Reefal and mud mound facies development in the Lower Devonian La Vid Group at the Colle outcrops (León province, Cantabrian Zone, NW Spain)

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Abstract In the locality of Colle (Cantabrian Zone, NW Spain), the upper part of the Valporquero Shale Formation (Emsian, La Vid Group) contains an interval of shales and marlstones (barren, greenish-grey shales and fossiliferous, greenish-grey or reddish shales/marlstones) with beds and packages of homogeneous and cross-bedded skeletal limestones. Metre-scale mud mounds and coral biostromes occur encased in the fossiliferous reddish and greenish-grey shale/marlstones, respectively, with the coral biostromes overlying conspicuous skeletal limestone bodies. These rocks were deposited on a carbonate ramp, ranging from above storm wave base for the cross-bedded skeletal limestones to below the storm wave base for the remaining deposits, organic buildups included. The vertical stacking of these facies and the occurrence of the two types of buildups are interpreted to reflect the interplay among several (possibly 4th and 5th) orders of relative sea-level variations, during a 3rd-order highstand. Coral biostromes occur in early 5th-order transgressive system tracts developed within late 4th-order highstand, and are interpreted to have thrived on a stable granular substrate (skeletal limestones) in non-turbid waters, being later aborted by the onset of muddy sedimentation. Biostrome features suggest that they developed under environmental conditions essentially different from those related to the sedimentation of their granular substrate. Mud mounds occur in 5th-order transgressive and early highstand system tracts tied to early 4th-order sea-level rise. Field relationships suggest that mud mounds grew coevally with muddy sedimentation, with high-frequency variations in carbonate vs. terrigenous mud sedimentation influencing their development.

Keywords Cantabrian Zone · Lower Devonian · Carbonate ramp · Coral biostromes · Mud mounds · Allocyclic controls

Introduction

Since the 19th century, the village of Colle (Cantabrian Mountains, León province, NW Spain; Fig. 1) is well known for its fossiliferous Lower Devonian deposits, which form an interval belonging to the Valporquero Shale Formation (Vilas Minondo 1971; Vera de la Puente 1989) of the La Vid Group (Comte 1959; García-Alcalde and Racheboeuf 1978). In literature, the fossils are described as “Sabero-fossils” named after a small mining town situated close to the village of Colle (Fig. 1). More than 200 species from different groups (e.g., brachiopods, crinoids, blastoids, bryozoa, corals, stromatoporoids, trilobites, nautilids) had been cited, but mostly without any relation to the facies and sedimentology of the succession and mainly from non-reefal strata. Previous studies on the Colle section with special emphasis on its facies were carried out mainly by Stel (1975) and Fernández et al. (1995). The study of Stel (1975) was focused on the origin of the rugose coral biostrome at the base of the Colle section, whereas Fernández et al. (1995) worked out a facies-oriented overview section of the complete calcareous succession in the La Vid Group near Colle. A brief overview of the biostromal reef development in the Colle section was published by Brouwer (1964) and Soto (1981, 1982). Finally, Schmid et al. (2001) summarized the features of the mud mounds that occur
some metres above the coral biostrome. Other papers deal with this interval of the Valporquero Formation, either mentioning the Colle section (Leweke 1982) or studying it in relatively nearby areas of the same structural unit (Esla Unit; Fig. 1B; Ruhrmann 1971) or in the broader context of the southern part of the Fold and Nappe Province (Keller 1988).

The aim of this paper is to provide for the first time a detailed and comprehensive data set on the faunal composition, facies types and the development of the organic build-ups (coral biostromes and mud mounds) of this well-known locality of Colle. Moreover, in the light of these results, we discuss the controls on the development of the deposits, including the various reef types, in the Colle section, mainly in the light of a sequence stratigraphic framework. The vertical arrangement of facies suggests that several orders of cyclicity modulated the sedimentation. According to the literature and the available data, it seems that tectono-eustatic 3rd-order cycles interplayed with higher order (4th–5th?) cycles, the latter being possibly the result of a combination of regional tectonics and of climate fluctuations.

**Geological setting**

The studied area is located in the Esla Unit of the Cantabrian Zone (Lotze 1945), which is the external zone of the Variscan Iberian Massif in NW Spain (Fig. 1A). From a tectono-stratigraphic point of view, the Cantabrian Zone has been subdivided into several provinces (Julivert 1971) and units (Pérez-Estaün et al. 1988), one of which is the Esla Unit (Fig. 1A and B). From Cambrian to Devonian and prior to the onset of the Variscan orogeny during Carboniferous time, the Cantabrian Zone hosted a shallow-marine sedimentation with some alluvial deposits on a passive continental margin (e.g. Ribeiro et al. 1990; Aramburu et al. 2004), which resulted in a rather complete stratigraphic record, punctuated by hiatuses of variable duration and areal extent. The Devonian record comprises an alternation of calcareous and siliciclastic formations, which have been grouped into two types of successions based on their facies. These are the so-called Palentine and Astur-Leonese facies (Brouwer 1964; Fig. 2). The Palentine succession, whose features suggest deposition in distal and deeper areas of the sedimentary basin, has an allochthonous character being restricted to the Variscan Palentine nappes (Fig. 1A), which were emplaced gravitationally in the Pisuerga-Carrión Province from the south (see Frankenfeld 1983; Marqués and Marcos 1984; Rodríguez Fernández and Heredia 1988). In contrast, the succession with Astur-Leonese facies (Fig. 2) has a shallower and more proximal character, and crops out in the Fold and Nappe Province. Its basal part comprises the Lochkovian-Emssian La Vid Group and its lateral counterpart in the north, the Rañeces Group, which crop out in the Somiedo-Correcillas and Esla units (Figs. 1 and 2).

The La Vid Group is dominantly formed of carbonates (limestones and dolostones), although a thick shaly unit is present in its middle and upper part (Fig. 2). The group has been subdivided into four formations: Felmín, La Pedrosa, Valporquero and Coladilla (Vilas Minondo 1971; Vera de la Puente 1989), although other subdivisions have been later proposed (Leweke 1982; Keller 1988; Fig. 2). The Valporquero Formation (Emsian) comprises the shaly unit of the middle-upper part. This shaly unit is rather monotonous, although its upper part contains a marly interval with limestone intercalations that constitutes the studied interval (Fig. 2). These limestones are mostly detrital, although some coral biostromes and mud mounds exist as well. Keller (1988) named this interval of the Valporquero Formation as the Sagüera Member of his Esla Formation. It also corresponds to the Intermediate Limestone Member of Leweke (1982).

The La Vid Group was deposited on a carbonate ramp and its stratigraphic succession has been tied to two 3rd-order transgressive–regressive cycles (Vera de la Puente 1988; Keller and Grötsch 1990; Keller 1997; Fig. 2). The shales of the Valporquero Formation are thought to record the highstand of the upper 3rd-order cycle. Within this shaly unit, the coarser-grained succession of the studied interval would record a minor (4th-order?) lowstand.

