Post-workshop field trip guide of the XI International Ichnofabric Workshop

Museo del Jurásico de Asturias (MUJA)

Colunga, Spain, 1st–5th July 2011

XI International Ichnofabric Workshop

Spain
Colunga (Asturias)
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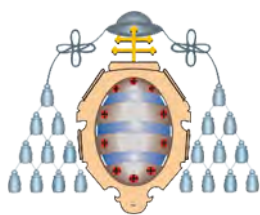
University of Oviedo

Post-workshop field trip guide of the XI International Ichnofabric Workshop

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XI International Ichnofabric Workshop

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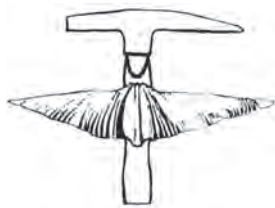Dinokinétics
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Figure 1. Location of the field stops. Stop 1: Vega beach. Stop 2: Villar sea cliffs. Stop 3: El Puntal sea cliffs. Stop 4: La Griega beach

Introduction to the Jurassic of Asturias

In the Asturias, region the most spectacular and best conserved Jurassic rock outcrops extend almost continuously along a narrow section of coast between the Cabo Torres, in Gijón, and the Arra beach, situated about two kilometers to the east of Ribadesella (Fig.2). The municipalities that make up this littoral sector are, from west to east: Gijón, Villaviciosa, Colunga, Caravia and Ribadesella. The Jurassic outcrops are part of the so-called Gijón-Villaviciosa Basin (Ramírez del Pozo, 1969), whose western end is represented by the Veriña fault, located a few kilometres west of Gijón, while to the east it ends at the Ribadesella fault (Figs.2,3), coinciding there with Arra beach. This vertical fracture of east–west orientation, which puts the Jurassic outcrops into contact with Carboniferous limestones, began to develop during the Cretaceous, allowing conservation of the Jurassic series in the northern fallen block, and was reactivated in the Palaeogene as an inverse fault during the tectonic inversion of the compressive alpine stage that led to the creation of the Cantabrian Range (Valenzuela *et al.*, 1986; García-Ramos & Gutiérrez Claverol, 1995 a,b; García Ramos *et al.*, 2002, 2004, 2006 a,b; Alonso *et al.*, 2009).

The Jurassic rocks of Asturias can be grouped into two high-order lithological units (Valenzuela *et al.*, 1986). The lower one is predominantly made up of carbonate rocks of littoral origin (Gijón Formation) and open marine origin (Rodiles Formation). The upper one mainly comprises siliceous rocks of continental origin (alluvial fan or fluvial) in the formations of La Ñora and Vega, and restricted marine and coastal (deltaic), respectively represented by the Tereñes and Lastres formations (Fig.4).

The Jurassic record of Asturias begins with a succession of limestones, dolomites and marls (Gijón Formation) that originated on a low and irregular coast, rich in carbonate muds and evaporites (sabkha). These lithologies also include some intervals of metric thickness, constituted by calcareous breccia. Their origin is related with processes of dissolution of gypsum layers intercalated among highly fractured limestones, which caused the fragmentation and collapse of these latter, generating stratiform accumulations of collapse breccias**.**

Later on, a gradual rise in sea level left a major part of the region submerged under the sea, with a depth that at times would have been over 100 m. The Rodiles Formation has two clearly distinct parts: the first meters are made up of nodular limestones with some thin marly layers, representing the proximal part of a carbonate ramp; in the rest of the formation, the layers of limestones and marls present a tabular geometry adopting a rhythmic character that represents the middle and external part of the ramp.

At the beginning of the Late Jurassic, a drastic change in the landscape took place as a consequence of the activity of several faults within an extensional regime, which eventually produced the uplift and emergence of part of the territory. As a result of this tectonic activity, the sea —which covered a good great of Asturias during the Early and Middle Jurassic— retreated abruptly, giving rise to new coastlines and emerged lands that would soon be colonized by dinosaurs and other coetaneous vertebrates such as turtles, crocodiles, lizards, fish, flying reptiles, etc. The most recent age of the carbonate marine succession truncated erosively by the terrigenous material of the Late Jurassic, which fills a unconformity, is Lower Bajocian.

