonate

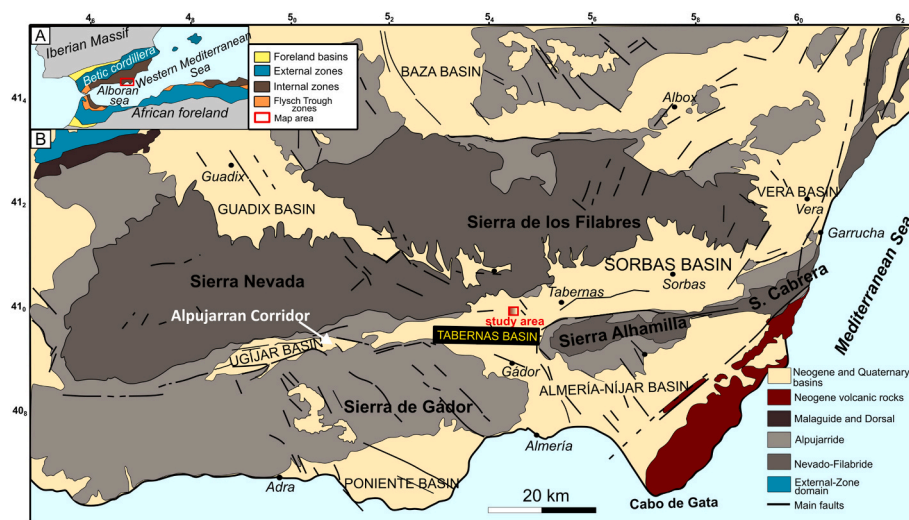


Fig. 1. A) Position of the Betic and Rif cordilleras, highlighting their main structural and depositional domains (modified from Lacerda-Orita et al., 2025; after Crespo-Blanc et al., 2016). B) Structural map of the internal zone of the Betic Cordillera, illustrating the main basement complexes and sedimentary basins (modified from Lacerda-Orita et al., 2025; after Sanz de Galdeano et al., 2010). The study area is indicated by the red square, and the location of the studied section is 37°02'40" N; 2°30'23" W. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

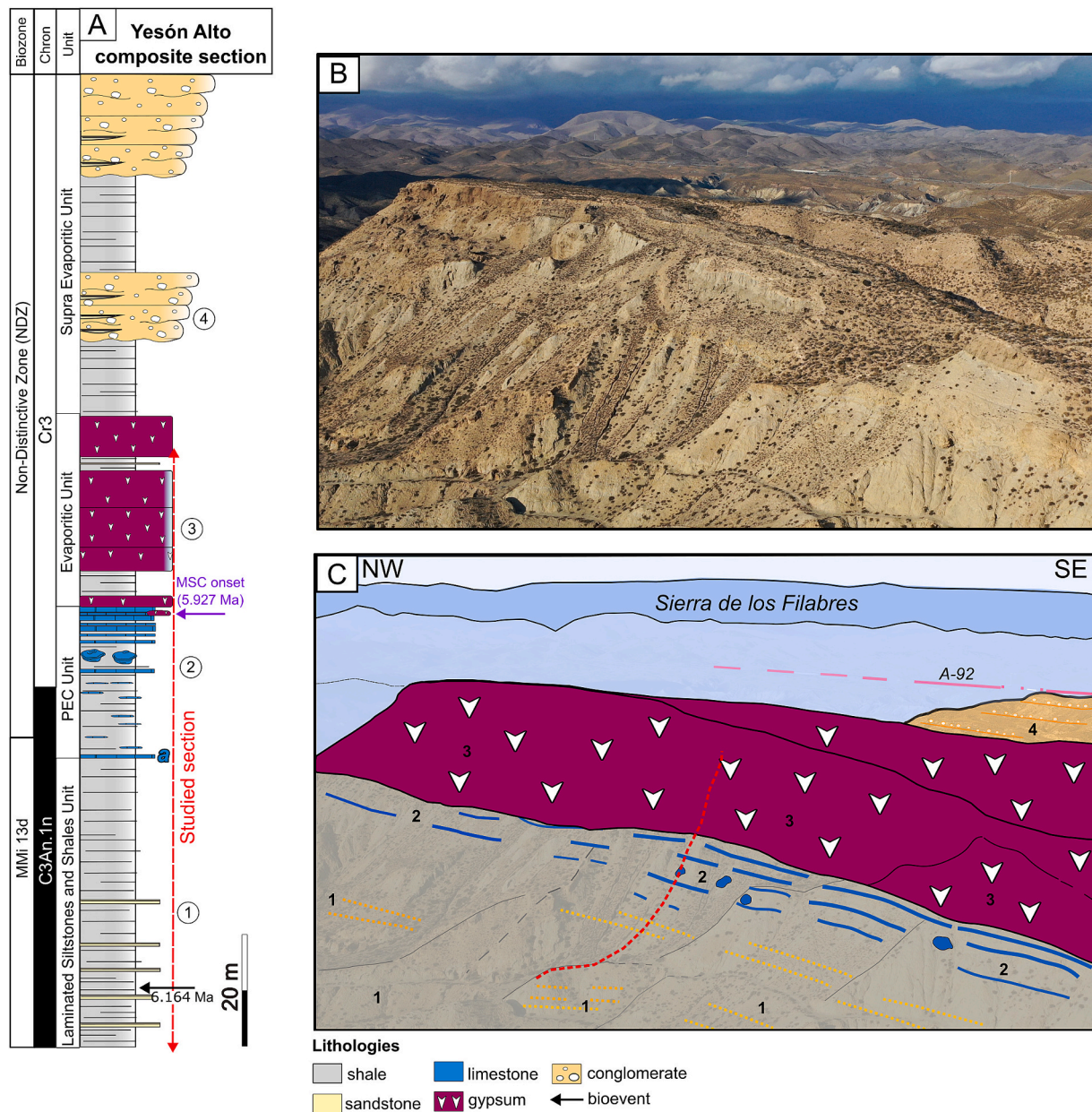


Fig. 2. A) The Yesón Alto composite section, displaying the main lithostratigraphic units (from 1 to 4). The bio-magnetostratigraphic data are based on [Lacerda-Orita et al. \(2025\)](#). The stratigraphic position of the studied section is highlighted in the dotted red line. B) Panoramic view of the study area, C) showing the main field observations and stratigraphic units. The dotted red line shows the section analysed in detail in this study. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

systems during the final stages of Mediterranean restriction, offering a high-resolution archive of environmental change immediately preceding the onset of the MSC. The Tabernas Basin succession analysed here represents a recently discovered outcrop, providing new insights into the paleoenvironmental conditions that prevailed during this time interval. The findings complement existing records from the Western Mediterranean and enhance the spatial resolution of pre-MSC reconstructions.

2. Geological context

The Betic Cordillera ([Fig. 1a](#)) was formed by the convergence between the African and Eurasian plates, representing the westernmost extension of the Alpine Orogeny in southern Europe ([Sanz de Galdeano, 1990](#); [Lonergan and White, 1997](#)). Structurally, the Betic Cordillera can

be divided into two main structural zones. The External Zones, which include the South-Iberian paleo-margin, are characterized by folded Mesozoic and Cenozoic sedimentary successions (Sub-Betic and Pre-Betic domains) ([García-Hernández et al., 1980](#); [Vera, 1988](#)). The Internal zones are represented by the Nevado-Filábride, Alpujárride, and Maláguide complex ([Weijermars et al., 1985](#); [Sanz de Galdeano, 1990](#); [Lonergan et al., 1994](#); [Braga et al., 2003](#)). The Nevado-Filábride complex is formed of Paleozoic metamorphic rocks ([Gómez-Pugnaire et al., 2000](#)), and the Alpujárride complex is composed of Paleozoic to Mesozoic metasedimentary units ([Delgado et al., 1981](#); [Martín and Braga, 1987](#)).

The Tabernas Basin (Southeastern Spain) is a Neogene intramontane basin bounded by the Nevado-Filábride and Alpujárride complexes in the internal Zones ([Fig. 1b](#)). It is approximately 10 km in length and is located between the Sierra de los Filabres to the north, the Sierra

Alhamilla to the south, and the Sierra de Gádor to the west. The basin is separated from adjacent Neogene basins, such as the Sorbas and Níjar, by structural highs. (Fig. 1B; Weijermars et al., 1985; Haughton, 2000). The sedimentary record of the Tabernas Basin is notably influenced by active tectonics (Kleverlaan, 1989; Dabrio, 1990; Haughton, 2000; Hodgson and Haughton, 2004; Baudouy et al., 2021).

A newly discovered Messinian succession (the Yesón Alto composite section) at the northern margin of the Tabernas Basin was dated to the late pre-MSC (6.184–5.97 Ma) and early stages of the MSC (Lacerda-Orita et al., 2025). In that study, a 180 m-thick composite stratigraphic section was assembled, recording the transition from the late pre-MSC interval into Stage 1 of the MSC. The Yesón Alto composite section (as described by Lacerda-Orita et al., 2025) comprises a vertically stacked succession with the following lithostratigraphic units:

(1) The Laminated Siltstones and Shales Unit, dominated by fine-grained hemipelagic deposits containing abundant calcareous microfauna, represents a basinal setting. Interbedded with the siltstones and shales, distal turbidite sandstones and slump structures indicate recurrent mass-transport processes triggered by tectonic instability along the margins of the basin. Biostratigraphically, the unit is assigned to the MMi 13d *Turborotalita quinqueloba* subzone (6.35–5.99 Ma, Sierro et al., 2001; 6.35–6.01 Ma, Blanc-Valleron et al., 2002), defined at its base by the coiling change of *Neogloboquadrina acostaensis* and at its top by the disappearance of planktonic foraminifera, criteria also documented elsewhere in the Mediterranean region (e.g., Lirer et al., 2019).

(2) The Pre-evaporitic Carbonates Unit represents the development of a hybrid carbonate shelf, marking approximately the transition from the MMi 13d subzone to the Non-Distinctive Zone (NDZ) (Lacerda-Orita et al., 2025). The NDZ is biostratigraphically defined at its base by the disappearance of planktonic foraminifera (see Lirer et al., 2019).

(3) The Evaporitic Unit is represented by the cyclic alternation of selenitic gypsum and siliciclastic deposits (Lacerda-Orita et al., 2025), controlled by orbitally induced climatic forcing (e.g., Krijgsman et al., 1999). Massive selenites precipitated during arid periods (precession maxima), while shales and sandstones were deposited during more humid periods (precession minima) (Lugli et al., 2010). An age of 5.971 Ma is assigned to the base of the first gypsum bed, corresponding to the onset of the MSC in the basin (Lacerda-Orita et al., 2025).

(4) The Supra-Evaporitic Unit is stratigraphically continuous with the underlying Evaporitic Unit and records the progradation of fluvio-deltaic systems into the evaporitic basin. It reflects a rapid continentalization of the basin during the early Stage 1 of the MSC (Lacerda-Orita et al., 2025).

3. Materials and methods

3.1. Fieldwork and sampling

Field campaigns were conducted to log and analyse the pre-evaporitic interval of the Yesón Alto composite section (37°02'40" N; 2°30'23" W), which has a total thickness of 80 m (Fig. 2). The study focused on facies analysis and sampling, with particular attention given to the Pre-Evaporitic Carbonates Unit. Within this unit, the geometry of limestone beds and their spatial relationships with surrounding fine-grained strata were characterized. Both carbonate deposits and interbedded shales and siltstones were sampled for petrographic, micropaleontological, and bulk-rock mineralogical analyses (XRD).

3.2. Mineralogy and petrology

Fifteen representative polished slabs of limestone facies were prepared at the thin section laboratory of the University of Alicante. The slabs were subsequently scanned to document rock textures, skeletal, and non-skeletal components. Fifteen standard thin sections were also produced and analysed under a ZeissTM Axiophot optical microscope for systematic petrographic characterization. Photomicrographs were taken

to document microfacies and diagenetic features. Carbonate samples ($n = 18$) were examined using SEM-EDS and XRD to determine their primary composition.

3.3. Micropaleontological Analysis

Shale and siltstone samples were collected at 1-m intervals to analyse their microfossil assemblages. Approximately 0.5 kg of sediment per sample was processed using wet sieving to extract the $>125 \mu\text{m}$ fraction. The sieved fraction was treated with hydrogen peroxide (H_2O_2) to remove organic matter and subsequently cleaned using an ultrasonic bath to eliminate clay particles. Micropaleontological analysis was conducted under a ZeissTM Discovery V12 binocular microscope.

Qualitative analysis of 72 samples was performed to characterize calcareous macro- and microfossils and the paragenetic relationships between their tests and diagenetic constituents. Taxonomic identification was mainly based on the studies of Van der Laan et al. (2005) and Corbí (2010), and a collection with the most representative taxa was prepared after picking. Based on this collection, well-preserved specimens were selected for examination and documentation at higher resolution using a Quanta 200 low-vacuum scanning electron microscope (SEM) at the Scientific and Technological Centers of the University of Barcelona (CCiTUB).

For semiquantitative analysis, 300 specimens of planktonic and benthic foraminifera were counted in 34 samples representative of the investigated interval to calculate the planktonic/benthic ratio. Additionally, 30 samples with a sufficient number of benthic foraminifera were selected for a more detailed semiquantitative analysis, in which 300 benthic specimens were identified and counted per sample. Relative abundances of key benthic taxa were determined relative to the total benthic assemblage per sample, consistent with the thresholds defined by Corbí et al. (2016): occasional presence ($<5\%$), common presence ($5\text{--}20\%$), abundant presence ($20\text{--}50\%$), and dominant presence ($>50\%$). Quantification was performed using a custom Python script developed for live species counting.

To reconstruct the palaeobathymetry, we considered the distribution and relative abundance of determined benthic taxa classified as water depth markers by Van der Laan et al. (2005). The estimated paleodepths were correlated to other paleontological and sedimentological data and attributed to different environmental settings following Corbí et al. (2020) and based on the bathymetric zonation defined by Wright (1978) for the Mediterranean. These settings are: (1) Shore zone (shallower parts of the littoral zone); (2) inner-shelf (from ~ 5 to ~ 50 m); (3) outer-shelf (from ~ 50 to $\sim 150\text{--}200$ m); (4) upper epibathyal (from $\sim 150\text{--}200$ to $\sim 500\text{--}700$ m).