The Devonian limestone formations of the Cantabrian Zone contain a wealth of reefal and reef-related deposits. These deposits, which mainly appear in the Astur-Leonese facies, occur from the Pragian (Nieva Formation of the Rañeces Group and Lemanza Formation of the Palentine nappes) until the Givetian (Candás-Portilla Formation), and locally the Frasnian (Crémenes limestone of the Nocedal Formation), forming five main reefal episodes (Fig. 2; Méndez-Bedia et al. 1994; Fernández et al. 1995; see a review in García-Alcalde et al. 2002). The coral biostromes and mud mounds of the studied interval represent the second reefal episode.

**Methods**

In the vicinity of Colle (Fig. 1C), the studied interval crops out patchily along a hill slope, where several gullies and footpaths provide sections of acceptable quality. The best exposed section was selected as the main section (Fig. 3), and two others, located within a distance of...
Fig. 2 Chronostratigraphic chart showing the Devonian lithostratigraphic units of the Astur-Leonese facies that have been defined in the political regions of Asturias (northern part of the Cantabrian Zone) and Leon (southern part of the Cantabrian Zone) and the distribution of the reefal episodes. Absolute ages based on Gradstein et al. (2004). The stratigraphic subdivision of the La Vid Group is that of Vilas Minondo (1971) and Vera de la Puente (1989), but subdivisions by Leweke (1982) and Keller (1988) are also shown. Note that, contrary to other authors, Leweke (1982) treats La Vid Group as a formation made up of members. The log on the right depicts the interpreted relationships between the general stratigraphy of the La Vid Group and the sea-level curve (based on Keller and Grootsch 1990) and shows the location of the studied interval of the Valporquero Formation.

The studied interval (Saguera Member of Keller 1988; see above) is composed of greenish and reddish shales and marlstones with interleaved limestones (Leweke 1982). Field features of the deposits were used to separate a number of facies that are dealt with in the following section. Rock slabs and stained thin sections were studied for detailed features of sediments and fossils.

Major lithological contacts were physically traced between sections where possible and drawn on field photo-mosaics and panoramic photographs to produce a stratigraphic cross section that summarizes both physical and inferred correlations, and used to determine the geometry of sedimentary bodies and their mutual relationships. The hill slope trends approximately NW–SE. This direction is oblique to the depositional dip of the Devonian basin, whose distal and deeper parts would be located to the south in this sector of the Cantabrian Zone (e.g., see Aramburu et al. 2004, and references therein).

200 m from the latter, were additionally studied. The studied interval approximately coincides with that discussed by Stel (1975, see his Fig. 1), and is bound above and below by poorly exposed shales with scarce limestone intercalations (Lower Shale and Upper Shale members of Leweke 1982). Field features of the deposits were used to separate a number of facies that are dealt with in the following section. Rock slabs and stained thin sections were studied for detailed features of sediments and fossils.

Facies

The studied interval (Saguera Member of Keller 1988; see above) is composed of greenish and reddish shales and marlstones with interleaved limestones (Fig. 3). Limestones are mainly detrital (bioclastic) packstones to grainstones that form thick bodies and thin beds. Subordinately, some small but locally conspicuous coral biostromes and mud mounds exist. Based on their lithological features and faunal content, the shale/marlstone and limestone deposits of the studied interval can be grouped into a number of facies, described and interpreted below. Special emphasis is given to the biostromal and mud mound facies.

**Facies A: barren, greenish-grey shales**

*Description*

It comprises greenish-grey shales to marly shales lacking any detectable macrofossil content (Fig. 4A). These deposits form principally the lower and upper parts of the
studied interval (Fig. 3) and are also the major constituent of the remainder of the Valporquero Formation.

**Interpretation**

This facies is interpreted as deposited by clay fallout from suspensions in a deep and low-energy environment well away from the action of currents transporting particulate sediment, as suggested by the lack of interleaved coarser-grained deposits. Absence of macrobiota could be due to several factors, not necessarily mutually exclusive: unfavourable, possibly somewhat stressed (oxygen-depleted?) bottom conditions (see also Stel 1975; Keller and Grötsch 1990), and the lack of a stable, particulate substrate suitable for a fauna to attach. A high rate of clay supply could have also contributed to the absence of macrobiota, although not being itself a determinant factor, since the mudrocks of facies B and C (see below) contain abundant fauna. In contrast to the other mud-rich facies (facies B and C), the absence of both coarse-grained sediment layers and benthos suggests that probably this facies represents the most distal and deepest setting of the ramp, possibly below the pycnocline.

**Description**

Facies B: fossiliferous, greenish-grey shales/marlstones with interleaved skeletal limestones

This facies mainly consists of greenish-grey shales to shaly marlstones. Locally, skeletal packstones to wackestones are present forming alternations with the shales/marlstones in subequal proportions, although more commonly they are a subordinate constituent (Figs. 3 and 4B–C). The shales/marlstones contain abundant macrobiota, which comprises slightly disintegrated crinoids and blastoids, brachiopods (spiriferids and terebratulids), bryozoans (fenestellids, ramose forms, mushroom-shaped fistuliporids), favositids (*Crenulipora*) and auloporids (*Bainbridgia, Schluerichonus*). The skeletal wackestones to packstones appear as cm-thick beds, which either are tabular or thin and pinch out laterally. Some beds laterally pass into the surrounding muddy rock due to mixing by burrowing. Bioclasts are variably bioabraded and/or iron-stained and essentially belong to the same taxa present in the muddy deposits. They comprise disintegrated echinoderm plates (mainly crinoid and blastoid ossicles), although some whole calyces exist. Fragments of bryozoans (*fenestellids, fistuliporids and ramose forms*), favositids (*Crenulipora*) and auloporids (*Bainbridgia, Schluerichonus*) and other bioclasts (such as brachiopods) that exist within the beds are interpreted as infauna and epifauna that colonized the granular substrate after currents laid down their bioclastic load (see interpretation of facies D). The related burrowing activity would also account for the bedding destruction and mixing of the granular beds with the underlying muddy sediment. Nevertheless, it cannot be precluded that in some cases the skeletal limestone beds may represent “condensed” intervals, in which, a diminished rate of clay fallout, possibly tied to minor transgressive episodes, could have resulted in a deposit enriched in skeletal components and in lime mud. Thus, as a whole, this facies would represent an environment close to or above the storm wave base, shallower than the setting of deposition of the former facies. It is conceivable that these shallower conditions, plus a more intense water circulation, partly related to the currents that sporadically swept the sea bottom and deposited the skeletal beds, would have resulted in more oxygenated bottom conditions, favourable to host benthic communities (see also Stel 1975).