This tectonic activity, controlled by extensional faults, represented the first stages of a period of rifting that culminates during the Lower Cretaceous. It generated a relief in the southwestern part of the region, within the so-called Western Asturian-Leonese Zone, whose erosion provided the first contributions of terrigenous material to the basin.

Once the carbonate successions of marine origin of the Lower Jurassic had emerged, a process of karstification took place, leading to the formation of decalcification clays, collapse breccias, and palaeovalleys excavated in the underlying calcareous rocks.

Shortly thereafter the first terrigenous materials (gravels and sands) began to accumulate in the eastern part of the region. They were of alluvial origin, proceeding from the erosion of the recently created palaeorelief, and initially filled in the palaeovalleys and karstic cavities up to colmatation.

The successions, of up to 60 m in thickness , are formed mainly by siliceous conglomerates with small intercalations of sandstones, and red mudstones with calcareous palaeosols, ordered vertically into meter-scale finning-upward cycles. They constitute the La Ñora Formation, of alluvial origin.
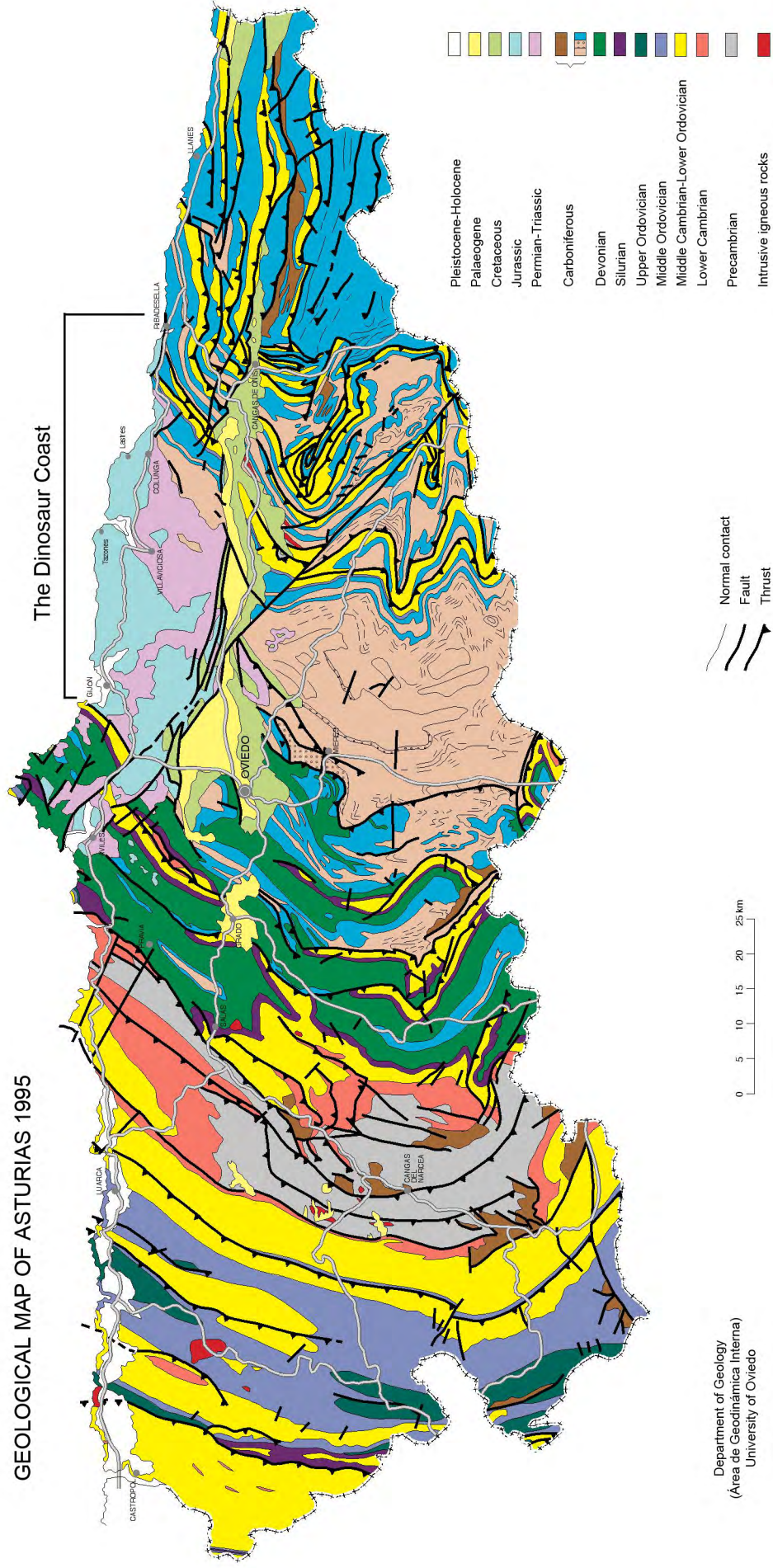
Figure 2. Geological map of Asturias. Department of Geology. University of Oviedo