4. Results

4.1. Yesón Alto: subsection A

Subsection A of the Yesón Alto composite section comprises a ~ 100 m-thick succession formed by three main lithostratigraphic units: The Laminated Siltstones and Shales Unit, the Pre-Evaporitic Carbonates Unit, and the overlying Evaporitic Unit (Fig. 2). The first two units record the pre-MSC stage in the Tabernas Basin, while the Evaporitic Unit corresponds to gypsum/shale deposits from Stage 1 of the MSC. This study focuses on the pre-MSC interval of the succession (i.e., The Laminated Siltstones and Shales Unit, the Pre-Evaporitic Carbonates Unit).

The bulk of the studied interval consists predominantly of fine-grained siliciclastic facies, including grey to orangish siltstones and shales that are commonly rich in calcareous microfossils. Medium- to very-fine-grained sandstone beds, up to 10 cm thick, are interbedded within the Laminated Siltstones and Shales Unit. The transition to the overlying Pre-Evaporitic Carbonates Unit is gradual and defined by the first appearance of a carbonate bed (a) at the upper part of the MMi 13d

subzone (Figs. 2 and 3A).

The Pre-Evaporitic Carbonates Unit is 29 m thick and consists of thirteen hybrid carbonate beds alternating with fine-grained facies. These hybrid carbonate beds exhibit ochre to yellowish colours, ranging from a few centimetres to over one meter in thickness, and form prominent and indurated layers with tabular, lenticular, or nodular geometries (Lacerda-Orita et al., 2025). Deformation features, including faulting and folding of beds, are particularly common towards the top of the unit, at the transition with the Evaporitic Unit (Fig. 3A). Notably, a soft-sediment deformation structure (SSDS) horizon marks the transition between the Pre-evaporitic Carbonates and Evaporitic Units (Fig. 3D).

4.2. Lithofacies

Seven facies and facies associations were determined based on sedimentary structures, primary texture, and skeletal and non-skeletal

components. Each facies association is described below and illustrated in Figs. 3–9.

4.2.1. Shales and siltstones

This facies association comprises light-grey shales and siltstones, which constitute the bulk of the Laminated Siltstones and Shales Unit, but also occur interbedded with hybrid carbonates within the Pre-Evaporitic Carbonates Unit (Figs. 2A and 3).

These shales and siltstones range from massive to finely laminated and are dominantly siliciclastic in composition, with muscovite, quartz, clinchlore, and paragonite as the principal mineral constituents. The facies association also includes minor amounts of very-fine to medium-sized sand, angular quartz and feldspar, platy-shaped muscovite, and schistose metamorphic fragments. These fine-grained facies contain abundant foraminifera and other marine fauna, which are documented in Section 4.3.

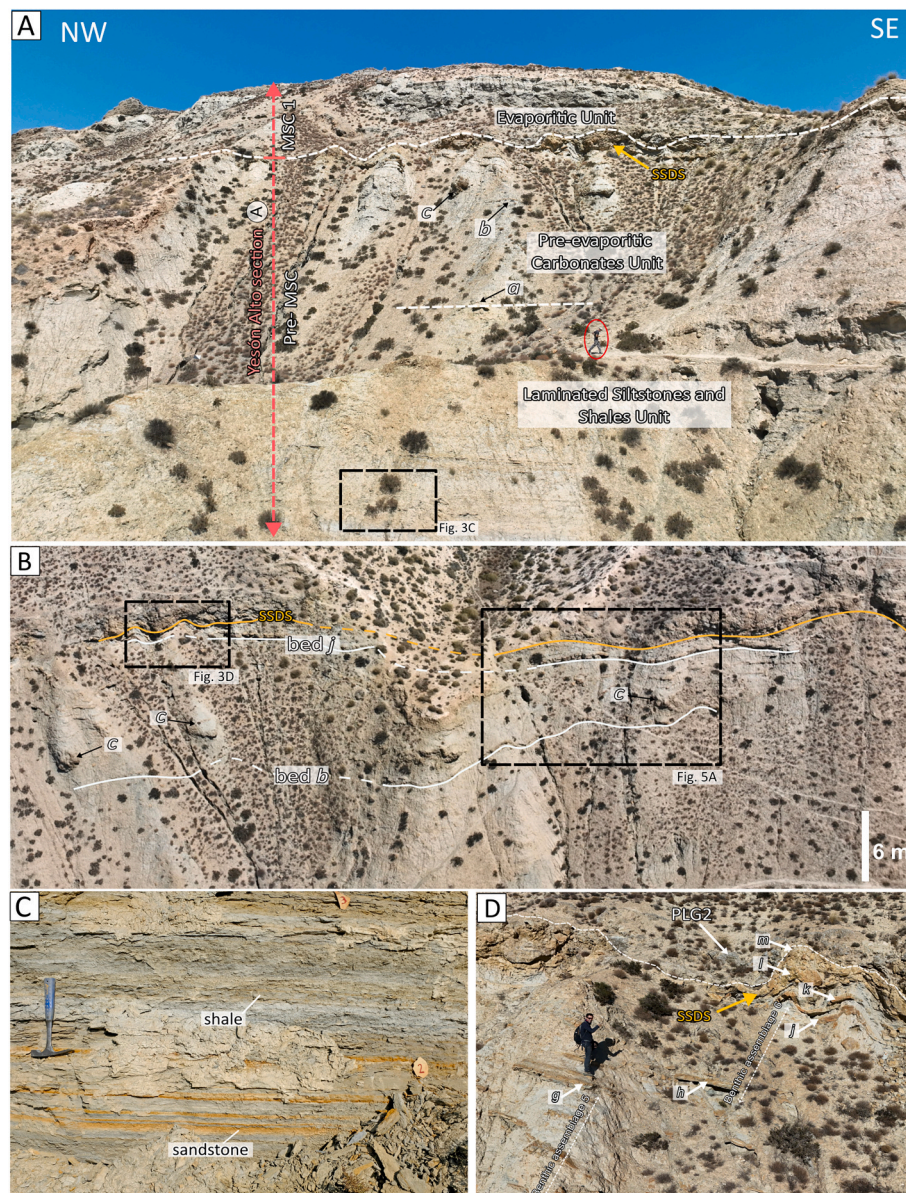


Fig. 3. A) Panoramic view of the Yesón Alto subsection A, showing its main stratigraphic units and features. Captions a, b, and c correspond to limestone facies referred below in the text. The red oval indicates a human for scale. B) Unmanned Aerial Vehicle (UAV) image of the upper part of the Yesón Alto subsection A, highlighting the lateral extension of carbonate beds and deformational structures. C) Typical facies association of the Laminated Siltstones and Shales Unit. D) Uppermost part of the subsection, showing the transition to the overlying Evaporitic Unit. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

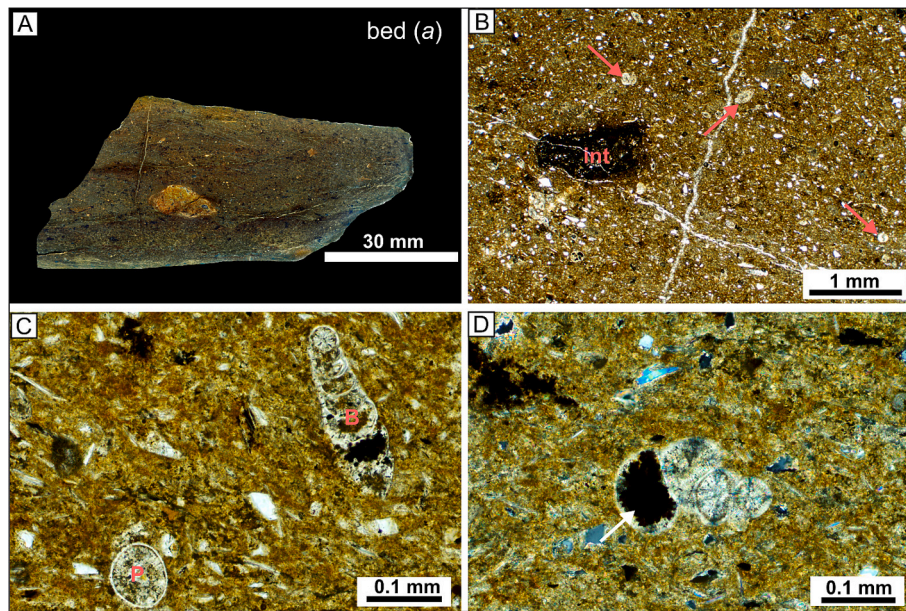


Fig. 4. Silty foraminiferal wackestone facies. A) Polished slab from bed (a). Red arrows show foraminiferal bioclasts. B) Microphotograph showing main textural and compositional features. Int = intraclast. Pink arrows point to foraminifera tests. C) Close-up of the skeletal particles, dispersed within a siliciclastic-influenced micritic matrix. B = benthic foraminifera test (likely *Bulimina* or *Uvigerina*), P = planktic foraminifera test (likely *Orbulina*). D) Microframboids of pyrite filling test chamber, highlighted by the white arrow. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

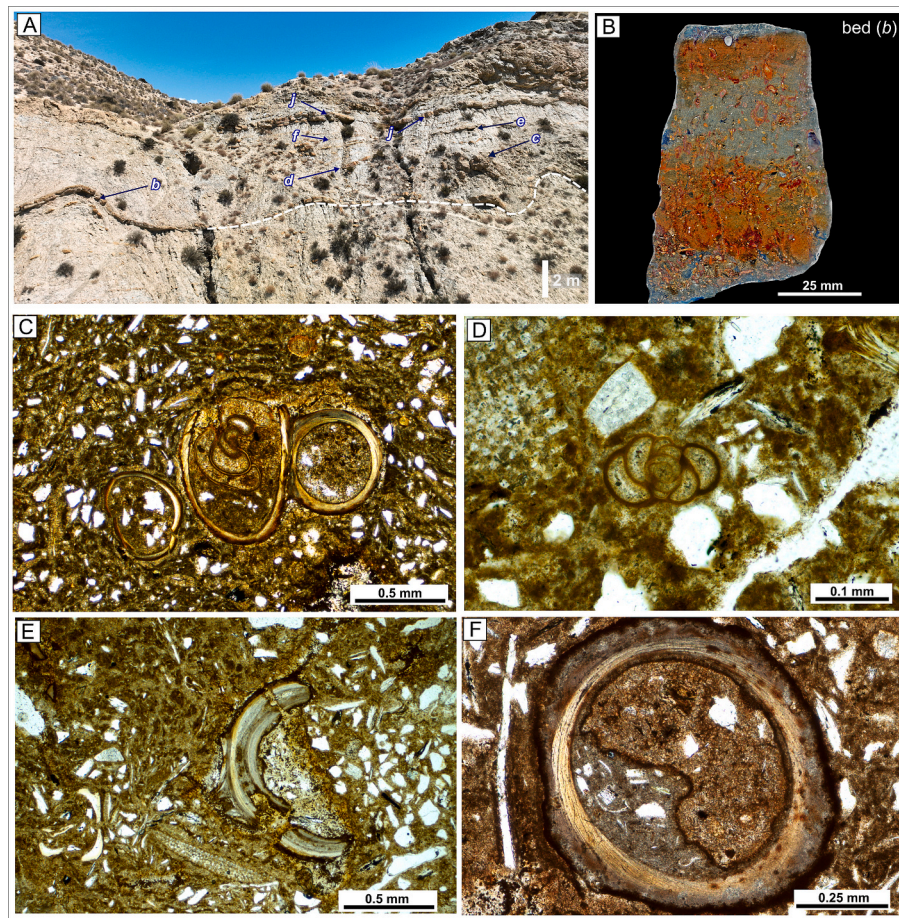


Fig. 5. Sandy bioclastic packstones. A) Outcrop image showing the relationships between different carbonate facies and bed deformation. B) Polished slab of bed (b). C) Serpulid bioclasts dispersed in mixed-siliciclastic matrix. D) Close-up view of a miliolid. E) Detail of a bivalve fragment. F) Serpulid exhibiting a micritized outer rim. Note the amount of detrital sand dispersed within the matrix.

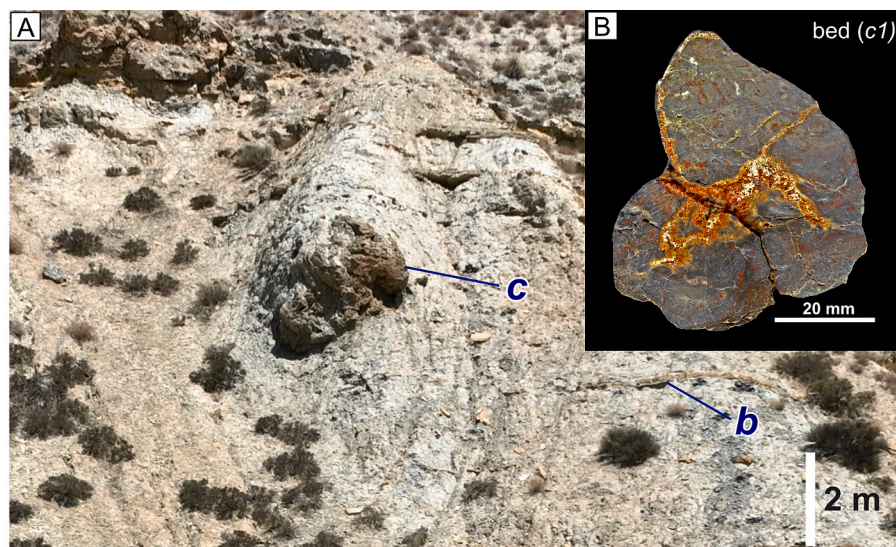


Fig. 6. Carbonate concretions. Unmanned aerial vehicle (UAV) image of outcrop, showing a large carbonate nodule overlying the tabular sandy bioclastic packstone bed *b*. B) Polished slab of carbonate nodule, exhibiting septarian cracks in the center.