Facies C: fossiliferous, reddish shales/marlstones with interleaved skeletal limestones

This facies is similar to the former (Fig. 4B–D), the main differences being the reddish colour of the shales/marlstones, and the generally higher fossil content of facies C. In addition, the muddy rock has a higher fine-grained carbonate content, judging from its more marly appearance. The fossil content is dominated by echinoderms (crinoids and subordinate blastoids), bryozoans (*fenestellids, mushroom-shaped fistuliporids and occasional ramose forms*), brachiopods (*spiriferids and terebratulids*), diverse tabulate corals, such as ramose favositids (*Crenulipora, Thamnoptychia, Dendropora*) and auloporids (*Schluerichonus, Cladochonus, Bainbridgia*). Skeletal limestone beds are identical to those of the former facies, and bear the same relationships to the host muddy sediment, containing the same types of bioclasts and whole fossils.
**Interpretation**

The setting of deposition of this facies seems to have been essentially much the same as in the case of the facies B above, as the faunal content and the types of skeletal limestones suggest. The most striking feature of this facies is the reddish colour of the shales/marlstones, which is likely a syn-sedimentary feature, generated by bacterial activity in the marine environment (see Bourque and Boulvain 1993; Preat et al. 1999; Boulvain 2001). The slightly greater abundance of fauna in the background muddy sediment, compared to the facies B, could indicate that conditions were more favourable for the development of benthic communities. The more marly appearance of this rock and the fact that mud mounds are present only in the intervals of this facies are suggestive of a lowered clay input (see below).

We speculate that this lowered clay input, probably related to an overall lower sedimentation rate, would have resulted in less turbid seawater favouring the growth of benthic organisms. So, improving growth conditions together with reduced “dilution” of skeletal remains by mud deposition account for the high faunal content of this facies. The vertical relationships between reddish- and greenish-grey-shale intervals suggest that the formation of these two facies was controlled by allocyclic, long-term factors, as it will be discussed below.

**Facies D: thick-bedded, skeletal limestones**

**Description**

This facies is formed of skeletal pack- to grainstones. Locally, they contain large fragments of corals and constitute michelinid or Synaptophyllum float/rudstones (e.g., unit 1 of Figs. 3 and 5B; see also Fig. 6). These limestones form cm- to dm-thick, laterally continuous to wedge-shaped beds. Some beds, especially the coarsest-grained float/rudstones, are structureless, although others display cross bedding. Cross bedding is commonly of a low-angle type (Fig. 4C), and is frequently recognized as hummocky cross-stratification (HCS). The beds of this facies are stacked and amalgamated to form dm- to m-thick packages (Fig. 4C). These packages may fine and thin upwards until they are formed of skeletal limestones with shale/marlstone interbeds. The uppermost limestone bed(s) of these packages are generally richer in a muddy matrix, and this matrix is identical to the overlying sediment (facies B or C). These packages have an erosive base; they are lenticular in shape and extend laterally for at least some tens of metres before they pinch out or interfinger with shales/marlstones until they finally pass into facies B or C intervals (e.g. unit 12 of Figs. 3 and 4B–D). These packages may appear as relatively thin packages capping a thickening-upward and coarsening-upward interval of alternations belonging to either the facies B or C (Fig. 3). More commonly, they form thicker packages that sharply overlie an interval of the facies B or C with a conspicuous erosional surface (Figs. 3 and 4C). In some cases, these erosional surfaces have a relief of at least 1 m, against which the beds forming the packages or their laterally equivalent alternations onlap (Fig. 4D). In places, the high relief of the erosional surfaces is caused by the presence of mound structures, which are more resistant to erosion than the adjacent muddy sediment. The rare coral float/rudstones (e.g. Figs. 4D and 5B) are present at the base of the packages above these major erosional surfaces (e.g. base of unit 1 in Fig. 3; see also Fig. 6).

Skeletal grains in this facies comprise bioclasts of echinoderms and bryozoans (fenestellids, fistuliporids and ramose forms) as the main components. Less common elements are coral fragments, brachiopods, trilobites, and ostracods. In some cases, whole brachiopods are found (Uncinulus, Atrypa and Xystostrophia). As in the case of the thin skeletal beds of facies B and C, bioclasts display borings and iron-stained margins and, in some cases, they constitute lithoclasts, because their diagenetic history differs from that of the hosting limestone, suggesting they were partially lithified before deposition. For example, some corals display cavities connected to outside that are only filled with cements, lacking the bioclastic muddy matrix that is found surrounding the coral. In other cases, coral cavities display overturned geopetal fills (Fig. 5B). Finally, the edges of some grains truncate the cements filling the grain cavities.

![Fig. 3](image-url) Simplified stratigraphic log of the main section described in the locality of Colle, showing the facies distribution and the coarsening-upward or fining-upward trends

**Interpretation**

The features of this facies suggest that it resulted from deposition from traction currents, which, at least in some cases, were produced by storms (HCS). A similar storm-related interpretation has been reached by Ruhrmann (1971), Stel (1975) and Keller (1988), for these Colle facies and for laterally equivalent deposits located elsewhere in the southern part of the Fold and Nappe Province. The variation of thickness and internal structuring of beds is thought to reflect proximal–distal trends or the magnitude of the currents (cf. Aigner and Reineck 1982; Aigner 1985). The deposits of the weakest currents, or the most distal deposits of the strongest currents, would be represented by the thinnest beds of this facies and by the thin beds interlayered in the facies B and C, whereas the proximal deposits of the strongest flows would consist of HCS thickest beds. Structureless beds could be due to flows not being able to exert enough traction on grains as they were being deposited (see a detailed discussion on this topic in Chapter 3 of Pickering et al. 1989; and references therein), or also to post-depositional homogenisation by burrowing. As a whole, this facies would have been deposited in an environment above storm wave base, which was episodically swept by high-energy events. The paucity and vertical distribution of fine-grained sediment layers in between the...
successive beds within packages suggest that calm conditions were of short duration and/or that the currents were able to erode the muddy sediment deposited during the fair weather episodes. The higher frequency of the muddy intervals to the top of packages would indicate a progressive waning of the successive storms, which, given the overlying muddy facies, is interpreted to record the progressive deepening of a relatively shallow environment. The types of grains suggest erosion, transport and deposition of relatively nearby bottom-communities by the currents. Nevertheless, in some cases, erosion was strong enough to affect a buried and partly lithified sediment and rip it up as lithoclasts (skeletal grains with an advanced diagenetic stage). Most of these lithoclastic grains (coral fragments) are not found in the neighbouring substrate, suggesting a longer-distance transport from shallower areas. Whole bra-
chiopods that are present within the beds are interpreted as in situ dwellers, which colonized the granular substrate after its deposition. The related burrowing activity would also account for the muddy matrix of the uppermost beds of the skeletal deposits, which is interpreted as the product of mixing of a muddy sediment with the granular bed.