These materials evolve laterally toward the northeast to alternating white, grey and reddish sandstones, with red mudstones and some conglomeratic beds (Vega Formation), likewise ordered in minor fining-upward cycles within a major cycle of the same character. They represent fluvial deposits formed by ephemeral and highy sinuous streams separated by inter-channel areas with calcareous palaeosols (caliches) and sporadic ponds with high microbial activity which gave rise to micritic grey limestones with oncoids. These ponds were partly fed by a number of freshwater carbonate-rich springs coming from faulted zones that affected to the limestones and dolomites of the Lower and Middle Jurassic rocks (García Ramos *et al.*, 2010).

Another rise in the sea level retreat the coastline toward the continent, meaning the coast was then situated in the interior of the territory that Asturias occupies today, giving it a northwest–southeast trend that is quite different than its current profile. In the front of this coastline, hardly influenced by the tides and waves except during occasional storms, a restricted shallow sea developed (shelf lagoon), separated from the ocean by a threshold or barrier of tectonic origin that impeded the entry of marine fauna from the outside. In the bottom of this small sea, a thick sequence of calcareous dark muds rich in organic matter and brackish invertebrate fauna accumulated (bivalves, gastropods, ostracods) now converted into shell beds (Tereñes Formation). This sea also served as a refuge for a multitude of crocodiles, turtles and fishes.