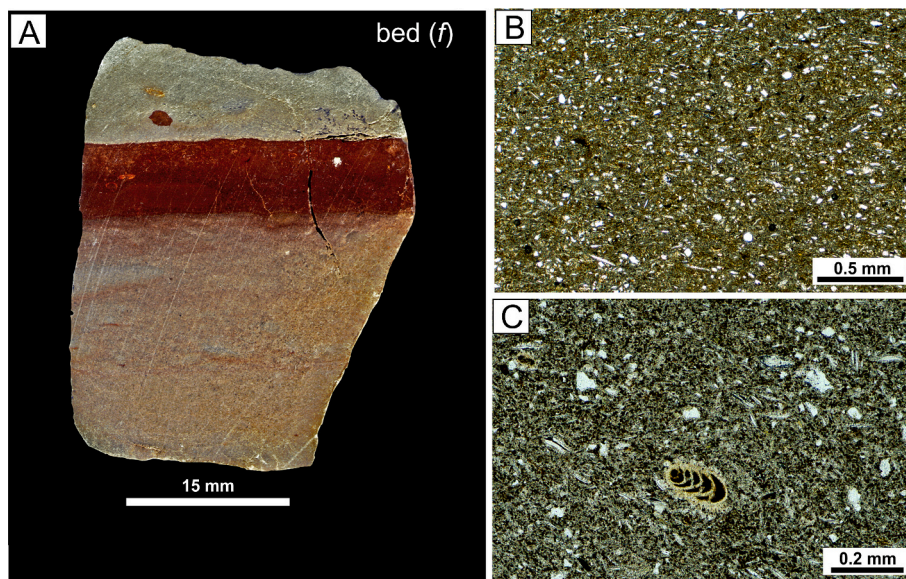


Fig. 7. Silty/sandy mudstones. A) Polished slab of a silty/sandy mudstone, characterized by layered Fe-oxide bands. B) Typical texture of the silty/sandy mudstone facies. C) Isolated foraminiferal tests dispersed within the sandy matrix.

4.2.2. Silty foraminiferal wackestones

This facies consists of greyish-brown to ochre mixed siliciclastic-influenced limestones (Fig. 4A), with beds typically ranging in thickness from 20 to 50 cm. It is represented by a bed (*a*), located at 40-m from the base of the investigated interval, which marks the base of the Pre-Evaporitic Carbonates unit (Figs. 2A, 3A, and 10). This bed exhibits a lenticular geometry and thins laterally across the outcrop. Its contacts with the surrounding shales are sharp at both the base and the top.

Bed (*a*) is massive and exhibits a wackestone texture, containing approximately 15% of bioclasts dispersed within a siliciclastic-influenced micritic matrix (Fig. 4B). The bioclasts are predominantly foraminifera, including both planktonic and benthic forms (Fig. 4C). XRD analysis indicates that the detrital fraction is dominated by silt-sized grains of quartz, feldspars, and muscovite (Table 1). Intra-basinal constituents, such as shale intraclasts, occur locally within the matrix.

The main diagenetic constituent is pyrite, which occurs as microcrystalline framboids that either replace or precipitate within the

micritic matrix or preferentially fill intraparticle pores within foraminifera shells (Fig. 4D).

4.2.3. Sandy bioclastic packstones

This facies is represented by bed (*b*), which is approximately 30 cm thick and located at 56-m from the base of the studied interval. This bed is tabular and exhibits lateral continuity across the outcrop, serving as a reliable marked bed (Fig. 3B). It is notably folded towards the south-eastern sector of the outcrop (Fig. 5A).

These hybrid carbonates are massive and exhibit a packstone texture, composed predominantly of densely packed bioclasts (Fig. 5B). These packstones have a more diverse assemblage of skeletal components in comparison to the silty wackestones, including serpulids (Fig. 5C), echinoids, bivalves, bryozoans, gastropods, and both benthic (Fig. 5D) and planktonic foraminifera. Some of the larger bioclasts, such as bivalves and serpulids, are often fragmented and exhibit evidence of reworking (Fig. 5E). Other minor intra-basinal constituents include

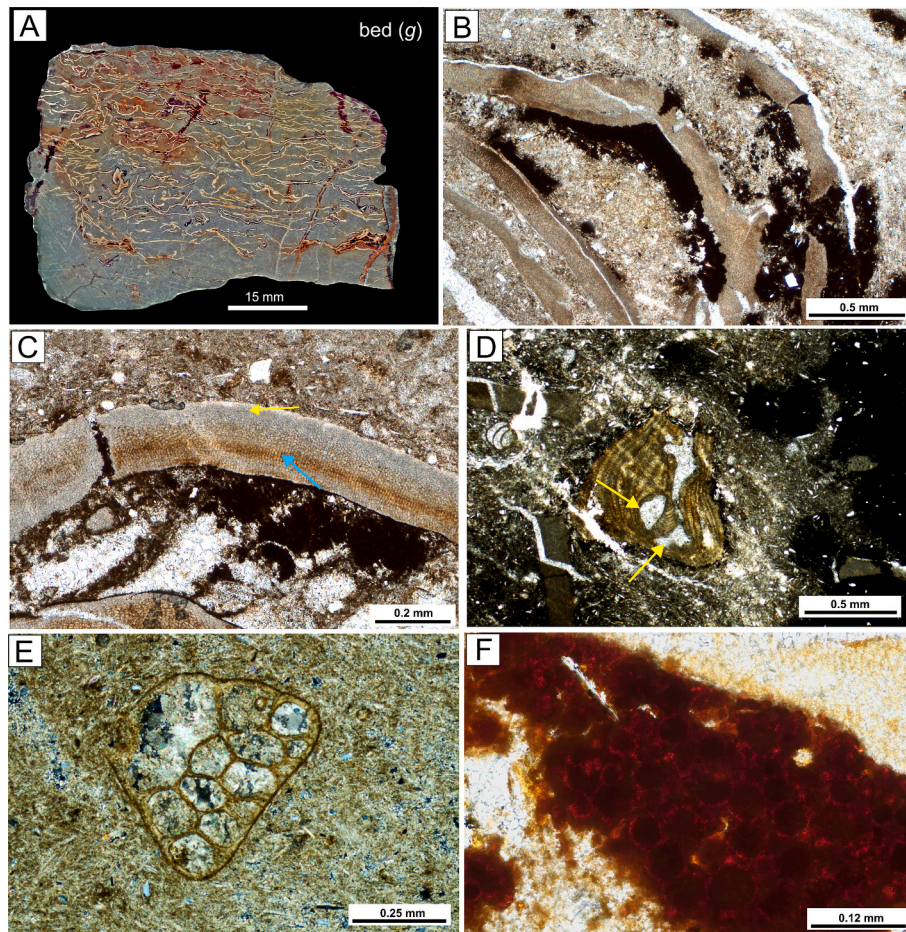


Fig. 8. Crustose coralline algal bindstones: A) Polished slab showing the framework structure built by crustose coralline algae. White arrows point to veins filled by sparry calcite. B) Microphotograph showing a detail of the coralline thalli constructing the frameworks. C) Detail of coralline thalli. Blue arrow = ventral core; Yellow arrow = peripheral filaments. D) Close-up view of sporangial conceptacles. Yellow arrows indicate calcite cements filling intraparticle pores. E) Detail of a fragment of a bryozoan filled by macrocrystalline calcite. F) Iron oxides replacing precursor mineral of roundish habit, likely corresponding to framboidal pyrite. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

limestone intraclasts.

The siliciclastic fraction comprises very-angular to angular silt- to medium-sized quartz (both mono- and polycrystalline), feldspars, muscovite, and foliated metamorphic rock fragments (Table 1). These grains are chaotically distributed within the micritic matrix. Bioturbation is occasionally present.

The dominant diagenetic processes that these sandy wackestones experienced include micritization of bioclasts, forming dark-coloured micrite envelopes around calcareous bioclasts (see Bathurst, 1966; Fig. 5F), and the replacement of the micritic matrix and carbonate particles by microcrystalline iron-oxides.

4.2.4. Carbonate nodules

This facies association is represented by spheroidal limestone nodules, up to 2.5 cm in diameter, which exhibit orangish to ochre colours, and occur interbedded within the surrounding shales and siltstones (Fig. 6A). Although their lateral distribution across the outcrop is variable, they predominantly occur within a stratigraphic horizon that is affected by slumps, approximately between 57 and 59 m from the base of the investigated interval, corresponding to bed c (Figs. 3B, 5A, 6A and 10).

The nodules are massive, displaying septarian cracks filled with sparry calcite (Fig. 6B). They have a mudstone texture with a very low detrital content, consisting of angular to sub-angular silt- to very fine sand-sized quartz, feldspars, muscovite, and metamorphic rock fragments, chaotically dispersed within the micritic matrix. Notably, no

bioclasts were observed.

4.2.5. Silty/sandy mudstones

This facies association consists of greyish-brown to ochre tabular to lenticular beds ranging in thickness from 20 to 50 cm. It is represented by beds d, e, f, and i, corresponding to the most common hybrid-carbonate facies in the upper part of the section (Figs. 5A and 10). The beds exhibit sharp lower and upper contacts with the surrounding shales and siltstones.

The hybrid-carbonate beds are either massive or show plane-parallel stratification often marked by centimetre-thick, reddish-coloured bands (Fig. 7A) and are frequently affected by micro-faulting and soft-sediment deformation. These facies contain abundant detrital grains, similar in composition to those described in previous facies, dispersed within the micritic matrix (Fig. 7B). Biogenic components are rare and include isolated foraminifera (Fig. 7C) or bivalve fragments dispersed within the matrix. Their abundance decreases noticeably towards the top of the succession. Other intrabasinal constituents, such as shale intraclasts, are rarely present. Diagenetic products in this facies association include calcite cements, framboidal pyrite, and iron-oxides.

4.2.6. Crustose coralline algal bindstones

This facies forms lenticular beds approximately 30 cm thick, represented by beds g and h (Figs. 3D and 10). The beds exhibit sharp contacts with the underlying and overlying shales and are folded due to the intense deformation affecting this interval.

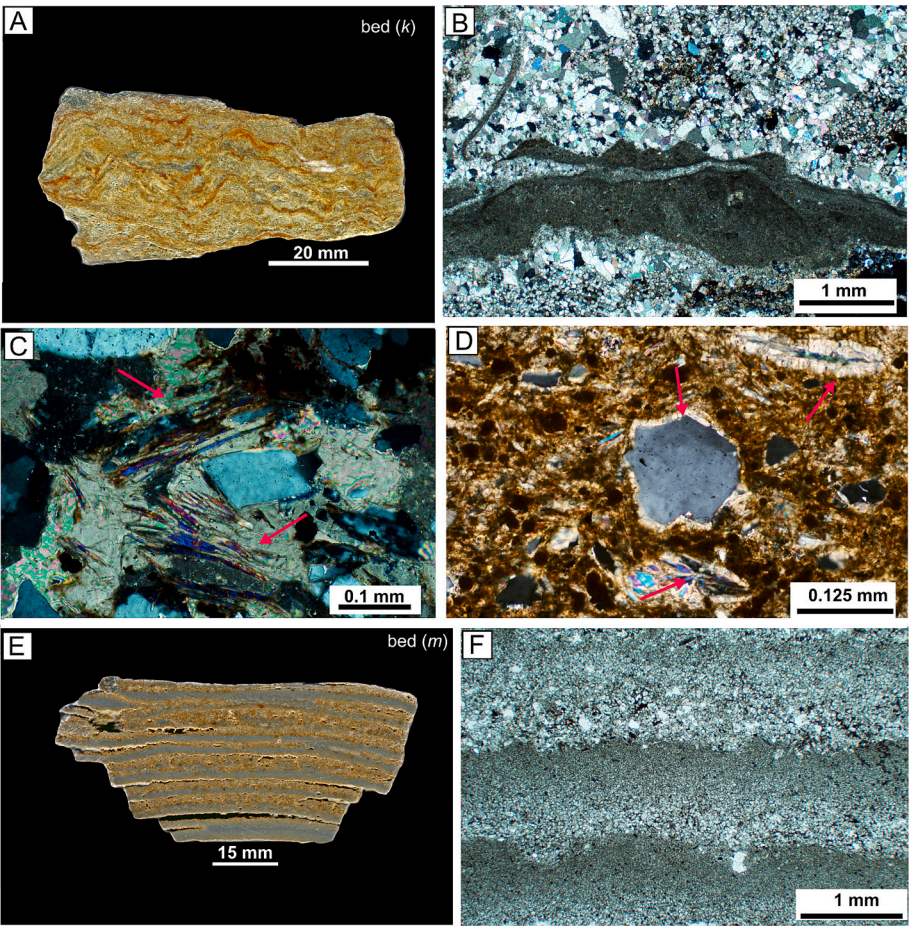


Fig. 9. Sandy laminated carbonates facies: A) Polished slab of a microbial layered structure showing a wrinkled lamination. B) Sparry calcite replacing original micritic-matrix. C) Mosaic calcite filling inter- and intraparticle pores. Note how calcite precipitated during early diagenesis is replacing and expanding muscovite (red arrows), indicating pre-compaction cementation. D) Calcite rims (red arrows) covering detrital grains. E) Polished slab of a microbial structure exhibiting a domal geometry and plane-parallel layering. F) Microphotograph of bed *m* displaying the lamination marked by the different granulometry, prior to extensive carbonate cementation. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1
Estimation of bulk mineral composition of carbonate facies from XRD data.