Facies E: *Synaptophyllum* biostromes

**Description**

This facies occurs in an interval of the lower part of the section (unit 1 of Fig. 3), which corresponds to unit C of Stel (1975). The deposits of this facies take place as a few thin (less than 30 cm thick) and laterally discontinuous (tens of m wide) biostromal horizons that alternate with skeletal limestones (facies D) and fossiliferous, greenish-grey shales/marlstones (facies B) forming several dm-thick sequences of facies. In each sequence of facies, a biostromal horizon overlies a bed of skeletal limestones, which in some cases are *Synaptophyllum* rudstones, and is in turn overlain by the fossiliferous, greenish-grey shales/marlstones or by the basal limestone bed of the next sequence of facies (Figs. 3 and 4E). The matrix between the colonies of the biostromes is a bioclastic and fossiliferous shale/marlstone identical to the overlying deposits (facies B). The organisms of the biostromes are mostly in situ and are dominated by branching colonial rugose corals represented by *Synaptophyllum* and by small numbers of massive colonies of the genus *Cantabriastrea* (Schröder and Soto 2003) <10 cm in diameter. Tabulate corals, stromatoporoids, some of which are overturned, and subordinate fistuloporidae bryozoans and atrypid chiopods are also present. Tabulate corals, which usually grow on dead *Synaptophyllum* branches and some of which are infected with tubes of the genus *Helicosalpinx*, are represented by Favositid, Alveolites, Heliolites and *Thannopora*. Stromatoporoids display symbiotic syringoporoids (*Caunopora* state) and belong to the genera *Salairella*, *Stromatopora*, *Neosyringostroma* and *Stromatoparella*.

Favositids and stromatoporoids mainly occur in the biostromal horizons of the uppermost sequences of facies (Fig. 4E) and comprise rather flat (laminar to low-relief domical) and relatively large forms, up to 20–30 cm in diameter. Heliolitid corals form small spherical colonies, whereas alveolitids occur as irregular encrusting forms. Finally, *Thannopora* colonies are small and display a scarce number of branches. Alveolitid, heliolitid and favositid corals show regular and continuous growth patterns, lacking the growth banding that is fairly usual in heliolitids, although in some cases it exists being very subtle. Also, stromatoporoids generally show regular latilamination.

Few and small necrotized areas affect the flanks of the coral and stromatoporoid skeletons. Internally, massive tabulate corals and stromatoporoids display some thin sediment laminae that veneered the upper surface of the skeleton and were later overgrown. These sediment laminae are mainly of micrite, although some of them are clay-rich.

Generally, the internal cavities of the colonies are only filled with cement. Nevertheless, the uppermost cavities of the colonies (e.g., see Fig. 5C) and those adjacent to the necrotic portions of the skeleton flanks are completely or partially filled with sediment. Three main generations of cement can be distinguished. The first one consists of a thin isopachous rim of inclusion-rich fibrous calcite; the second generation is formed of inclusion-rich columnar crystals of calcite with undulose extinction (radiaxial-fibrous calcite?); and finally the third generation consists of equant crystals of calcite, which in some cases are also inclusion-rich. These generations of cement were seen in thin sections made for palaeontological determinations that had not been stained, and thus ferroan vs. non-ferroan calcite could not be distinguished. The sediment infilling the cavities can be of two types, of rather pure micrite or of shale/marlstone. Both types contain a variable amount of bioclasts. The uppermost internal cavities of the colonies only contain the clay-rich type of sediment, whereas those cavities close to the necrotic portions of the flanks may display both the micritic and the clay-rich types. In some thin sections, the sediment that fills the uppermost cavities of the colonies postdates the first generation of (isopachous) cement (Fig. 5C).

**Interpretation**

The occurrence of the *Synaptophyllum* biostromes atop skeletal beds reveals the need of a stable substrate for the colonies to develop, a conclusion also reached by
This is also proven by the alveolitid colonies, which belong to forms with an encrusting habit, and by the morphology of the alveolitid and heliolitid colonies, which is indicative of restriction to lateral growth. Thus, the granular substrate constituted by the skeletal limestone beds could be regarded as the stabilization stage of the “reefoid” episode. Nevertheless, this does not mean that both deposits, the granular bed and the biostromal interval, are related and form a continuum. Since the granular sediment of the substrate is not found between the colonies and these occur clearly on top of the granular bed, we conclude that both deposits formed under very different environmental conditions in two separate episodes of sedimentation. On the other hand, the clay-rich sediment that surrounds the colonies and is also found as internal sediment in their cavities is very likely a later feature, not coeval to the biostrome development either. Several data back this conclusion. The presence of clay-rich sediment post-dating cements in coral cavities clearly indicates that this sediment records an essentially later episode of sedimentation. The few and small necrotic areas affecting the flanks of the skeletons suggest that killing processes by sedimentation (“growth-interruption banding” of Young and Kershaw 2005) were not frequent or important, and thus that sediment input during biostrome development was neither frequent nor intense. In addition, the rather pure carbonate sediment that fills the internal cavities of the skeletons adjacent to these necrotic areas suggests that the rare sedimentation events related to killing of skeleton flanks were clay-poor, clearly different from the type of sedimentation that is recorded in the clay-rich deposits found among the colonies and in the cavities open to outside. Summarizing, it seems more likely that three stages can be distinguished in biostrome development. During the initial stage of high energy, granular sediment was deposited forming the basal skeletal limestone bed. After the first stage, environmental conditions became more tranquil and biostromes developed during the second stage. In this second stage, it is possible that episodic currents were able to rip up and rework the tabular skeletons of stromatoporoids that are found overturned. These relatively high-energy episodes would also account for the presence of isopachous rims of fibrous cement that are in some cases seen to pre-date the sediment within cavities. This relatively early phase cementation and the radiaxial-fibrous calcite cement are assigned to a phreatic marine diagenetic environment. Several criteria back this interpretation, their fibrous and radiaxial-fibrous crystal habits, their occurrence forming the earliest pore-fillings, the comparison of the fibrous calcite with the non-ferroan fibrous calcite cement found in the mud mound facies. An early marine cementation is favoured by an active fluid circulation through pores, and this pumping is particularly active where sea floor is swept by currents (see Chapter 7 of Tucker and Wright 1990). During this second stage, the biostrome developed under relatively non-turbid-water and uniform conditions (low, clay-poor sediment input). The uniform growth patterns of the colonies, lacking marked discontinuities, suggest a monotonous environment, not prone to marked cyclical variations of environmental parameters such as light, temperature (?) and water turbidity (e.g., see Kershaw and Young 1997). Finally, in the latest stage, environmental conditions changed and led to the onset of clay-rich muddy sedimentation, which very likely was responsible for the abortion of the biostrome and finally buried it. This three-step evolution occurred several times giving rise to the repetition of the sequence of facies formed of the basal skeletal bed containing reworked coral colonies, the biostrome and the capping muddy interval plus the muddy matrix surrounding the biostrome.