The coast that was traversed by fluvial-dominated deltaic systems was also frequently visited by dinosaurs and other reptiles that left their footprints into the sediment. The alternating lutites, marls and sandstones that accumulated on the former littoral now constitute the lithostratigraphic unit known as the Lastres Formation, culminating the Jurassic succession in this region.
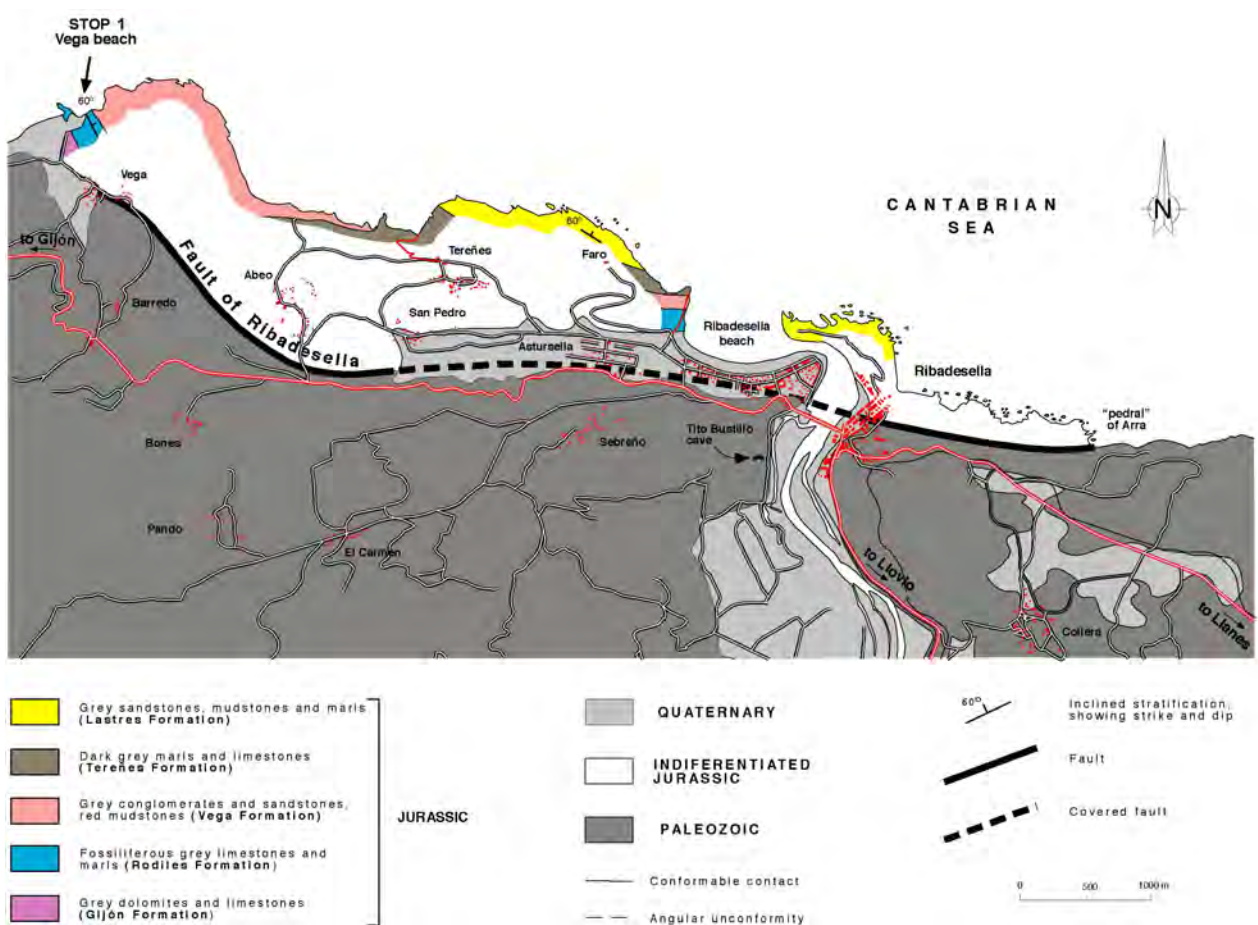
Figure 3. Geological map of the region close to Ribadesella, showing the Ribadesella fault (coarse black line) and the situation of the Stop 1, Vega beach
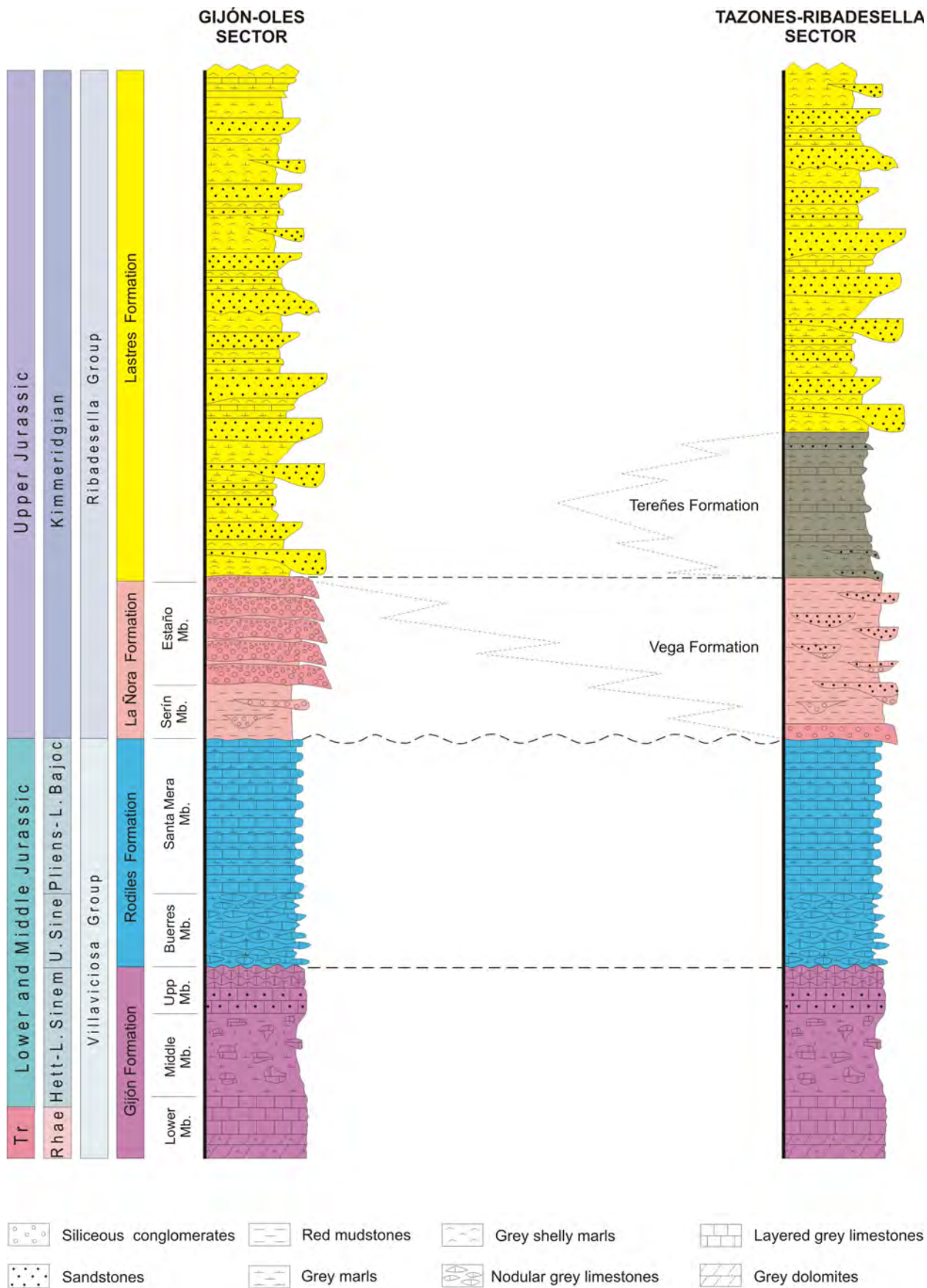
Figure 4. General stratigraphy of the Jurassic rocks in Asturias. Not to scale. Modified after García-Ramos *et al.*, (2006)