Sample	Facies assoc.	Quartz	Fe-Dolomite	Calcite	Muscovite	Gypsum	Dolomite	Clinochlore	Paragonite
2CB_1	(bed a) Silty bioclastic wackestones	10.9	5.2	69.8	7.1	1.9	0.9	3.2	1
2CB_2	(bed b) Sandy bioclastic packstones	14	0.5	76.4	6.3	0	0.1	1.7	1
2CB_3A	(bed c) Carbonate nodules	9.2	0.6	83.8	3.6	0.4	0.8	1.2	0.4
2CB_3B	(bed c) Carbonate nodules	13.4	0	64.4	8.4	9.9	0.1	2.7	1.1
2CB_4	(bed d) Silty/sandy mudstones	16.6	0	51.4	11.2	15.8	0.4	2.9	1.6
2CB_5	(bed e) Silty/sandy mudstones	13.8	0	64.6	9.1	7.9	0.1	3.5	1.1
2CB_5	(bed e) Silty/sandy mudstones	14.2	0	67.6	5.7	9.4	0	2.3	0.8
2CB_6	(bed f) Silty/sandy mudstones	16.1	0	64.5	6.9	7.9	0.3	3.4	0.8
2CB_7A	(bed g) Crustose coralline algal frameworks	16.1	0.6	62.3	11.8	0.8	0.3	6.2	1.7
2CB_7B	(bed g) Crustose coralline algal frameworks	8.5	0	74.5	8.4	4	0.3	3.1	1.2
2CB_8	(bed h) Crustose coralline algal frameworks	10.3	0.3	74.6	6.3	4.9	0.2	2.9	0.6
2CB_9.raw	(bed i) Silty/sandy mudstones	19.6	0	63.3	7.3	5.7	0.1	2.8	1
2CB_10A	(bed j) Sandy laminated carbonates	14.6	0.6	70.2	6.4	3.5	0.3	3.6	0.7
2CB_10B	(bed j) Sandy laminated carbonates	12.4	0.4	72.7	6.9	3.8	0.2	2.8	0.8
2CB_11	(bed k) Sandy laminated carbonates	18.9	0.3	44.4	11.8	18.8	0.2	4.4	1.3
2CB_12	(bed l) Sandy laminated carbonates	16.4	0.2	70.2	4.1	6.5	0.2	1.9	0.5
2CB_13A	(bed m) Sandy laminated carbonates	11.2	0.4	79.2	5.6	0.6	0.3	2.1	0.7
2CB_13B	(bed m) Sandy laminated carbonates	11.1	0.2	80.2	4	2.2	0.4	1.5	0.5

The algal frameworks consist of moderately to densely packed coralline algal crusts, exhibiting a sub-parallel orientation (Fig. 8A) within a micritic matrix (Fig. 8B). The crustose thalli exhibit a distinctive reticulate cellular microstructure, where a thicker ventral core (or hypothallium) and peripheral filaments can be distinguished (Fig. 8C).

The peripheral filaments are composed of a more densely packed cell arrangement, while the ventral core, in most cases, displays a less regular and generally non-coaxial arrangement (see e.g., Hrabovský et al., 2016). Sporangial conceptacles, which are reproductive structures, are also visible along the thalli of certain specimens (Fig. 8D). A slight

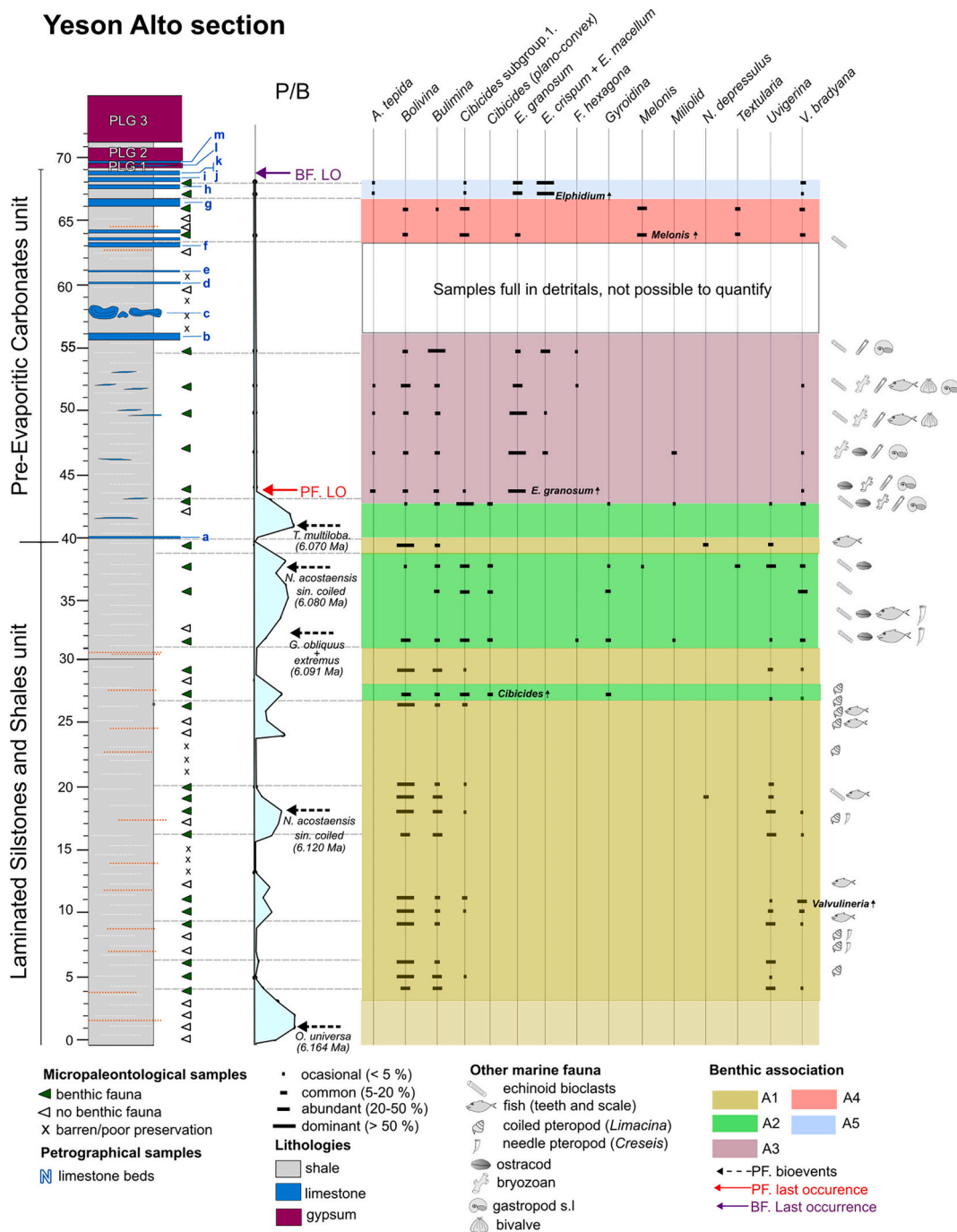


Fig. 10. Distribution of benthic foraminiferal assemblages and other main marine organisms. Key planktonic and benthic foraminifera bioevents are indicated in the P/B ratio column and are based on Lacerda-Orita et al. (2025).

upward increase in coralline algal crust density is observed in the polished slab of bed g shown in Fig. 8A.

Other skeletal components, including bryozoans, benthic foraminifera, and echinoids, are present in minor amounts. Silt- to very fine sand-sized siliciclastic grains are also present in the matrix. Intraparticle pores within bryozoan and foraminiferal tests are filled with microcrystalline calcite (Fig. 8E). Micrite envelopes are locally present around bioclast margins. Iron oxides occur as microcrystalline aggregates, replacing both the micritic matrix and a precursor mineral with rounded morphology, likely framboidal pyrite (Fig. 8F).

4.2.7. Sandy laminated carbonates

This facies association is represented by beds (j), (k), (l), and (m),

which occur in the uppermost part of the succession (Figs. 5A, 9, and 10). The geometry of these beds is difficult to define due to the intense deformation associated with soft-sediment deformation structures (SSDS) near the transition to the overlying Evaporitic Unit (Fig. 3B). Beds (j) and (k), located at the base of the SSDS interval, exhibit greater lateral continuity across the outcrop. In contrast, the contacts of beds l and m with the surrounding hybrid carbonates and the first gypsum bed are more diffuse and difficult to trace. This facies is characterized by wrinkled lamination or well-developed layering. Beds (j) and (k) exhibit millimetre-thick grey and yellowish wrinkled laminae (Fig. 9A). Under the microscope, this lamination is defined by alternating micritic and sand-rich layers. The sandy laminae are poorly sorted and composed of silt- to fine-grained sand, with angular to sub-angular grains mainly

oriented sub-parallel to the lamination. Calcite cementation occurs after the dissolution of the micritic matrix (Fig. 9B), or by replacing detrital grains, and expanding mica grains (Fig. 9C). It also forms fibrous rims coating sand grains (Fig. 9D).

Along the same lines, bed (m) exhibits plane-parallel lamination, with distinct alternating greyish and yellowish laminae (Fig. 9E). These laminae correspond to alternating greyish micritic laminae and yellowish sand-rich laminae, both extensively cemented by micro- and macrocrystalline calcite (Figs. 9F).

4.3. Micropaleontological analysis of foraminifera and associated fauna

Five benthic assemblages were identified based on the distribution of selected benthic foraminifera taxa and associated fossils with palaeoecological and depositional significance. Table 2 summarizes the main genera and species identified, while Fig. 10 displays the semi-quantitative data for benthic foraminifera.

Table 2

List of genera and species of the foraminifera identified in the pre-evaporitic succession of the Yesón Alto section. Some genera, such as *Cibicides* and *Elphidium*, were subdivided in the semiquantitative analysis due to the different paleoenvironmental significance of different species.

<i>Ammonia tepida</i> Cushman, 1926
<i>Bolivina alata</i> Seguenza, 1862
<i>Bolivina dentellata</i> Tavani, 1955
<i>Bolivina dilatata</i> Reuss, 1850
<i>Bolivina punctata</i> d'Orbigny, 1839
<i>Bolivina spathulata</i> Williamson, 1858
<i>Bulimina aculeata</i> d'Orbigny, 1826
<i>Bulimina striata</i> Guérin-Méneville, 1832
<i>Bulimina echinata</i> d'Orbigny, 1852
<i>Bulimina elongata</i> d'Orbigny, 1846
<i>Cibicides dutemplei</i> d'Orbigny, 1846
<i>Cibicides pseudoungerianus</i> Cushman, 1922
<i>Cibicides refulgens</i> Monfort, 1808
<i>Cibicides ungerianus</i> d'Orbigny, 1846
<i>Dentalina catenulata</i> Brady, 1884
<i>Elphidium complanatum</i> d'Orbigny, 1839
<i>Elphidium crispum</i> Linnaeus, 1758
<i>Elphidium granosum</i> d'Orbigny, 1846
<i>Elphidium macellum</i> Fichtel and Moll, 1798
<i>Favulina hexagona</i> Williamson, 1858
<i>Globigerina bulloides</i> D'Orbigny, 1826
<i>Globigerinella obesa</i> Bolli, 1957
<i>Globigerinita</i> sp. (genero) Brönnimann, 1951
<i>Globigerinoides bulloideus</i> Crescenti, 1966
<i>Globigerinoides extremus</i> Bolli and Bermúdez, 1965
<i>Globigerinoides obliquus</i> Bolli, 1957
<i>Globigerinoides trilobus</i> , Reuss, 1850
<i>Globigerinoides sacculifer</i> Brady, 1877
<i>Globobulimina affinis</i> d'Orbigny, 1839
<i>Gyroldina soldanii</i> d'Orbigny, 1826
<i>Gyroldina umbonatus</i> Silvestri, 1898
<i>Laevidentalina leguminiformis</i> Batsch, 1791
<i>Lenticulina calcar</i> Linnaeus, 1758
<i>Lenticulina cultrata</i> Monfort, 1808
<i>Lobatula lobatus</i> Walker and Jacob, 1798
<i>Marginulinopsis costata</i> (Batsch, 1791)
<i>Marginulina hirsuta</i> d'Orbigny, 1826
<i>Melonis pompilioides</i> Fichtel and Moll, 1798
<i>Neoglobobulimina acostaensis</i> (Blow, 1959)
<i>Nonion depressulus</i> Walker and Jacob, 1798
<i>Orbulina universa</i> D'Orbigny, 1839
<i>Orbulina suturalis</i> Brönnimann, 1951
<i>Planorbulina mediterranensis</i> d'Orbigny, 1826
<i>Reussella spinulosa</i> (Reuss, 1850)
<i>Textularia</i> sp. (genera) DeFrance, 1824
<i>Turborotalita multiloba</i> Romeo, 1965
<i>Turborotalita quinqueloba</i> Natland, 1938
<i>Uvigerina cylindrica</i> d'Orbigny, 1826
<i>Uvigerina peregrina</i> Cushman, 1923
<i>Uvigerina striatissima</i> Perconig, 1955
<i>Valvulineria bradyana</i> Fornasini, 1900
<i>Valvulineria complanata</i> d'Orbigny, 1846

4.3.1. Benthic assemblage 1

This assemblage is the most represented within the first 30 m of the Laminated siltstones and shales Unit (Fig. 10), and is characterized by a low diversity of species, dominantly the infaunal genera *Bolivina*, *Bulimina*, and *Uvigerina*. *Bolivina* is the dominant genus, and comprises species such as *Bolivina dentellata* (Tavani, 1955), *Bolivina dilatata* (Reus, 1850), *Bolivina punctata* (d'Orbigny, 1839), and *Bolivina spathulata* (Williamson, 1858) (Fig. 11). The *Bulimina* genus is mainly represented by *Bulimina echinata* (d'Orbigny, 1852) and *Bulimina aculeata* (d'Orbigny, 1826), with *Bulimina elongata* (d'Orbigny, 1846) occurring occasionally. The *Uvigerina* genus is represented by *Uvigerina cylindrica* (d'Orbigny, 1826) and *Uvigerina peregrina* (Cushman, 1923), while *Uvigerina striatissima* (Perconig, 1955) occurs in smaller amounts. *Valvulineria bradyana* (Fornasini, 1900), a juvenile form of *Valvulineria complanata* (d'Orbigny, 1846), also occurs, but rarely exceeds more than 5% of the assemblage. Less common but paleoecologically important taxa include *Globobulimina affinis* (d'Orbigny, 1839), and *Cibicides* sub-group 1, formed by *Cibicides dutemplei* (d'Orbigny, 1846), *Cibicides*