**Facies F:** mud mounds

**Description**

This facies forms small (0.3–0.8 m thick and ~1–4 m wide) mounds, and more rarely flat, bed-like bodies. They display ragged margins and are encased within intervals formed of facies C (Figs. 3 and 4B, C and F); only one very small and low-relief mound appears within an interval of facies B. The individual mounds are isolated from each other (Fig. 4F) or grouped either as laterally offlapping or as vertically stacked bodies (Fig. 4C). These
mound structures display discontinuous and irregular reddish shale/marlstone partings that can be traced into the ragged margins and connect with the surrounding marly rock (facies C). The mounds mostly consist of a reddish and greenish micrite containing a relatively abundant macrobiota (<25%), and displaying different types of submillimetric to millimetric cavities (cf. Schmid et al. 2001). The micrite is structurally heterogeneous, with several sediment generations (polymuds of Lees and Miller 1985) revealed by differences in colour or texture. Under the microscope, three types of carbonate sediment are distinguished, namely types 1, 2 and 3 (Fig. 5D–F). Types 1 and 2 are both micrite, and their mutual relationships suggest that they are coeval. Type 1 is a dense and dark micrite, with a homogeneous appearance, although it is locally peloidal (Bathurst 1975; Fig. 5D). It forms masses, which can have a positive relief with steep to vertical margins, surrounded by type 2 micrite. Type 2 micrite is a lighter coloured, homogeneous micrite. Sparse minute bioclasts are present in both types of micrite, although they are more abundant in the lighter homogeneous type 2 micrite. Type 3 is a microsparitic material with scarce minute bioclastic fragments (Fig. 5D–F). Cross-cutting relationships show that it is a later sediment than types 1 and 2 (see below).

The macrobiota of this facies does not differ significantly from that of the reddish marlstones and shales (facies C) in which the mud mounds are encased, and, mainly in the case of the corals, shows a variable degree of bioabrasion (microborings). The most prominent organisms are fenestellids and platy fistuliporids (Fig. 5E). They are encrusted by the dense and dark micrite masses (type 1 micrite) and, in turn, encrust the micrite types 1 and 2 and the grains. Other organisms are branching bryozoans, disarticulated plates of crinoids and blastoids, tabulate corals, rare brachiopods or their disarticulated valves and sponge remains (Fig. 5D–E). Tabulate corals, which markedly differ from those of the coral biostrome facies, display a high diversity at both species and genus levels and mainly comprise *Crenulipora*, a ramose favositid; other subordinate forms are *Hamarilopora* and reptant and erect auloporids (*Schlueterichonus* among others). Scarce sponge spicules are present whereas some samples taken contain...
whole sponges (Fig. 5D). Remarkably, sponges exhibit a chambered structure indicative of a sphinctozoan nature. Spicules are partly preserved revealing hexactinellid arrangement in lattice formation. Although a detailed taxonomic description of the sponge material from the Colle mounds has still to be done, this is the first record of hexactinellid sphinctozoa from the Devonian (Krautter pers. comm.). Until now, only a few representatives of hexactinellid sphinctozoa are known, exclusively derived from Upper Triassic and Jurassic strata (Senowbari-Daryan and García-Bellido 2002).

Apart from intraparticle porosity and very rare shelter porosity, the rock contains three different types of millimetric cavities, namely A, B and C (Fig. 5D–F). None of these cavities can be considered as typical stromatactis porosity, although some cavities of type B share all characteristics of stromatactis except for network and/or isopachous cements, and can be termed as stromatactis-like cavities according to terminology by Matyszkiewicz (1997) and Schmid et al. (2001). Type A consists of elongate pipe-like cavities found in the micrite of types 1 and 2 (Fig. 5D and E). This type of cavities is filled by the type 2 micrite. It is thought to result from burrowing in a soft sediment. Type B comprises elongate or irregular pores in type 1 and 2 micrites and is sealed by type 3 sediment (microsparitic material) which completely fills the pores or just floors them giving rise to geopetal structures (Fig. 5D–F). The elongate cavities are burrow-like (e.g., see Fig. 5D and E), whereas the irregular pores (e.g., see Fig. 5D and F) are larger (up to 1 cm) and display scalloped margins that truncate older sediment (micrite types 1 and 2) suggesting an origin by, at least partially, dissolution (cf. Lees and Miller 1995). Nevertheless, scalloped margins have also been interpreted as indicative of sponge-boring activity (Schmid et al. 2001, see their Fig. 17). Type C cavities are elongate burrow-like pores in type 3 sediment filled with the same type of sediment being only distinguished by subtle variations in colour (Fig. 5D and F).

Three generations of cement are found in the cavities of these facies. The first generation is a thin and poorly developed isopachous crust of inclusion-rich fibrous calcite. This generation is found in intra-skeletal pores and in type B porosity. Cross-cutting relationships reveal that these cements started growing during the final stage of the microsparite sedimentation and continued after its end. The second generation is made of inclusion-rich columnar crystals of calcite to slightly ferroan calcite with undulose extinction and is found in some intraparticle pores and in type B porosity. The third generation is a mosaic of clean, slightly ferroan to ferroan calcite and is poorly developed occluding the remnant voids in intraparticle and type B porosity.

**Interpretation**

The textural features and geometry of these mounds are comparable to those of mud mounds formed of microbial boundstones (see Lees and Miller 1995; Monty 1995; Pratt 1995). Following these authors, and despite the absence of fossilized microbial filaments, which can be attributed to filament degradation or to the small size of microbes, the dark micrite of type 1 is interpreted to result from cyanobacterial activity, judging from its dark colour, occasional peloidal texture, and the erect and steep-margin morphology of the masses that it forms, which is suggestive of cohesiveness. The lighter-coloured micrite of type 2 is thought to represent a sediment deposited mechanically. This type of sediment likely originated within the mud mound proper, given the terrigenous mud-rich environment of the mounds. This interpretation of a local provenance of the carbonate mud in the mud mounds is a common theme found in the literature and is based on the same criteria (see Bosence and Bridges 1995: 5). It is considered that degradation of the organic masses of type 1 micrite and of the skeletal grains by grazing and boring organisms would have led to this “mechanical” micrite. This interpretation is backed by the conspicuous early burrowing activity that is recorded by the type-A pores. The scattered sphinctozoan sponges probably produced some additional mud (automicrite) during degradation. Nevertheless, a partial external source of carbonate mud cannot be ruled out, since also carbonate mud is found outside the mud mounds. Apart from the microbial communities, it seems that the organisms that played a significantly active role in the mud mound stabilization were fenestellids and fistuliporids, by binding one another, the microbial micritic masses, and the available sediment (lime mud of type 2 micrite, clay particles and skeletal grains). In some instances, these bryozoans are found to be roofing type-B cavities, which suggests that they might have encrusted a soft body that later disappeared, although, in some of these cases, it seems that the bryozoans could have grown downwards from the cavity roof. The other organisms, chiefly crinoids and blastoids, are thought to have mainly played a passive role by providing grains, i.e., their whole or disarticulated skeletons, to the deposit. The small number of coral colonies suggests that they did not exert a significant baffling or binding role. The described biota are fairly similar to those of Devonian examples from Algeria (Wendt et al. 1997), Kess-Kess mounds of Morocco (Brachert et al. 1992) and the suggested encrusting role of bryozoans has also been claimed in Early Devonian examples from the Clifton Saddle (west-central Tennessee, USA; Gibson et al. 1998).