Stop 1. Vega beach (Ribadesella). Origin and ichnofabrics of Lower Jurassic (Pliensbachian) marl-limestone rhythmites

Location and geological setting of the Vega beach

The small coastal village of Vega, situated some 8 km west of Ribadesella (Fig.1), lends its name to one of the biggest beaches in Asturias, nearly 2 km long, which attracts many tourists to the area, particularly in the summer season. By the central part of the beach there is a field of dunes, considerably affected by anthropic activity.

Deserving special attention is the karstic system of Entrepeñas, of great beauty, included by the Government of the Principality of Asturias in the system of Natural Protected Spaces under the figure of Natural Monument (Fig.5). This karst is not developed upon carbonate rocks as is usually the case, but rather on quartzites of Ordovician age, strongly wheathered by an intensely fractured zone subjected to hydrothermal processes during Permian times (Gutierrez Claverol *et al*., 2008).

Also noteworthy in the beach area are the vestiges of old fluorite mines, now abandoned. The Permian mineralization was partly stratiform in character and partly associated with faults having a predominantly NW-SE orientation (García-Ramos & Gutiérrez Claverol, 1995; Gutiérrez Claverol *et al*., 2009).

Figure 5. The "karstic" system of Entrepeñes (Natural Monument by the Asturian Government) developed into the Ordovician quartzites. Access to the Vega beach

The rhythmic succession in the Jurassic of Asturias

Introduction to the rhythmic marl-limestone series

Rhythmic alternations of limestones and marls has been described from every geological time periods in the Phanerozoic. Although such interbedding may originate in many diverse environments, its development takes place almost exclusively in warm water environments of tropical and subtropical climates, because these are potentially suited for the aragonite formation in the proximity to a carbonate factory, with limited supply of terrigenous sediments. (Westphal & Munnecke, 2003; Westphal, 2006).

It has been understood for some time (Bathurst, 1971) that the limestone beds have undergone as many different diagenetic processes as well as the marl interbeds. These diagenetic processes have exaggerated the initial differences present at the moment of deposition, (Ricken, 1986, 1987) not only in the composition of sediment, but also in the proportion and diversity of the fossil content, geochemical parameters, and other physical properties (Westphal, 2006).

Although Ricken (1986, 1987) interpreted that limestone bed cementation had taken place during a stage of deep burial, it is now understood that that interpretation is disputable. The associated macrofauna and ichnofossils show no evidence of deformation due to pressure, suggesting that commonly the cementation of limestones took place prior to compaction. This phenomenon appears, nevertheless with some exceptions, as we will see later on.

These rhythmic successions are characterized by the selective dissolution of aragonite, especially within the marly interbeds. This in turn, become unsteady due to a lessening of microbial pH control resulting from the acids liberated during decomposition of the organic matter into the sediment surface. This phenomenon produces a lowering in the initial volume of the marl, which acts to differentially compact the limestone cemented with calcite as a result of the pressure due to the sediment load (Westphal, 2006).

With respect to the cement source required to form the limestone beds, it is necessary consider that the initial porosity of micritic sediment is quite elevated, which requires a large volume of cement in order to fill the pores, but at the same time these micritic deposits have an extremely low initial permeability which inhibits fluid flow. Therefore, the calcium carbonate that will serve as cement has to progress through the adjacent marl interval and precipitate in a very early stage of diagenesis prior to compaction (Westphal 2006).

As a final result of selective dissolution, these processes generate an absence of aragonitic components in the marly layers and to the contrary, a preservation of the same transformed in calcite or just as molds within the limestone beds.

In a rhythmic succession of interbeds, it is necessary to recognize first of all the initial differences during sedimentation, secondly the early and late diagenetic changes that take place during shallow and deep burial respectively, and lastly, the final effects of differential weathering in the outcrop. Accordingly, we can make use of the concepts of a primary or initial sedimentary stratification, a diagenetic stratification, and a final or weathered stratification, overprinting the former two.

According to Ricken (1986), the sharp contacts in the outcrops between the limestone beds and the marly interbeds, are not match exactly with the differences in the proportion of original sedimentary carbonate, nor to the diagenetic carbonate; each limestone or marl bed has a maximum and a minimum of carbonate respectively, located the central portion of the same and the transitions between the two are gradual, not well defined, as one can see presently in very recent outcrops which are not greatly affected by weathering or in core wells.

The Pliensbachian marl-limestone rhythmite

The alternations of limestones and grey marls that make up the Pliensbachian marine successions of the Jurassic Basin of Asturias (Rodiles Formation) crops out in several places of the coast between Gijón and Ribadesella localities (Figs.1,2,3).