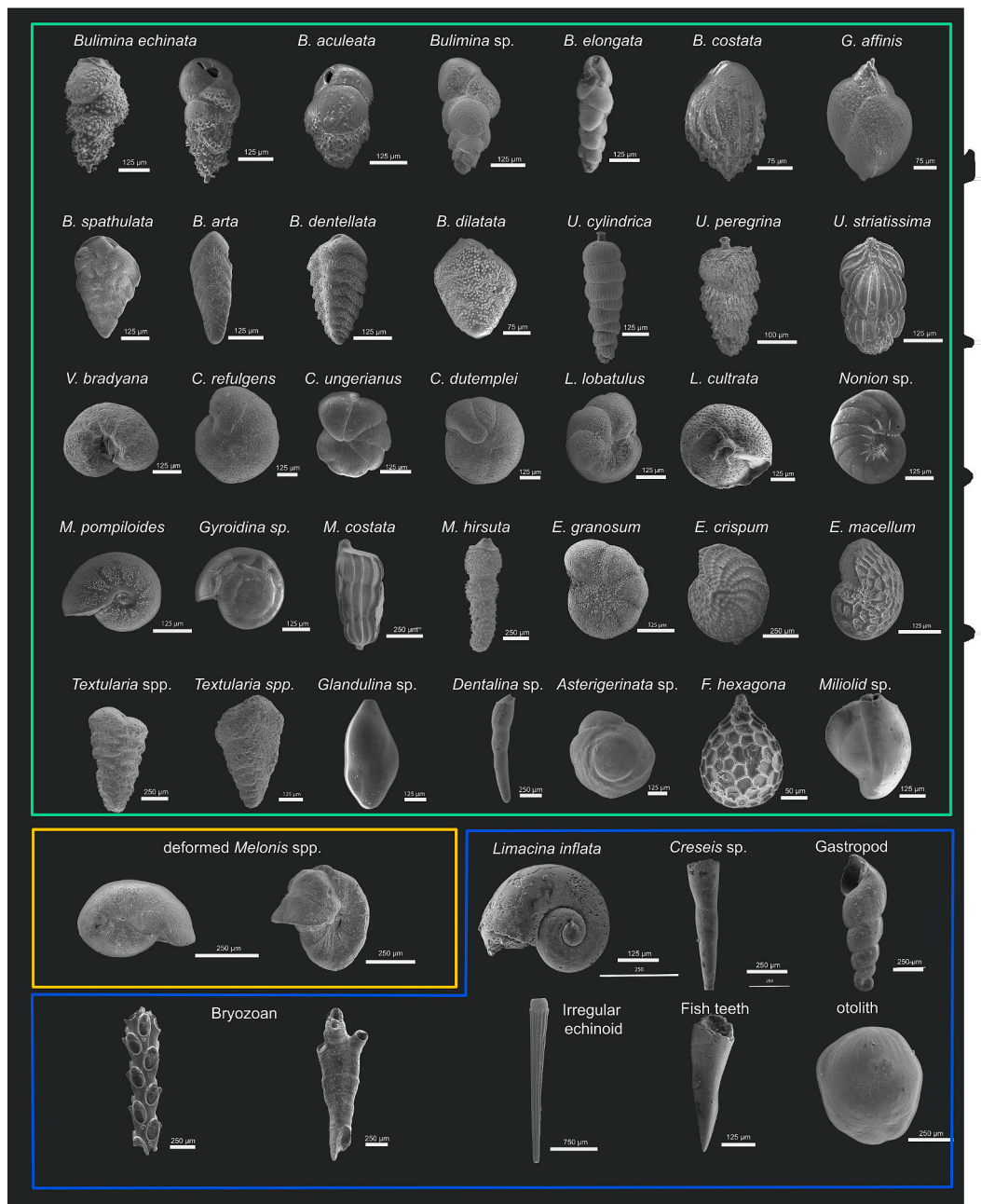


Fig. 11. SEM images of the main fossils identified in the Yesón Alto section. Green square includes the main studied benthic foraminifera. Yellow rectangle highlights deformed tests of *Melonis* spp. from assemblage 5. Blue area illustrates other important marine bioclasts. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

ungerianus (d'Orbigny, 1846), and *Cibicides pseudoungerianus* (Cushman, 1922). Also, two notable influxes of *Nonion depressulus* (Walker and Jacob, 1798), forming >15% of the benthic assemblage, occur at the 29 and 49 m mark (Fig. 10). Other bioclasts, such as fish teeth, scales, and otoliths, were also observed (Fig. 11).

The upper limit of the benthic assemblage 1 is at around 39 m from the base, immediately below the first silty foraminiferal wackestone of bed *a*, marking the transition with the overlying Pre-Evaporitic Carbonates Unit (Fig. 10). In the uppermost sample characterised by benthic assemblage 1 interval, specimens of *Uvigerina* and *Bulimina* notably exhibit more elongated tests, *U. cylindrica* is dominant within the genus, and *B. aculeata* exhibits strongly inflated upper chambers (Fig. 11).

Planktonic foraminifera (PF) are abundant throughout the

stratigraphic interval characterised by benthic assemblage 1, comprising up to 99% of the total foraminiferal assemblage in some samples (Fig. 10). These PF taxa were previously described by Lacerda-Orita et al. (2025) and correspond to two different ecological groups. The first group includes *Orbulina universa* (d'Orbigny, 1839), *Orbulina suturalis* (Brönnimann, 1951), *Globigerinoides obliquus* (Bolli, 1957), *Globigerinoides bulloideus* (Crescenti, 1966), *Globigerinoides extremus* (Bolli and Bermúdez, 1965), *Globigerinoides sacculifer* (Brady, 1877), *Globigerinoides trilobus*, (Reus, 1850), and *Globigerinella obesa* (Bolli, 1957). The second comprises *Neoglobobulimina acostaensis* (Blow, 1859), *Turborotalita quinqueloba* (Natland, 1938), *Turborotalita multiloba* (Romeo, 1965), *Globigerina bulloides* (d'Orbigny, 1826), and *Globigerinita* sp. (genus, Brönnimann, 1951).

Pteropods, which are a type of holoplanktonic gastropod, are also

common in the interval. They are usually preserved as shell moulds infilled with framboidal pyrite (see Section 4.4). Two taxa are identified: the sinistrally coiled *Limacina inflata* (d'Orbigny, 1834) and *Creseis* (Rang, 1828), distinguished by its elongated, needle-shaped shells (Fig. 10).

4.3.2. Benthic assemblage 2

This assemblage dominates in the upper part of the Laminated siltstones and shales Unit (between ~27 and 43 m from the base of the studied interval) and is characterised by an increase in species diversity compared to benthic assemblage 1. The most common genera include *Cibicides*, *Lobatula*, *Gyroidina*, *Valvulinaria*, *Bolivina*, *Bulimina*, and *Uvigerina*.

Within these, *Cibicides* subgroup 1 (*C. dutemplei*, *C. ungerianus*, and *C. pseudoungerianus*) is very abundant, followed by plano-convex *cibicides*, including *Lobatula lobatulus* (Walker and Jacob, 1798) and *Cibicides refulgens* (Monfort, 1808). *Gyroidina soldani* (d'Orbigny, 1826) and *Gyroidina umbonatus* (Silvestri, 1989), and *V. bradyana* are also common and occur in similar proportions. *B. aculeata* is the most common species of the *Bulimina* genus, though *B. echinata* and *Bulimina striata* (Guérin-Méneville, 1832) are also present. Benthic assemblage 2 further includes the occasional presence of *Lenticulina calcar* (Linnaeus, 1758), *Lenticulina cultrata* (Monfort, 1808) *Textularia* spp. (genus, DeFrance, 1824), *Dentalina catenulata* (Brady, 1884), *Laevidentalina leguminiformis* (Batsch, 1791), *Marginulina hirsuta* (d'Orbigny, 1826), *Marginulinopsis costata* (Batsch, 1791), *Reussella spinulosa* (Reuss, 1850), and *Planorbilina mediterraneensis* (d'Orbigny, 1826).

Uvigerina, notably *U. peregrina*, are restricted to a single stratigraphic level at 37 m (Fig. 10), forming approximately 30% of the benthic assemblage. This level also contains *Melonis pompilioides* (Fichtel and Moll, 1798), which accounts for approximately 5% of the assemblage.

Planktonic foraminifera present in these samples are *Orbulina* spp., *Globigerinoides* spp., *G. bulloides*, *Globoturborotalita* sp. and neogloboquadrinids. Moulds of pteropods from the genus *Creseis* were also observed. Other bioclasts include spines of echinoids, ostracods, fish teeth, and otoliths (Fig. 11).

4.3.3. Benthic assemblage 3

This assemblage characterises the interval comprised between 43 and 56 m from the base (i.e., the lower half of the Pre-evaporitic Carbonates Unit), and its presence also coincides with the disappearance of planktonic foraminifera from the Yesón Alto section (Fig. 10). The most representative benthic genera in this assemblage are *Elphidium*, *Bolivina*, *Bulimina*, *Valvulinaria*, *Cibicides*, *Ammonia*, as well as miliolids. *Elphidium* occurs for the first time in the succession, with *Elphidium granosum* (d'Orbigny, 1846) being the dominant taxon in most samples. Other species within the genus, such as *Elphidium crispum* (Linnaeus, 1758), and *Elphidium macellum* (Fichtel and Moll, 1798), are present in smaller proportions. *Bolivina* is mainly represented by *Bolivina alata* (Seguenza, 1862), while *B. aculeata* is the most common species within the *Bulimina* genus. *Ammonia tepida* (Cushman, 1926) is also occasionally present.

The basal sample of the stratigraphic interval including benthic assemblage 3, is notably dominated by *C. dutemplei*, comprising up to 76% of the assemblage, although its abundance progressively decreases upwards. *V. bradyana* is also common at the base of the interval.

This assemblage is also characterized by a diversification of other carbonate-producing fauna, including bryozoans, echinoid spines, ostracods, serpulids, bivalves, and gastropods (Fig. 10). Bryozoans are particularly abundant and dominate the biogenic fraction in some samples. These bioclasts are normally cylindrical and whitish, and exhibit branching morphologies, with some specimens exhibiting well-preserved zoecial apertures (Fig. 11). Gastropods are approximately ~1 mm in size and are often found fragmented. Serpulids are attributed to the genus *Ditrupa* (Berkeley, 1835). Bivalve shells are normally fragmented, which masks their taxonomic identification.

Some intervals, particularly the one immediately below carbonate

bed *b*, exhibit a high proportion of sand, reaching up to 70%. The elevated detrital content suggests that this interval might result from a downslope transport and that a portion of the biogenic assemblage might be allochthonous, potentially compromising the paleoenvironmental interpretation of the uppermost interval of benthic assemblage 3

4.3.4. Benthic assemblage 4

Benthic assemblage 4 is characterized by a moderately diverse benthic association, dominated by *Melonis*, *Cibicides*, *Textularia*, *Valvulinaria*, *Elphidium*, and miliolids. This assemblage is found after an interval containing barren layers or with limited fossils (between 63 and 66 m from the base), corresponding to the upper part of the Pre-Evaporitic Carbonate Unit (shales interbedded with the silty/sandy mudstone facies) (Fig. 3D). Its upper boundary is located immediately below the coralline algal frameworks of bed *g* (Fig. 10).

Melonis pompilioides is the most representative species of benthic assemblage 4, and many specimens exhibit deformed and abnormal tests, particularly affecting the final chambers (Fig. 11). Within the *Cibicides* genus, *C. dutemplei* and *C. ungerianus* are the most representative species. *E. granosum*, *V. bradyana*, and *Textularia* sp. occur in similar proportions. *Textularia* sp. In addition to foraminifera, benthic assemblage 4 also contains other marine fossils, such as irregular echinoid spines. Very fine to fine sand is common in some samples.

4.3.5. Benthic assemblage 5

This assemblage flourished in the uppermost part of the pre-evaporitic interval (Fig. 3A), within the shales interbedded between the crustose algal frameworks (beds *g* and *h*) and the stromatolitic structures (bed *i*) (Fig. 10). Further above, benthic foraminifera disappear from the Yesón Alto section, preceding the deposition of the first gypsum bed.

Benthic assemblage 5 records a reduction in diversity compared to assemblage 4 and is characterized by the absence of other marine microfossils. It is notably dominated by the *Elphidium* genus, which comprises over 90% of the assemblage. *E. crispum* and *E. macellum* are the most abundant species, but *E. granosum* is also common. Other benthic foraminifera, including *C. dutemplei*, *A. tepida*, and *V. bradyana*, are occasionally present. Samples from this assemblage contain a considerable amount of angular fine-grained sand and carbonaceous fragments.

4.4. Pyrite authigenesis and paragenetic relationships

Pyrite is a common diagenetic constituent in the studied interval, primarily displaying framboidal and euhedral habits, and scanning electron microscopy (SEM) analysis reveals its association with benthic and planktonic foraminifera, pteropods, and other bioclasts tests. In more elongated tests of benthic foraminifera such as *Uvigerina* spp. and *Bulimina* spp., framboidal pyrite aggregates commonly occupy intra-shell pores (Fig. 12A), often preserving portions of the original calcitic test. Some tests also exhibit evidence of partial dissolution (Fig. 12B). Within the intraparticle spaces, pyrite is present both as microframboids and as slightly larger subhedral crystals (Fig. 12C).

In planktonic foraminiferal tests such as *Orbulina* spp., pyrite normally occurs as well-rounded, polyframboidal aggregates reaching up to 50 µm in diameter (Fig. 12D), as well as smaller, more subhedral framboids averaging ~15 µm (Fig. 12E). Framboidal pyrite is also found as isolated, irregular clusters dispersed within the matrix (Fig. 12F). At low degrees of pyritization, framboids tend to precipitate preferentially along the outer margins of foraminiferal tests (Fig. 12G). In contrast, phosphatic bioclasts such as fish teeth and scales typically show no evidence of pyritization.