The sediment forming the mud mounds soon seems to have been semi-lithified as the features of the type-B porosity suggest. This second generation of porosity could have a variable origin (see a detailed discussion in Monty 1995). It seems to have resulted from dissolution and burrowing, maybe accompanied by current driven winnowing of unconsolidated sediment beneath a lithified crust (cf. Schmid et al. 2001). In other instances, it seems to represent growth-framework or shelter cavities variably modified by the former processes. Finally, it could also result from the degradations of pre-existing soft-bodied organisms (cf. Lees and Miller 1995; see above the comment about the bryozaoo-roofed cavities). Type-B porosity is filled by the microsparitic material (type 3 sediment). Bur-
rowing activity was a long lasting process over the mound life and the last burrowing stages gave rise to the burrowing structures that cut and are filled by the type 3 sediment. The absence of pore-filling clay suggests that the three stages of burrowing took place before mud-mound abortion and burial by the clay-rich deposits of facies C. In addition, type-3 sediment pre-dating the isopachous fibrous calcite cement indicates that it was deposited before a substantial early marine-phreatic cementation took place. It is worth mentioning that burrowing represented by type-B and, mainly, type-C porosity makes these examples different from other deep-water mud mounds, which are generally devoid of infaunal soft-sediment dwellers (cf. Pratt 1995: 107), but similar to the Silurian examples from Anticosti Island, Canada, described by Schmid et al. (2001), which also show analogous infaunal organisms.

The ragged margins and the shaly partings of the mud mounds suggest that mound development was essentially coeval to sedimentation from fallout of clay and carbonate mud particles forming the surrounding deposits (facies C), and they would fall within the “sediment-continuum” mound category of Schmid et al. (2001). These two features indicate that clay-fallout rate and/or carbonate production rate varied in intensity, possibly in relation to some high-frequency external factor. It is interpreted that with high clay-fallout rates mound growth probably was somewhat inhibited and clay veneered the mound surface, resulting in the shaly partings. On the contrary, during moments of reduced clay-fallout, mounds would have grown faster encroaching over the surrounding marly sea bottom.

Given the inferred temporal relationships between mud mound development and off-mound sedimentation, it follows that the environmental conditions were the same for the mud mounds and for the surrounding facies C sea-bottom, and thus, mud mounds formed in a low energy, relatively deep environment. The occasional presence of thin skeletal beds suggests that this environment could have been close to the storm wave base. Several authors (e.g., Wendt et al. 1997, 2001) have claimed a deep-water setting, being close to the storm wave base. Several authors (e.g., Wendt et al. 1997, 2001) have claimed a deep-water setting, being close to the storm wave base.

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Allocyclical controls on buildup development

Sequence stratigraphy framework

Individualization and identification of depositional sequences and sequence boundaries must be taken with caution, mainly in single well logs, cores and outcrops. Nevertheless, even in these unfavourable cases some diagnostic criteria can help in the discrimination of depositional sequences. These are abrupt basinward shifts in facies, and changes in the mode of vertical stacking of parasequences or higher-order depositional sequences (see van Wagoner et al. 1990). These criteria can be easily applied in the Colle outcrop where the vertical and lateral relationships among the facies above described permit to group the studied interval into a series of units bound by sharp surfaces (see Figs. 3 and 6A), whose general features are summarized
in Fig. 7. Although, at small-outcrop scale, the basal sharp surfaces are identifiable as erosional only in some instances (e.g., base of unit 4 in Fig. 3; see also Fig. 4D), when they are traced over a larger area (see Fig. 6A), they clearly show truncation and onlap. Additionally, their erosional character is revealed by the presence of coral clasts with a diagenetic history different from that of the hosting sediment (see the description and interpretation of facies D). These erosive surfaces display a metric-scale relief (Figs. 4D and 6A), show no evidence of subaerial erosion and are located at the inflection points between coarsening-upward and fining-upward trends (Fig. 3). The units bound by these sharp surfaces have a well-defined internal structuring and vertical stacking of facies (Fig. 7). Each unit sharply commences with a well-developed body of skeletal limestones (facies D; Fig. 4C) that laterally passes into a skeletal-bed-rich interval of facies B or C (Fig. 4D; see also Fig. 6A). It is important to stress that, as discussed above, these skeletal limestone bodies show evidence of resulting from the amalgamation of the deposits of successive storm events. Thus, they record a time span characterized by frequent storms affecting the sea floor, and should not be considered each as the geologically instantaneous deposit of a single storm. This basal interval fines upwards and passes into a middle interval composed of shales/marlstones belonging mainly to facies B or C, although the middle part of this interval may be formed of facies A (Figs. 3, 6A and 7). This middle interval forms the remaining of the unit or it passes into an upper interval of alternations as the shales/marlstones start containing skeletal limestone beds with a coarsening and thickening upward trend (Figs. 3, 6A and 7). The skeletal bodies of the basal interval would be storm-related deposits laid down in a ramp environment above the storm wave base (see above). Their occurrence, abruptly overlying a mud-rich interval instead of forming the top of coarsening upward intervals (i.e., instead of evolving from an underlying interval progressively richer in limestone beds), indicates that the relatively shallow-water and high-energy (storm influenced) environment sharply succeeded the deeper and more tranquil environment of deposition of the mud-rich facies (essentially a deep ramp below the storm wave base). That is, these deposits represent a sharp basinward shift of facies. Consequently, they are interpreted to record a relative sea-level drop accompanied by a significant submarine erosion of the substratum, judging from the relief of the basal erosional surfaces and the lack of evidence of subaerial exposure, and represent lowstand or early transgressive deposits of a depositional sequence. The gradational upwards passage of these packages into the middle interval of muddy facies (mainly facies...
cies B or C; see Fig. 4C) indicates a gradual decrease of energy, very likely tied to a gradual deepening of the environment, and thus this middle interval would constitute the transgressive system tract of the depositional sequence. In some depositional sequences, this middle interval passes upwards into the coarsening-upward upper interval of alternations, reflecting a gradual increase in the energy of the environment likely due to shallowing. This upper interval clearly corresponds to the highstand system tract, although it cannot be precluded that the upper part of the middle interval could include the distal counterparts of the highstand system tract. This vertical arrangement of facies, and the interpretation that we propose, contrast with those given by Ruhrmann (1971), who studied the same stratigraphic interval in nearby localities of the Esla Unit (located in the same thrust sheet but also on the relative autochthon; see Fig. 1B), and by Keller (1988). Both authors described this interval as formed of coarsening upward sequences comprising mud-rich deposits, comparable to our facies B and C, capped by skeletal limestones (our facies D) and concluded that these are shallowing upward sequences related to tectonic uplift. In addition, Keller (1988) strikingly mentioned that frequently the skeletal limestones have a sharp, channel-like base and interpreted these limestone bodies as channel fills. Because of the vertical organization that we have described and taking into account that no channel s.s. features have been found (i.e. we have not found evidence of erosional conduits acting as ways of transport for long periods of time during which only coarse lag deposits are left), we consider that, in the case of the Colle deposits, our interpretation is more plausible. On the other hand, Keller’s interpretation fails to explain why channels would suddenly form in the relatively deep and low energy settings represented by the underlying sediments.