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The carbonate succession accumulated in an ancient epicontinental sea with an irregular floor containing depressions and highs, probably controlled by a reactivation of ancient Variscan faults similar to those described by Wetzel *et al* (2003) in the NW of Switzerland. These events took place within a general context of thermal subsidence between two successive extension periods focused in the Triassic and in the Late Jurassic-Early Cretaceous. During the latter phase, a series of fluvial and marine sub-basins were formed associated with the initial opening of the Bay of Biscay (Ziegler,1990). The epicontinental sea rich in carbonates, prior to the afore mentioned rift, occupied a large extent over a great portion of Eastern Europe, situated in an intermediate zone between the Boreal and Thethys domains and separating the Iberian Massif to the South and Armorican Massif to the North. The palaeontological data currently available appears to indicate a predominance of boreal faunas, although also containing diverse incursions of those coming from the Tethys (Comas-Rengifo & Goy, 2010).

The progressive increase in the thickness and the palaeobathymetry of the rhythmic marl-limestone successions within the Asturian Jurassic Basin, as well as the marl/limestone ratio from E to W along the coast, suggests the existence of an intra-platform elongated depression situated around the westernmost area of the coastal cliff outcrops (East of Gijón)

The rhythmic succession of the Vega beach

The rhythmic marl-limestone succession of the Pliensbachian that makes up the Rodiles Formation at Vega beach, is 45 m thick (Figs.6,7). Some of the limestone beds, which have a maximum thickness of 45 cm, show a lenticular geometry at outcrop scale, although the lateral extension of those lenses is quite variable, ranging between just a few centimetres (lenticular nodules) and more than ten metres. The origin of these lenticular beds is sometimes controlled by depositional processes, such as the case of limestone bodies having convex bases and flat tops, which represent scours, having bioclasts or whole faunas at their bases and locally displaying normal grading (Figs.8,9). In contrast, others appear to have an exclusively diagenetic origin without any apparent primary control. In this case the limestone beds tend to display a biconvex form.

Within the *black shale* intervals, continuous lateral beds of limestones bounded by planar surfaces and having internal parallel lamination are common, as a result of the diagenetic cementation of the *black shales* (Weedon, 1986; Arzani, 2004). The carbonate content in the limestone beds of the rhythmic series ranges between 70 and 92% and the texture varies from mudstone to packstone (Bádenas *et al.,* 2009) depending not only of the bathymetry but also of the variations in hydraulic energy.

The thickness of the marl interbeds can reach up to 1.25 m, with the larger values present in the dark grey marls and *black shales*. Many of these marly interbeds show one or more erosively based, bioclastic levels, commonly with normal grading trends and a sparse parautochtonous fauna. In other levels, bioclasts and the parautochtonous faunas appear dispersed in a very fine-grained matrix. The colour of the marls varies between medium grey (lighter at the surface due to weathering) and dark grey to black, depending on the content of organic matter and the presence of scattered pyrite of very fine grain size.

Among the invertebrate prevailing macrofauna, molluscs such as ammonites, belemnites, and bivalves, are represented, along with brachiopods, crinoids and echinoids. Within the limestone beds, the fossils appear almost always without deformation due to compaction, while in the marls, many of them are crushed with the exception of those with a thick calcitic shell. The greatest values of flattening are found within the internal moulds of ammonites preserved in episodes of *black shales*. These differences result from the cementation processes that took place in the early diagenesis and during a stage of shallow burial (Westphal, 2006), evidence of which is shown by the localized presence of infilled internal moulds of ammonites traversed by *Chondrites* (Fig.10). One exception is found in the black laminated limestones that locally reveals flattened ammonites, which means that cementation was delayed late diagenetic and formed after compaction.