In aragonitic tests of pteropods, pyrite occurs as framboids or rosette-like aggregates of euhedral crystals (Fig. 12H), infilling the intra-shell pore and completely replacing the original carbonate skeleton. These relationships preserve the external morphology of the test, forming

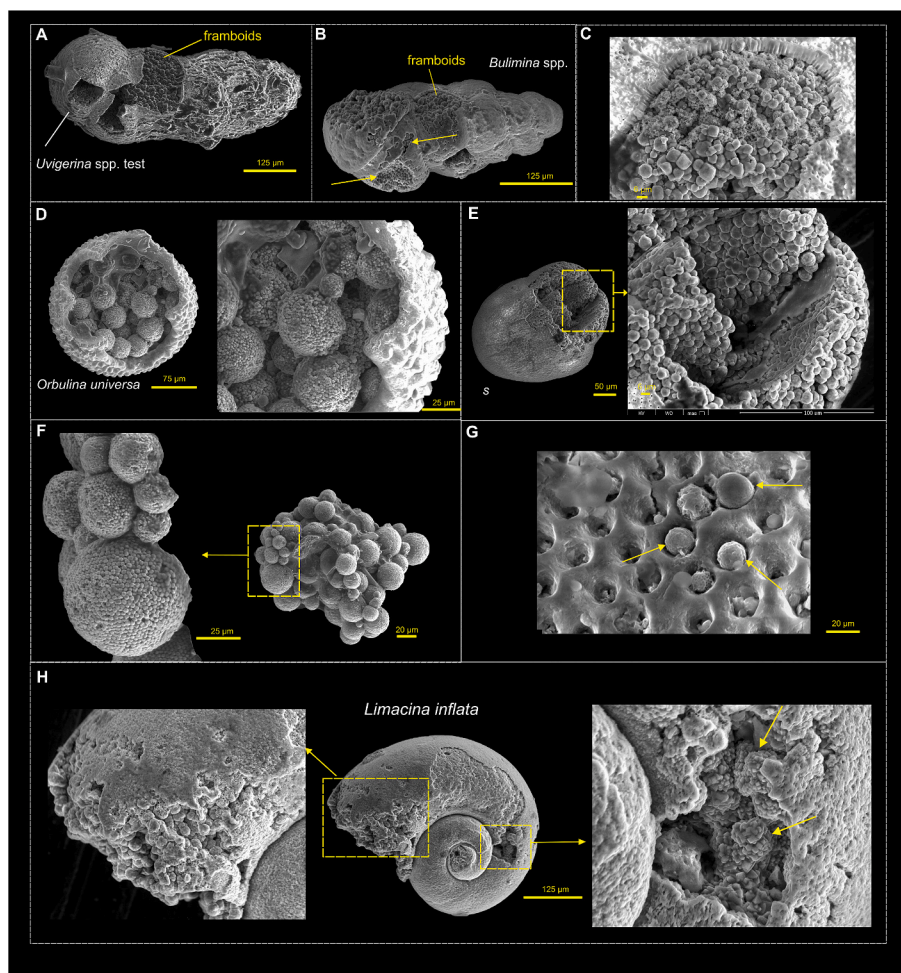


Fig. 12. Paragenetic relationships between pyrite and calcareous tests. A and B) Partially dissolved benthic foraminiferal tests filled by framboidal pyrite. C) Close-up of pore-filling pyrite showing framboidal and euhedral habits. D) Homogeneous polyframboids in planktonic foraminifera. F) Isolated aggregates of polyframboids. G) Pyrite polyframboids precipitated in the outer part of a planktonic foraminifera. H) Pteropod moulds fully replaced by framboidal and euhedral “rosette”-like pyrite.

detailed moulds that enable taxonomic identification despite the loss of original mineralogy (Fig. 12H).

Pyrite is more abundant in the lower part of the studied succession, particularly in samples associated with benthic assemblages 1 and 2. In contrast, it is less common in the interbedded shales of the Pre-evaporitic Carbonates Unit, where it is mainly concentrated in the lower part of the unit. In the silty foraminiferal wackestone bed (a), the intraparticle pore-filling relationships described in this section can also be identified (Fig. 4C and D). In other carbonate facies, pyrite normally occurs as isolated framboidal aggregates, either replacing the micritic matrix or partially infilling foraminiferal tests.

5. Discussion

5.1. Paleoenvironmental and sedimentary evolution of the Pre-MSC interval in the Tabernas Basin

A detailed micropaleontological analysis of the fine-grained facies enables the reconstruction of the evolution of the depositional environments and the major paleoenvironmental changes recorded during the pre-MSC stage in the Yesón Alto area. This interval was dated between 6.184 and 5.97 Ma (Lacerda-Orita et al., 2025) and cyclostratigraphically correlated with the reference section of the Sorbas Basin, as originally described by Sierro et al. (2001). Accordingly, the studied strata are interpreted within the context of progressively increasing

restriction across the Western Mediterranean, corresponding to the “consolidation phase” (sensu Corbí et al., 2020) of marine restriction of the Mediterranean.

Lacerda-Orita et al. (2025) interpreted that the Laminated Siltstones and Shales Unit (Fig. 2) was deposited in a pelagic environment intermittently disrupted by turbidity currents. The observed oscillations in the planktonic/benthic foraminiferal (P/B) ratio (Fig. 10) suggest the influence of precession-driven climatic forcing, which affected marine assemblages by controlling the conditions of the water column during an interval of limited water exchange with the Atlantic (Sierro et al., 2003). This led to episodes of bottom-water anoxia and stressed upper waters in the Mediterranean basins during the pre-MSC stage (see e.g., Blanc-Valleron et al., 2002; Sierro et al., 2003; Kouwenhoven et al., 2006). These cyclic fluctuations are consistent with paleoenvironmental signals documented in other Mediterranean basins during this interval (e.g., Sierro et al., 2003; Gennari et al., 2018). Nevertheless, caution should be taken when interpreting the variability in the P/B ratio obtained, as it may be biased by the influence of turbidite sedimentation (e.g., Corbí et al., 2020), oxygen depletion (Van Hinsbergen et al., 2005), and selective dissolution of juvenile planktonic foraminifera (Van der Zwaan et al., 1990).

Based on the data of the current study, benthic assemblage 1 (which characterises the interval between 0 and 30 m from the base of the studied portion of the succession), displays abundant *Uvigerina*, associated with *B. striata*, *C. ungerianus*, and *C. pseudoungerianus*, and a high P/

B ratio. This points towards an estimated depth of around 200–500 m (Van der Laan et al., 2005), corresponding to an upper-epibathyal setting. The occasional presence of shallower settings taxa, such as *C. dutemplei* (Fig. 10) may result from a minor influence of downslope transport by turbidity currents (e.g., Ortiz et al., 2011).

The abundance of stress markers in this assemblage (Van Hinsbergen et al., 2005) suggests persistent stressed conditions at the seafloor of the Tabernas Basin around 6.164 Ma (Fig. 10). *B. aculeata* is indicative of highly eutrophic, dysoxic settings (Fontanier et al., 2002). *B. echinata* is known to tolerate low oxygen and elevated organic matter content (Kouwenhoven and Van der Zwaan, 2006). Similarly, *G. affinis* is a species well-adapted to survive in nearly anoxic conditions (Corliss, 1985; Corliss and Emerson, 1990).

The co-existence of planktonic foraminifera and pteropods with dysoxic to suboxic taxa from the benthic assemblage 1 indicates a stratified basin, where oxic conditions prevailed in the upper water column. Most pteropod species have a narrow tolerance for temperature and salinity oscillations, being typical of tropical warm waters, and species such as *Limacina inflata* and *Creseis acicula* (Rang, 1828) (Fig. 11) prefer stable, warm, normal-salinity conditions (Herman, 1998).

The establishment of benthic assemblage 2 records a shallowing-upward trend from an upper-epibathyal setting towards outer-shelf conditions. The transition is characterized by an increase in species that can occur in outer-shelves, such as *C. dutemplei* (Van Hinsbergen et al., 2005), that along with an increase in benthic bioclastic material (chiefly echinoids and ostracods), suggests outer shelf conditions. The occasional presence of *Uvigerina* spp., *Gyroidina* spp., *Lenticulina* spp., and *M. pompilioides* indicates a bathyal influence in a predominant outer-shelf setting (Van Hinsbergen et al., 2005; Corbí, 2010; e.g., Corbí et al., 2016). The increase in the diversity of benthic species suggests enhanced bottom-water oxygenation, but the presence of stress-tolerant taxa such as *Bolivina* spp., *Bulimina* spp., and *V. bradyana* points to sustained environmental stress. At 39 m from the base of the investigated interval, the reappearance of species from benthic assemblage 1 suggests a decrease in oxygen levels at the seafloor or a short deepening phase preceding the deposition of the first carbonate bed.

The onset of carbonate deposition in the Yesón Alto section started between 6.080 and 6.070 Ma (Fig. 10), reflecting the regional expression of the shallowing-upward trend recorded in the uppermost pre-MSC stage in other basins of the Betic Cordillera, including the Sorbas (Goubert et al., 2001) and Bajo Segura (Corbí et al., 2016, 2020) basins. This trend is likely influenced by a global sea level drop of around 60 m, reported after 6.35 Ma and prior to the onset of the MSC, and associated with the beginning of a Northern Hemisphere glaciation (Hodell et al., 2001; Van der Laan et al., 2005; Holbourn et al., 2018; Jiménez-Moreno et al., 2013; Pérez-Asensio et al., 2013; Ohneiser et al., 2015).

Bed (a) is the lowermost carbonate bed of the Yesón Alto section (Figs. 3A and 10), deposited between the last occurrences of sinistrally coiled neogloboquadrinids and *Turburotalia multiloba*, dated at 6.080 (Sierro et al., 2001; Lourens et al., 2004) and 6.070 Ma (Blanc-Valleron et al., 2002; Lirer et al., 2019; Gennari et al., 2024), respectively (Fig. 10). The predominance of benthic and planktonic foraminiferal bioclasts dispersed within the micritic matrix in this silty foraminiferal wackestone facies (Fig. 4A–C), is suggestive of relatively low hydrodynamic energy in the outer-shelf setting. The occasional occurrence of limestone intraclasts suggests episodic reworking by potential contour currents or, internal wave action, storm waves or minor downslope transport. The abundance of terrigenous silt indicates a persistent siliclastic input into the basin from the adjacent emerged Sierra de Los Filabres (Fig. 1).

The presence of benthic assemblage 3 in the interval between carbonate beds (a) and (b) (43–56 m from the base) reflects further shallowing, as suggested by the disappearance of planktonic foraminifera (Fig. 10). The dominance of *Elphidium granosum* and *E. crispum*, associated with the notable increase in bryozoans, serpulids, bivalves, ostracods, gastropods, and irregular echinoids suggest a transition to shelfal

settings (possibly middle to inner shelf context, around 50 m of water depth). Oscillations in salinity are not recognized in this interval. This observation is reinforced by the abundance of irregular echinoids, which are highly sensitive to salinity variations and tolerate only limited ranges (a few ppm; Néraudeau et al., 2001). The diversity of irregular echinoids increased in the Sorbas Basin from the upper Tortonian to the uppermost pre-MSC layers, prior to their decline during the MSC (Néraudeau et al., 2001). Yet, restricted conditions are suggested by the predominance of the euryhaline *E. granosum* (Debenay et al., 2000), and by stress markers such as *B. aculeata*, *B. echinata*, and *V. bradyana*, which usually indicate intermittent dysoxia in bottom waters (Corbí et al., 2016).

The sandy bioclastic packstone facies (bed b) likely deposited in the inner shelf, where fauna typical of shallow marine settings, such as abundant miliolids (Fig. 5D), serpulids (Fig. 5C), and gastropods thrived. Also, the presence of very fine- to medium-grained quartz, feldspar, muscovite, and metamorphic rock fragments (Fig. 5C) indicates continuous detrital input from nearby terrestrial sources. The presence of micritic and mixed carbonate-siliclastic intraclasts indicates remobilization by wave action, yet their angular shapes, together with the low degree of bioclast fragmentation, suggest a limited degree of transport and reworking. The formation of micrite envelopes around skeletal grains (Fig. 5F) is interpreted as a product of bioerosion by microboring organisms (Bathurst, 1966), a process that typically occurs in shallow-water environments (Swinchatt, 1969; Bernard-Dumanois and Delancey, 1983). The tabular geometry and lateral continuity of bed b (Figs. 3B and 5A) support its deposition within a well-developed inner shelf environment, likely associated with a more extensive and productive carbonate factory.

The occurrence of carbonate nodules (Fig. 6A) with septarian cracks (Fig. 6B) is likely related to the influence of syn-sedimentary tectonics or early diagenetic processes in a muddy substrate (Raiswell and Fisher, 2000). The origin of septarian cracks remains debated, and proposed processes for their formation include in situ deformation (Marshall and Pirrie, 2013) and syneresis following bacterial extracellular polymeric substances (EPS) decay in muddy sediments (Hendry et al., 2006). Pratt (2001) proposed a model for soft-sediment deformation generated by earthquakes, which promotes the internal shrinking of muddy sediments, suggesting that septarian cracks may represent valuable indicators of syn-sedimentary tectonics in individual basins (Pratt, 2001). The intense deformation observed in the Yesón Alto section (Figs. 3B and 5A), associated with faulting, folding, slumps, and SSDS (Fig. 3D), particularly in the interval corresponding to bed c (Fig. 5A and 6A), may indicate that syn-sedimentary deformation influenced the deposition of the carbonate nodules. Moreover, the intense deformation of silty mudstone beds in the interval may also have generated carbonate bodies resembling ovoid concretions.

The lack of reliable micropaleontological data from the interval comprised between 57 and 63 m from the base of the section limits depth estimation. Petrological analysis from the silty/sandy mudstones (beds d, e, and f; Section 4.2.5) suggests deposition under moderate- to low-energy conditions in a shelfal setting, with a continuous supply of terrigenous sediments from emerged land. The formation of reddish bands (Fig. 7A) is interpreted as the result of meteoric water percolation during telodiagenesis, which promoted the oxidation of Fe-rich diagenetic constituents such as pyrite.

The establishment of benthic assemblage 4 (Fig. 10) records a brief deepening phase in the uppermost pre-MSC interval of the Yesón Alto section (from 63 to 66 m from the base). The dominance of *M. pompilioides*, *C. dutemplei*, and *C. ungerianus*, which usually dominate in outer shelfal to bathyal settings, suggest a shift towards deeper conditions (Van Hinsbergen et al., 2005; Corbí et al., 2020). In this setting, crustose algal frameworks (beds f and g) developed over a lime-mud soft substrate (e.g., Ingrassia et al., 2019). Accordingly, a reduction in siliclastic influx, and/or the presence of a deeper and calmer environment, was likely necessary to allow the encrustation of soft substrates by

coralline algae (Fig. 8A). However, the occurrence of fractured and randomly oriented thalli (Fig. 8B) suggests some degree of reworking and destabilization of the original framework.

Coralline algae are common organisms of Messinian reef units across the Western (Braga and Martín, 1996; Martín et al., 1997; Braga and Aguirre, 2001) and Central Mediterranean (Bosellini et al., 2001, 2002; Braga et al., 2009). Braga et al. (2009) also observed that the Messinian coralline fauna exhibits more affinity to present-day Mediterranean assemblages than to contemporaneous tropical species, as a response to the progressive cooling and isolation of the Mediterranean from the Indian Ocean from the Miocene onwards. Sparry calcite cements, mainly within intraparticle pores of the bioclasts (Fig. 8D and E), indicate the occurrence of early carbonate cementation. The occurrence of precursor framboidal pyrite replaced by telogenetic iron oxides (Fig. 8F), associated with the presence of the stress markers *V. bradyana*, *Bolivina*, and *Bulimina* in the interbedded shales, suggests intermittent low-oxygen conditions in this outer-shelf environment.