Our interpretation is also backed by the mode of vertical stacking of the depositional sequences of the studied interval. The diagram of Fig. 6B shows that they form a set of sequences that display a systematic vertical variation of (1) sequence thickness, (2) importance of the basal skeletal limestone packages and (3) distribution of and mutual relationships among the three muddy facies (barren shales of facies A, fossiliferous greenish-grey shales/marlstones of facies B and fossiliferous reddish shales/marlstones of facies C). We speculate that these variations result from the interplay between two or more orders of cyclicity. Following Keller and Grötsch (1990) and Keller (1997), the La Vid Group was deposited during two 3rd-order cycles, and the highstand deposits of the upper cycle (Valporquero Formation) record a minor lowstand (4th-order cycle?), which is represented in the studied interval. It thus follows that the depositional sequences described in the studied interval may be attributed to 5th-order cycles.

The portions of the Valporquero Formation that underlie and overlie the studied interval crop out very patchily, so little can be said about them. Nevertheless, Leweke (1982) found that they are mainly composed of shales, corresponding to his Lower Shale and Upper Shale members, respectively. The succession immediately below the studied interval comprises several tens of metres of barren grey shales, that, mainly in the upper part, contain subordinate and thin intervals of skeletal limestones and of fossiliferous shales (see also Stel 1975). All these rocks are similar to the facies A, B and D described above. Atop the studied interval, the Valporquero Formation continues with barren grey shales (attributable to facies A), which occasionally host laterally discontinuous skeletal limestone beds and packages of beds (similar to facies D). Judging from these stratigraphic features and from the facies distribution across the 5th-order sequences of the studied interval, it seems clear that the second 5th-order sequence (sequence #2 in Fig. 6A) represents the maximum regression linked to the base of the upper lower-order (4th-order) cycle (depositional sequence II in Fig. 6B). This second 5th-order sequence displays the most proximal character containing the thickest lowstand deposits and the best-developed highstand tempestites. The successive overlying depositional sequences thin and fine upwards suggesting that they form a retrogradational sequence set, which would belong to the transgressive system tract of the lower-order (4th-order) depositional sequence II. The overlying poorly exposed interval of the Valporquero Formation precludes drawing further conclusions about this topic. Nevertheless, this shaly interval could represent at least a part of the highstand system tract of the 4th-order sequence, since a higher clay input has been suggested to take place in the deeper basinal areas during rising sea-level periods (Pasquier and Strasser 1997; Elrick and Snider 2002).

Within this scheme, the reddish marlstones/shales of the facies C seem to occur in the 5th-order sequences associated with the maximum regression/early transgression of the 4th-order sequences (see Figs. 6 and 7). Notably, a similar location within a sequence stratigraphy framework has been postulated for marine red beds by Loydell (1998). This author points out that marine red beds appear to have been deposited during a minor sea-level fall immediately after a period of very high sea-level. In our case, this would correspond with the 4th-order sea-level fall that takes place during the 3rd-order high sea-level. In turn, the barren shales of the facies A occur preferentially in the lowermost and uppermost 5th-order sequences of the studied interval (Fig. 6). These deposits, which are interpreted to represent the deepest environment recorded in the area (Figs. 7 and 8), would appear in the transgressive and highstand system tracts of those 5th-order sequences that belong to the late transgressive and at least the early highstand parts of the 4th-order sequences (Figs. 6B and 7; see also the comments in the previous paragraph).

The sequential setting of the *Synaptophyllum* biostromes and the mud mounds

Given the inferred sequence stratigraphy framework discussed above, it seems clear that the *Synaptophyllum* biostromal intervals and the mud mounds occur in specific positions within depositional sequences and thus that their development was controlled, at least partially, by the factors that determined the different orders of cyclicity and
their interplay (Figs. 7 and 8). Nevertheless, the few occurrences of Synaptophyllum biostromes, restricted to the lower part of the section (Fig. 3), imply that conclusions about them must be taken with caution. On the contrary, mud mounds are relatively common in the studied succession, where they show a close relationship to facies C (Fig. 3), and this makes conclusions about them to be more reliable. Synaptophyllum biostromes are probably related to the early transgression of a 5th-order cycle prior to the maximum regression in the boundary between two 4th-order cycles (Figs. 7 and 8A–C). On the contrary, mud mounds seem to be related to late transgressive or highstand episodes of 5th-order cycles during the late lowstand and early transgression of a 4th-order cycle (Figs. 7 and 8D and E).

During the 5th-order early transgression, progressive deepening, coupled with relatively non-turbid-water conditions led to the onset of Synaptophyllum biostromes in an environment below and close to the storm wave base (Fig. 8B). Later on, as clay supply was progressively re-activated the biostrome was killed (Fig. 8C). This process was repeated several times through time (see unit 1 in Fig. 3). This repetition could be the result of a higher-order (6th-order?) cyclicity or merely the result of recurrent storm activity during the early transgression. Finally, continuing transgression brought the area to a deeper environment in which barren shales were deposited.

Later on a 5th-order cycle framework, during the sea-level rise and highstand (transgressive to highstand systems tracts), when the environment in the area of study was deeper and of lower energy compared to that existing during the development of the Synaptophyllum biostromes, and clay supply had been fully re-established, mud mounds developed (Fig. 8D and E). This situation, which can clearly be deduced from the sequential arrangement of the succession, notably coincides with that inferred by Elrick and Snider (2002) for the deep-water Cambrian mud mounds of House Range (Utah, USA). Nevertheless, as said above, mud mounds do not occur in the same 5th-order sequences as the Synaptophyllum biostromes, suggesting that tuning between 4th- and 5th-order cyclicity played a role in determining mud mound development (Fig. 7). The shale partings of the mud mounds also record high-frequency variations in the environmental conditions. These variations in the clay fallout rate could be of the same frequency as the repetitions shown in the Synaptophyllum-bearing deposits (unit 1, Fig. 3), composed of several packages, each one consisting of a skeletal bed, overlain by a biostrome episode, which in turn is followed by a muddy interval, and be related to 6th-order cyclicity.