The transition to benthic assemblage 5 (A5) marks the reestablishment of inner-shelf conditions during the last stages of the pre-evaporitic phase in the Tabernas Basin. *E. macellum* and *E. crispum* are typical epiphytic species found in Mediterranean seagrass habitats (Murray, 2006; Mateu-Vicens et al., 2014), and their dominance, associated with common *A. beccari*, suggest a water depth of less than 50 m (Debenay et al., 2000). The rare occurrence of *V. bradyana* in both assemblages 4 and 5 may indicate intermittent dysoxia and/or the influence of continental runoff (Donnici and Barbero, 2002; Rossi and Vaiani, 2008), which is supported by the presence of carbonaceous fragments, derived from surrounding vegetated continental areas.

The sedimentary structures and morphological characteristics observed in the sandy laminated carbonates facies, particularly wrinkled-lamination (Fig. 9A) and cemented, parallel-laminated carbonates (Fig. 9E, F), indicate a stromatolitic origin (e.g., Martín et al., 1993). Similar stromatolitic fabrics, defined by their abundant siliciclastic grains and well-developed lamination, were described in the MSC succession of the Sorbas Basin (Martín et al., 1993; Braga et al., 1995; Bourillot et al., 2010). Major controls in the formation of these stromatolites include a significant supply of siliciclastic sediments combined with carbonate cementation, which promoted sediment trapping and preservation of the stromatolite's structure (Martín et al., 1993). Although the composition of the microbial-mat assemblage is difficult to determine (Riding, 2000), cyanobacteria were interpreted to be the dominant organism colonizing the slope setting in the Sorbas Basin (Braga et al., 1995). Other studies related MSC-bearing microbialites to chemotrophic-dominated microbial mats capable of colonizing environments from shallow to outer shelf depths (Dela Pierre et al., 2011, 2014). However, the estimated bathymetry and paleoenvironmental constraints given by benthic assemblage 5, carbonate cementation, and the textural and paleoenvironmental affinities with the Siliciclastic Stromatolites from the Sorbas Basin (see Martín et al., 1993; Braga et al., 1995), support a shallow-water origin for the stromatolites of the Yesón Alto section. Their deposition marks the highest degree of restriction in the inner shelf, preceding the deposition of the evaporites from Stage 1 of the MSC (Lacerda-Orita et al., 2025).

Finally, the amplification of precession climate-forcing in the sedimentation during the pre- and syn-MSC Stage (e.g., Krijgsman et al., 2001), likely played a role in the sedimentation of carbonates in the increasingly restricted Tabernas Basin. The influence of precession in carbonate productivity and platform growth has been well recognized in the Fringing reef unit, in the pre-MSC interval of the Sorbas Basin (Braga and Martín, 1996), and correlated to the sapropel/marl cycles of the basinward Abad Member (Sierro et al., 1999). In the Sorbas Basin, platform growth was related to insolation maxima periods, when the combination of higher surface water temperatures, normal salinities and small-scale sea-level rise favoured carbonate production (Sierro et al., 1999; Sanchez-Almazo et al., 2007; Roveri et al., 2009). Conversely, during insolation minima, a decrease in surface water temperatures,

increasing salinity, and sea-level fall limited the carbonate factory (Sierro et al., 1999; Roveri et al., 2009). In the Yesón Alto area, the lack of reliable biostratigraphic data within the Non-Distinctive Zone (Fig. 2) combined to a large SDDS (Fig. 3) in the uppermost pre-evaporitic interval limits a precise correlation between the carbonate beds and precessional cyclicity.

5.2. Pyritization of calcareous tests and organic matter influx

Paragenetic relationships between pyrite and calcareous tests allowed the characterization of the water-column redox-conditions and structure in the Tabernas Basin during an interval marked by important oceanographic changes in the Mediterranean. From approximately 6.4 Ma onwards, increasingly anoxic conditions were reported to develop in the bottom waters of different Mediterranean basins (Sierro et al., 2003; Kouwenhoven et al., 2006; Gennari et al., 2018) in response to the increasing limited circulation in the basin during the “consolidation phase” of the pre-MSC restriction of the Mediterranean (sensu Corbí et al., 2020).

The precipitation of authigenic pyrite in marine sediments is mainly controlled by the interplay between dissolved sulphate, organic matter availability, bio-mediated reactions, and poor oxygenation of bottom waters (Berner, 1978; Curtis, 1980; Hudson, 1982; Raiswell, 1997). In most marine settings, pyrite forms mainly through bacterial sulphate reduction, a biochemical process driven by anaerobic sulphate-reducing bacteria (SRB), occurring especially in fine-grained sediments containing a significant amount of organic matter (Hudson, 1982). Sulphate reduction by SRBs normally occurs at the sediment-water interface in anoxic environments, where SRB use sulphate as an electron acceptor to metabolize organic matter (Fig. 13). This process generates hydrogen sulphide (H_2S) as a byproduct, which can react with dissolved iron available in the pore waters to form pyrite (FeS_2) (Fig. 13).

In the Tabernas Basin, an elevated primary productivity in the upper waters is indicated by the presence of eutrophic planktonic taxa such as *Turborotalita*, *G. bulloides*, or neogloborotalids (Lacerda-Orita et al., 2025). This condition likely enhanced the flux of organic matter from the upper waters to a suboxic to anoxic seafloor (Fig. 13), fuelling microbial sulphate reduction processes and controlling the benthic fauna. The benthic taxa dominating in benthic assemblage 1, including *Bulimina*, *Globobulimina* spp., and *U. peregrina*, have been reported to even tolerate hydrogen sulphide (Rathburn et al., 2003). However, experimental analysis reveals that the tolerance of some infaunal foraminifera to sulphidic conditions is limited to survival, and prolonged exposures inhibit their reproduction and growth (Moodley et al., 1998). The association of *Uvigerina* and *Bulimina* has been related to high-productivity waters (Loubere and Fariduddin, 1999), indicating that sulphide- low oxygen tolerant taxa benefited from increased food supply (Enge et al., 2014) and from the reduced presence of aerobic scavengers. Increased organic matter flux may also explain the morphological variability observed in *B. aculeata* (Fig. 11) (e.g., Collins, 1989).

The precipitation of framboidal pyrite within intra-shell pores of both planktonic and benthic foraminifera (Figs. 12A-B, D-E) likely occurred through preferential replacement of soft tissues (Sawłowicz, 1993; Sawłowicz, 2000). Such cavities also provide reducing microenvironments that protect SRB colonies from more oxygenated waters or predation (Hudson, 1982), and the lower solubility of calcite tests favoured the precipitation of pyrite in the intra-shell space (Fig. 12A-E; 13B).

The co-occurrence of framboidal pyrite with disseminated euhedral crystals (Fig. 12H) and intermediate textures (Fig. 12C) likely reflects variations in the saturation conditions during diagenesis (Sawłowicz, 2000). In Eastern Mediterranean sapropels, framboidal pyrite precipitates rapidly under high saturation levels within sapropel levels, while euhedral habits result from slower precipitation under low saturation states below the sapropels (Passier et al., 1999).

The pervasive replacement of pteropod tests (Fig. 12H) reflects the

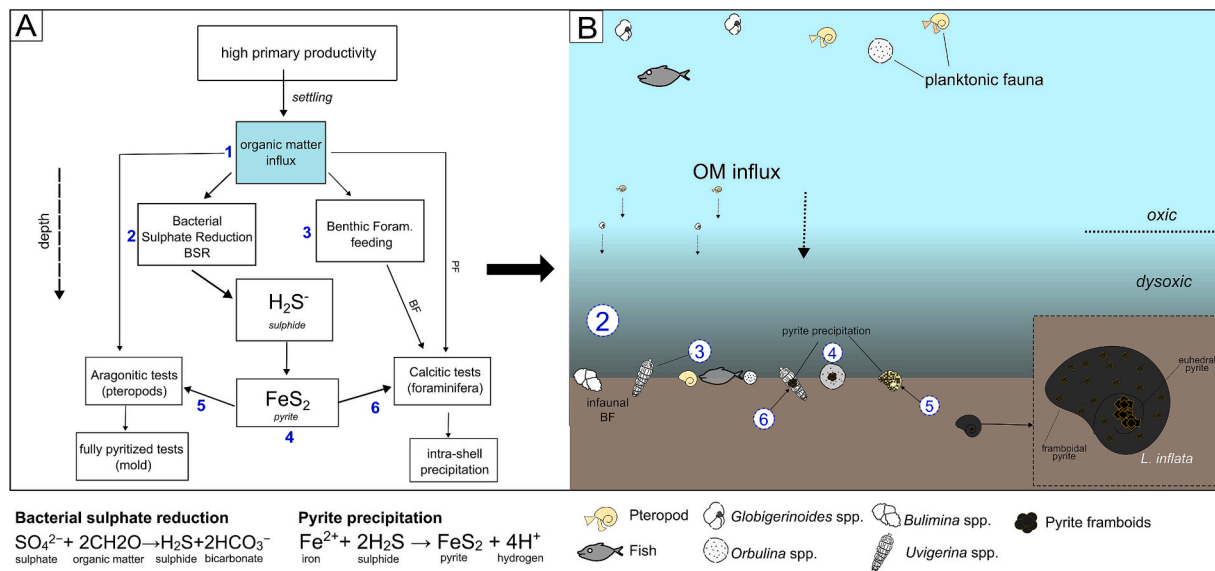


Fig. 13. Schematic model illustrating biotic and abiotic processes and reactions in a stratified Tabernas Basin. A) Diagram showing the processes and diagenetic pathways controlling the pyritization of calcareous tests. Numbers (1–6) indicate different processes, not necessarily in parasequence order. Their spatial distribution in the seafloor environment is illustrated in the model in B.

higher solubility of their aragonitic shells compared to the calcite tests of foraminifera (Herman, 1998). The pyritization of the aragonitic shell preserved its morphology and enabled taxonomic identification (Fig. 11), while the preferential replacement of the soft parts contributed to stabilizing the shell mould, thus reducing susceptibility to fragmentation during compaction or by bottom currents (Sawłowicz, 2000; Fisher, 1986).

Taphonomic biases can be generated by diagenetic processes, particularly by the dissolution of calcareous tests. These processes may represent a challenge for accurately reconstructing MSC-related paleoenvironments (Mancini et al., 2022). In contrast, in the Yesón Alto section, the paragenetic relationships between pyrite and the observed tests highlight the potential of early diagenesis for preserving microfossils during episodes of rapid environmental and ecological changes. During more oxic periods, the absence of pteropod fossils does not necessarily imply their disappearance from the upper waters but may instead reflect the dissolution of their aragonitic tests with the lack of pyritization at the seafloor. Therefore, water-column stratification and anoxic conditions at the bottom waters were essential for preserving an important part of the micropaleontological record, allowing a more complete reconstruction of pre-MSC marine ecosystems in the Tabernas Basin.

5.3. Indicators of environmental stress and biotic responses before the MSC

During the pre-MSC stage, evidence of “environmental stress” has been documented in basins from the Western (Sierro et al., 2001; Corbí et al., 2016, 2020; Lacerda-Orita et al., 2025), Central (Dela Pierre et al., 2012), and Eastern Mediterranean (Gennari et al., 2018, 2024; Bigi et al., 2024; Lanterna et al., 2026). Within the MSC framework, the term environmental stress is applied to characterize specific paleoenvironmental conditions that are interpreted as responses to the major oceanographic changes that occurred during this time, including water-column stratification, oxygen depletion at intermediate to deep depths, and abrupt salinity fluctuations. Conventionally, these stress conditions are typically inferred from their corresponding impacts on marine microfauna, particularly on foraminifera and calcareous nanoplankton (e.g., Gennari et al., 2024). These include large oscillations in the P/B ratio (from 0 to 100%), the increase of benthic species related to stressful

conditions (i.e., adapted to eutrophic, low-oxygen environments sensu Van Hinsbergen et al., 2005), and the progressive decline of calcareous microorganisms (Bertini et al., 2024). Our integrated micropaleontological and petrographic data show that water-column stratification developed in the Tabernas Basin at 6.164 Ma, consistently with the “consolidation phase” of the restriction of the Mediterranean (sensu Corbí et al., 2020). The intensification of water-column stratification and the poor ventilation of bottom waters were likely influenced by the superimposition of local tectonic activity in the Tabernas Basin (Lacerda-Orita et al., 2025) with the overall regionally driven restriction of the Mediterranean during the pre-MSC stage (Flecker et al., 2015).

The shift of benthic assemblages, from upper-epibathyal to an outer-shelf transition, may also indicate an increase in oxygenation levels of the bottom waters. This likely reflects the local effect of shoreface and coastal processes favouring the mixing of the water-column. In contrast, in the deeper parts of the basin water-column stratification and poor-ventilated bottom-waters was likely sustained or even intensified as restriction progressed.

The disappearance of foraminifera is another important event related to the shifting environmental conditions approaching the “restriction climax”. A reduction in foraminifera abundance before their disappearance during the pre-MSC stage is consistently recorded in many successions from the Eastern and Central Mediterranean (Blanc-Valleron et al., 2002; Kouwenhoven et al., 2006; Gennari et al., 2018; Manzi et al., 2007, 2018; Bigi et al., 2024). According to Gennari et al. (2024), foraminiferal reduction took place at slightly different intervals across different basins, but their disappearance occurred within the same precessional cycle that immediately preceded the MSC (~5.990 Ma).

In the Yesón Alto section (Fig. 10), planktonic foraminifera disappear approximately two precessional cycles before the overall decline of the marine fauna, coinciding with a transition to an inner shelf setting and the establishment of a mixed carbonate-siliciclastic environment (Fig. 14). This suggests that local basin dynamics promoting shallowing and restriction, likely associated with a global sea-level fall after 6.35 Ma (Jiménez-Moreno et al., 2013; Pérez-Asensio et al., 2013), played a significant role rather than Mediterranean-wide salinity fluctuations alone. Such a rapid shallowing was likely influenced by the uplift of the Betic Cordillera, whose ranges progressively emerged following a west-to-east trend during the Messinian (Braga et al., 2003). This differential

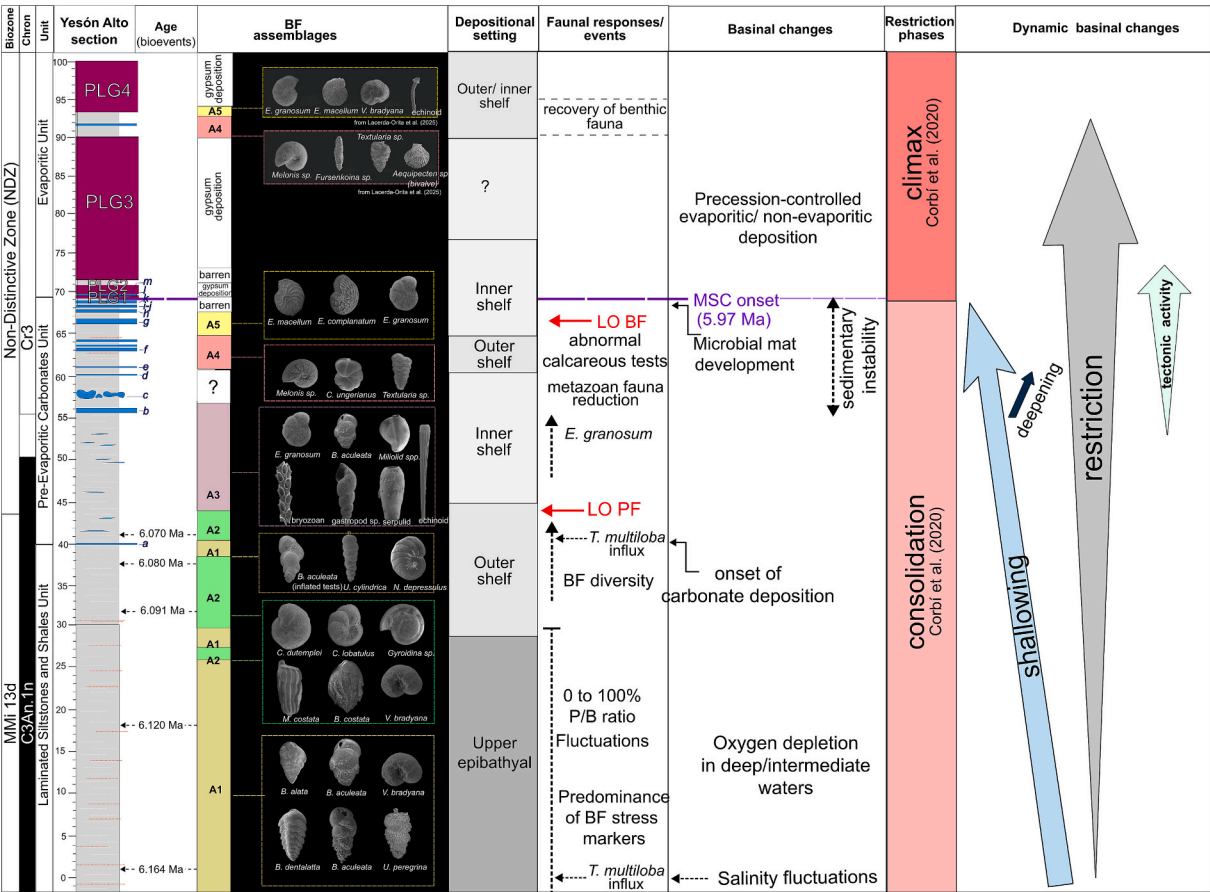


Fig. 14. Schematic chronostratigraphic framework illustrating the pre-MSC stage in the Yesón Alto section, summarizing the main paleoenvironmental, biotic, and sedimentary changes preceding and during the first phase of Stage 1 of the Messinian Salinity Crisis.

uplift largely controlled the progradation of sedimentary systems as the sea retreated eastwards, resulting in the limited development of evaporites during Stage 1 of the MSC in the Tabernas Basin compared to the Sorbas and Bajo Segura basins (Lacerda-Orita et al., 2025).

Salinity variations are likely recorded before the planktonic foraminifera disappearance, as suggested by two distinct influxes of *T. multiloba* (approximately at 0.1 and 41 m from the base respectively) (Fig. 14). This species can tolerate elevated salinities and is usually common in restricted marine environments (Sierro et al., 2003; Kouwenhoven et al., 2006; Lanterna et al., 2026). The abundance of deformed tests of *Melonis pompilioides* in benthic assemblage 4 (Fig. 11) further supports the influence of salinity fluctuations in triggering environmental stress in the bottom-waters. The deformation of foraminifera tests is linked to disrupted growth and is usually related to environmental stress (Geslin et al., 2002), but the link between a specific stress condition and particular test abnormalities remains unclear (Geslin et al., 2000; Murray, 2006). Yet, a high proportion of deformed tests of foraminifera reported in lagoons and estuaries from Brazil have been related to stress conditions such as hypersalinity, salinity fluctuations or seasonal acidification (Geslin et al., 2002). In the Yesón Alto section, the species *Bolivina dentellata* and *Bulimina echinata*, abundant in benthic assemblages 1 and 3, can also be associated with hypersaline environments (e.g., Kouwenhoven et al., 2006). A recent study reported a drop in salinity of surface waters at 6.015 Ma in the Eastern Mediterranean, attributed to an influx of fresher waters from the Paratethys, which resulted in a lid of fresh water above a layer denser and more saline bottom-waters where *B. dentellata* and *B. echinata* thrived (Lanterna et al., 2026). While the precise cause of deformation in *Melonis* spp. specimens in the Tabernas Basin remain uncertain, our integrated data suggest that it is most likely related to salinity fluctuations

during the late pre-MSC stage (Fig. 14).

A statistical study of a large fossil record dataset evaluated the impact of the MSC on the Mediterranean marine biodiversity (Agiadi et al., 2024). The model showed that water-column stratification and salinity fluctuations were directly correlated with a major decline in plankton species and in the biodiversity of demersal habitats, including the extinction of >20% of demersal fish species, 35.3% of bivalves (see Agiadi et al., 2024; and references therein). The crisis particularly affected coral reefs, which ceased to develop in the Mediterranean from the Pliocene onward (Agiadi et al., 2024). Despite the evident correlation between evaporite deposition and marine biodiversity loss, studies from different Mediterranean basins report that some marine organisms, including foraminifera, diatoms, and calcareous nanoplankton, persisted during gypsum deposition stages (Pellegrino et al., 2021), interbedded with halite bodies (Bertini et al., 1998; Meilijson et al., 2019) and in inter-evaporitic deposits (Goubert et al., 2001; Corbí et al., 2016; Lacerda-Orita et al., 2025; Sierra-Ramirez et al., 2026). In the Betic Cordillera region, marine fauna experienced a marked decline in the uppermost pre-evaporitic interval (this study Fig. 14; Sierro et al., 2001; Goubert et al., 2001). This trend is usually followed by a periodic recovery of benthic fauna limited to one precessional cycle intercalated with the deposition of thick, massive gypsum beds during Stage 1 of the MSC (Goubert et al., 2001; Lacerda-Orita et al., 2025). Due to their partial isolation, marginal basins may have represented possible biodiversity hotspots during the Late Miocene (Corbí, 2010), which highlights the importance of further integrated stratigraphic and paleoenvironmental studies in these settings.

5.4. Development of microbial mats and the transition to the Evaporitic Unit

The Primary Lower Gypsum is a lithostratigraphic unit deposited during Stage 1 of the MSC (5.97–5.6 Ma; Roveri et al., 2014) and has been recognized across many marginal basins of the Mediterranean (see Lugli et al., 2010, and references therein). It consists of a cyclic alternation of gypsum and siliciclastic/carbonate facies, forming evaporitic/interevaporitic couplets. The notable regularity of these cycles has been attributed to precession-paced climatic forcing, similar to the sedimentological cyclicity observed during the pre-MSC stage (Krijgsman et al., 1999). According to this model, evaporites precipitated during arid phases and increased restriction, while interevaporitic cycles were deposited during more humid phases and reduced restriction. Other marginal basins successions across the Mediterranean record sequences of carbonates below the Primary Lower Gypsum, which are often referred as the term ‘calcare di base’ (Manzi et al., 2011).

The evaporitic succession of the Tabernas Basin comprises alternating massive selenite beds and predominantly fine-grained siliciclastic facies (Lacerda-Orita et al., 2025) and has been correlated with the first four basal cycles of the Primary Lower Gypsum in the Sorbas Basin (Manzi et al., 2013; Roveri et al., 2018). A key marker is the third interevaporitic bed, notably rich in benthic fauna, and that has also been described in the same stratigraphic interval in the Sorbas Basin (Saint-Martin et al., 2000; Goubert et al., 2001; Néraudeau et al., 2002; Sierra-Ramirez et al., 2026). In the Tabernas Basin, this interval records a shallowing-upward shift from an outer-shelf assemblage dominated by *Melonis* spp. and *Fursenkoina schreibersiana*, to an inner-platform assemblage dominated by *Elphidium* spp., irregular echinoids, bivalves, and bryozoans (Fig. 14, Lacerda-Orita et al., 2025). In the middle of the interval, a thin sandy bioclastic packstone bed resembling bed *b* from the pre-evaporitic stage occurs (Figs. 5 and 13). This interval has been interpreted as being deposited during the reestablishment of nearly normal marine conditions in the basin during a precession minimum (Lacerda-Orita et al., 2025), following the marked decline in marine fauna observed prior to the onset of the MSC (Fig. 14).

In the uppermost part of the pre-MSC succession, benthic assemblages 4 and 5 (Fig. 10) records a similar faunal and bathymetric shift. We thus suggest that minor sea-level oscillations or changes in marine connectivity, potentially linked to tectonic activity, controlled restriction levels during the late pre-MSC stage and early stages of the MSC in the Tabernas Basin. In contrast to Stage 1 of the MSC, when a new cycle of massive selenites was deposited after a shift towards more arid conditions (Lacerda-Orita et al., 2025), during the late pre-MSC Stage, this shift led to the establishment of microbial mats (Figs. 9 and 14).

At the northern margin of the Sorbas Basin, microbialite-dominated sequences developed, with salinity fluctuations identified as the primary driver of microbial blooms (Bourillot et al., 2008, 2010; Bosellini et al., 2002). A reduction in competition with metazoans is considered to favour stromatolite formation (e.g., Martín and Braga, 1994), as observed in modern settings, such as in Shark Bay (Australia), where seasonal hypersaline conditions reduce metazoan grazers (Logan, 1961). Similar conditions likely occurred in the inner-shelf of the Yesón Alto area during the late Pre-MSC stage, where the combination of increasing restriction, shallowing, fluctuating salinity, and reduction of metazoan fauna (Fig. 14) may have favoured the development of microbial mats in the climax of the restriction (sensu Corbí et al., 2020), marking the transition from a carbonate-dominated system to evaporitic conditions.

6. Conclusions

The establishment of a mixed carbonate-siliciclastic shelf between 6.080 and 5.971 Ma in a marginal marine basin of the Western Mediterranean (Tabernas Basin) was influenced by the major oceanographic and paleoenvironmental changes associated with the advanced state of basin restriction immediately preceding the Messinian Salinity Crisis.

Between ~6.164 and ~6.080 Ma, upper-epibathyal settings were already subjected to environmental stress primarily driven by the stratification of the water column. This resulted in fluctuations in the planktonic/benthic foraminifera ratio and in the increase in sulphide/dysoxia-tolerant benthic taxa at the seafloor. Paragenetic relationships between framboidal pyrite and calcareous tests indicate that high primary productivity in the upper water column generated a positive flux of organic matter to deeper water, fostering microbial sulphate reduction and controlling the ecology of benthic fauna. Rapid precipitation rates of pyrite in this reducing environment favoured the differential preservation of different types of calcareous tests.

A shallowing-upward trend led to the deposition of silty foraminiferal wackestones in the outer-shelf, marking the onset of carbonate deposition in the basin as restriction intensified. The relative increase in benthic carbonate producers, such as bryozoans, echinoids, serpulids, and gastropods, and the deposition of laterally continuous sandy bioclastic packstones, followed this trend and marked the transition to an inner-shelf. The presence of abundant terrigenous sediments highlights the continuous influence of marginal depositional systems on the carbonate factory.

Test deformation in *Melonis pompilioides* and the progressive disappearance of marine fauna indicate deteriorating marine conditions during the late pre-MSC stage, likely related to fluctuations in salinity. The deposition of thin crustose algal frameworks occurred during episodes of reduced siliciclastic inflow. Further shallowing towards a restricted inner shelf environment, occurring at the climax of the restriction, was accompanied by a total loss of benthic foraminifera and metazoan fauna. In this setting, microbial mats colonized the substrate, marking the highest degree of restriction prior to the deposition of evaporites related to Stage 1 of the MSC.

The progressive restriction of the Tabernas Basin recorded in the upper pre-MSC deposits was the main factor controlling the depositional, diagenetic, and ecological evolution of the mixed carbonate-siliciclastic shelf. This study provides new insights into the dynamic interplay between basin restriction, diagenesis, and paleoenvironmental changes, revealing a detailed chronology of events preceding the Messinian Salinity Crisis in a marginal setting in the Western Mediterranean.

CRediT authorship contribution statement

Gustavo Kenji Lacerda-Orita: Writing – review & editing, Writing – original draft, Visualization, Validation, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Fernando Pérez-Valera:** Writing – review & editing, Visualization, Validation, Methodology, Investigation, Formal analysis, Data curation. **Hugo Corbí:** Writing – review & editing, Visualization, Validation, Supervision, Methodology, Investigation, Formal analysis, Data curation. **Jesús M. Soria:** Writing – review & editing, Visualization, Validation, Methodology, Investigation, Formal analysis. **Telm Bover-Arnal:** Writing – review & editing, Visualization, Validation, Methodology, Investigation, Formal analysis, Data curation. **Enrique Gomez-Rivas:** Writing – review & editing, Visualization, Validation, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis. **Luis Gibert:** Writing – review & editing, Visualization, Validation, Supervision, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.palaeo.2026.113791>.

Data availability

The authors confirm that all data necessary for supporting the scientific findings of this paper have been provided.

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