Controlling factors of cyclicity

The several orders or cyclicity here described are governed by different driving mechanisms. With respect to the lower-order cyclicity, as Johnson et al. (1985) pointed out, eustatic sea-level variations during greenhouse periods, like most of Devonian times, must be of tectonic origin and related to variations in mid-ocean ridge volume or, in the case of the third-order cycles here mentioned, to mid-plate thermal uplift and submarine volcanism. With respect to the higher orders of cyclicity, available data both on the stratigraphic interval and on the time span involved are very limited. Classically, it has been interpreted that Palaeozoic sedimentation in the Cantabrian Zone and in the neighbouring West Astur-Leonese Zone was largely governed by syn-sedimentary, extensional or strike-slip tectonics (e.g., see de Sitter 1962; Evers 1967; Kullman and Schonenberg 1978; see also a review in Aramburu et al. 2004), frequently accompanied by volcanic activity with a variable importance. Regional tectonics resulted in the individualization of uplifted and subsiding blocks, the former acting temporarily as shallow, starved areas with condensed sedimentation, or as emerged landmasses delivering sediments to the latter. In the case of the studied deposits of the Valporquero Formation, several authors have interpreted that their occurrence and location are the product of tectonics that led to the individualization of shallow areas where mainly storm-related processes were responsible of laying down high-energy deposits and also of taking skeletal debris and transporting them elsewhere (Ruhmann 1971; Stel 1975; Keller 1988; Keller and Grötsch 1990; Keller 1997). Specifically, Stel (1975) invoked a palaeohigh located in the South (Narcea High), where sediment was produced and then transported as skeletal debris northwards to the sectors of the basin where the Colle section is nowadays. The other authors invoked a southward transport of skeletal debris, with a role of a palaeohigh located in the north (Pardomino High) and a pattern of facies distribution following another palaeohigh related to the so-called Sabero-Gordón Line.

Despite this, there is a broad agreement on the general tectonic model for pre-Variscan times, some discrepancies exist about several pre-Variscan features and their bounding faults (e.g., León Line, Sabero-Gordón Line, Narcea High, Pardominos Ridge; see Fig. 1B and C) mainly during the Devonian. Disagreement concerns the extent, age, exact location and even existence of these features, as well as their potential role during Variscan (Carboniferous) times. Some authors claim that some of them are actually Variscan or post-Variscan features (e.g. see Marcos 1968; Aller 1986; see also the debate about the meaning of some Variscan structures maintained by Nijman and Savage 1989, 1991; Alonso et al. 1991). Even if some of these controversial features were inherited structures, they would have been subsequently modified to such an extent that considering them the same structures with the same position as their potential precursors would be an extreme oversimplification. For instance, the so-called Sabero-Gordón Line, interpreted to be a pre-Variscan feature, strikingly cross-cuts several Variscan thrust sheets (see Fig. 1B and C; see also Fig. 1 of Keller 1988 and Fig. 2 of Keller 1997). Also, the Narcea and Pardominos highs fit the present-day outcrops of lower structural levels of some Variscan thrust sheets exhumed after the emplacement of the Narcea Antiform and the Esla Nappe, respectively (Fig. 1; see Pérez-Estaún et al. 1988; Alonso 1985 for a structural explanation of these two features).
Fig. 8 Schematic cartoons showing the inferred timing of coral biostrome and mud-mound development along an idealized 5th-order cycle (see text for details). Compare with Figs. 6A and 7.
Given these facts, and thus being cautious about the identification and location of any potential palaeohigh, we conclude that regional tectonics must have played a significant role in the facies distribution and the vertical organization of the studied deposits. The relative sea level variations could reflect a varying balance between subsidence and sediment-input rates, although we disagree with Ruhrmann (1971), who interpreted that the shallow-upward sequences purely resulted from tectonic uplift. Apart from regional tectonics, other factors could have played a role, by themselves or interacting with tectonics. Cyclicality could be the result of astronomically induced climatic variations within the Milankovich frequency band or of shorter period (e.g., see Keller’s 1997 discussion on the onset of the carbonate sedimentation of the La Vid Group), involving or not the potential existence of continental polar ice caps (see comments in Pasquier and Strasser 1997; Immenhauser 2005, and references therein). In this sense, Elrick (1995) and Elrick and Hinov (1996) interpreted the high-frequency cyclicity in Middle Devonian rocks of the eastern Great Basin (North America) to be of glacio-eustatic origin and, in the case of the highest-frequency cycles, due to millennial climate fluctuations not related necessarily to waning and waxing of ice caps.

Conclusions

The studied interval of the Valporquero Formation comprises a set of detrital and organically bound deposits laid down in a mixed, terrigenous/carbonate ramp. The detrital facies are skeletal lime pack- to grainstones deposited above the storm wave base, and mud-rich deposits (barren greenish grey shales, fossiliferous greenish-grey shales and marls, and fossiliferous reddish shales and marlstones) with some thin-bedded skeletal limestone beds deposited in a low energy environment below storm wave base. The organically bound facies are of two types: colonial rugose coral biostromes forming patches a few metres across and mud mounds a few metres wide and up to a metre thick. Both were deposited below the storm wave base, although the latter seem to be of deeper water than the former.

All those facies stack vertically in a predictable manner. Analysis of this vertical structuring of facies suggests that it reflects a series of 5th-order sequences that developed during a minor (4th-order) sea level fall occurring in the middle of a major (3rd-order) highstand. The style of vertical stacking of the 5th-order sequences and their facies composition is thought to result from the interplay among these three (3rd, 4th and 5th) orders of cyclicity. This interplay seems to have been a major factor in determining the occurrence of the two types of organic buildups, the colonial rugose coral biostromes and the mud mounds. The former seem to occur in 5th-order sequences formed during the late 4th-order highstand, whereas the latter seem to occur in 5th-order sequences developed in the late lowstand to early transgressive 4th-order system tracts. In addition, detailed facies relationships permit to decipher the timing of development of build ups with respect to the surrounding facies.

The colonial rugose coral biostromes take place on skeletal limestone beds and the colonies are surrounded by fossiliferous grey shales/marlstones. Field relationships and detailed scale (thin section) studies suggest that biostromes developed on a stable granular substrate (skeletal limestone beds) that was deposited during the lowstand and early transgressive system tracts. Before clay input resumed later in the transgression, extensive colonization of the sea floor by colonial rugose corals, alveolitids, favositids and stromatoporoids took place in non-turbid, possibly sporadically agitated waters. This resulted in low relief patches (biostromes) tens of metres wide, whose development was then aborted as clay sedimentation commenced.

The mud mounds are encased in fossiliferous reddish shales/marlstones and have a uniform facies composition that suggests uniform ecological conditions over the mud mounds lifetime. They are dominated by microbial micrite, and fenestelid and fistuliporid bryozoans. Other metazoans (sponges, corals and crinoids) present seem to have played a limited role, acting as grain suppliers. Distinctively, these mud mounds display common burrowing structures and lack stromatolitic cavities. In contrast to coral biostromes, mud mounds do not seem to have any particularly particle-rich substrate requirement to develop, although their muddy substrate contains an appreciable amount of skeletal material. Mud mounds formed during transgression to stillstand and they are essentially coeval with sedimentation by settling from suspensions of the clay and carbonate mud that form the encasing shale/marlstone deposits (facies C). Nevertheless, detailed structuring of mud mounds suggests that clay-fallout rate varied with time giving rise to periods of mud mound growth and encroaching on the substrate and periods of mud mound inactivity and clay veneering of its surface. In any case, it remains obscure what factors led mud mounds to be so clay-free if compared to the adjacent sediments deposited some metres away. Carbonate-mud sedimentation rates outpacing low clay-fallout rates and, possibly, relief-induced sheltering of the mud mounds from thin hyperpycnal clay-laden flows are the most likely mechanisms.

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