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Figure 6. General view of the marl-limestone alternations in the eastern part or the Vega beach (Stop 1)
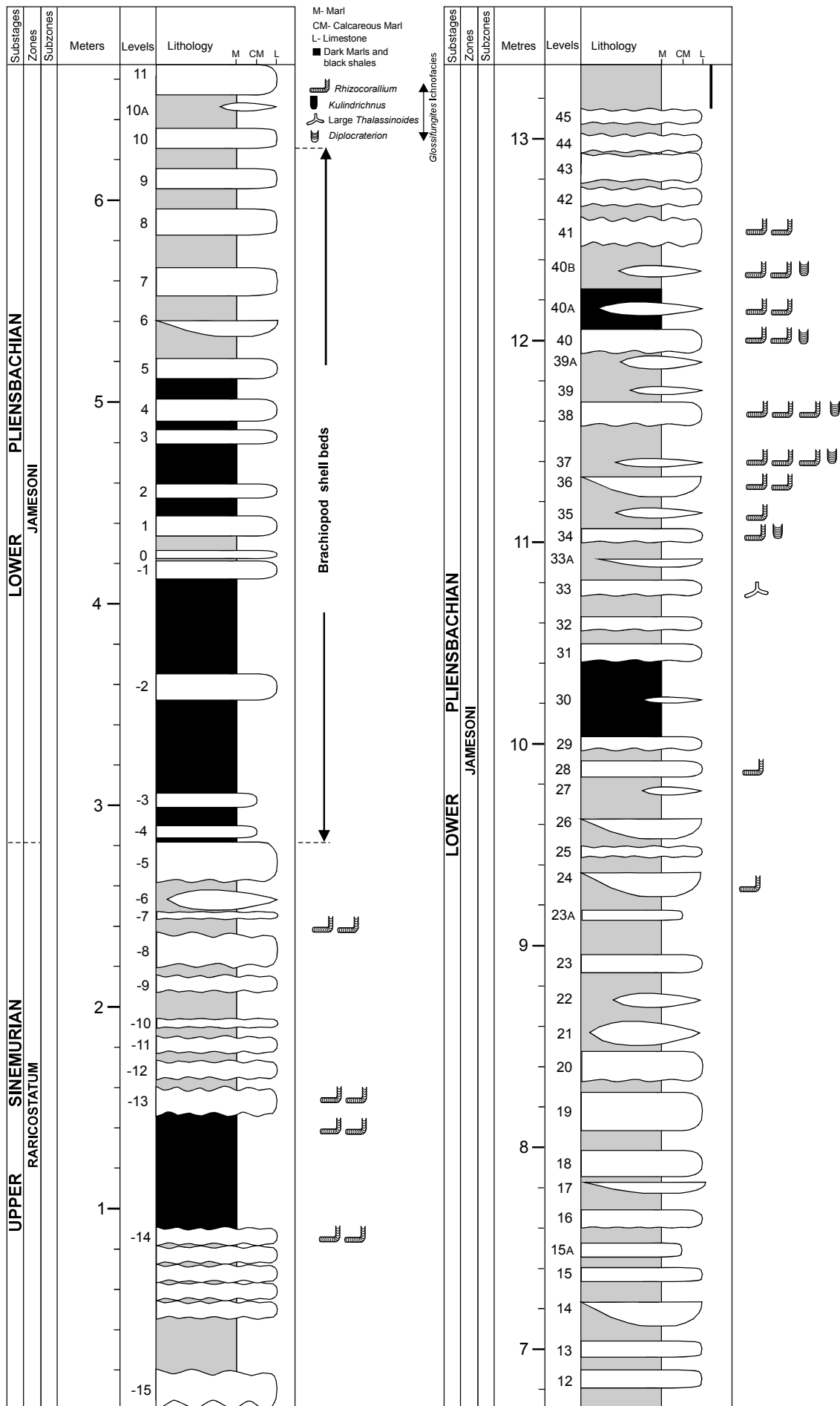
Figure 7. Stratigraphic log of the middle and upper part of the Rodiles Formation in the Vega beach (Ribadesella)

Figure 7. Cotinuation

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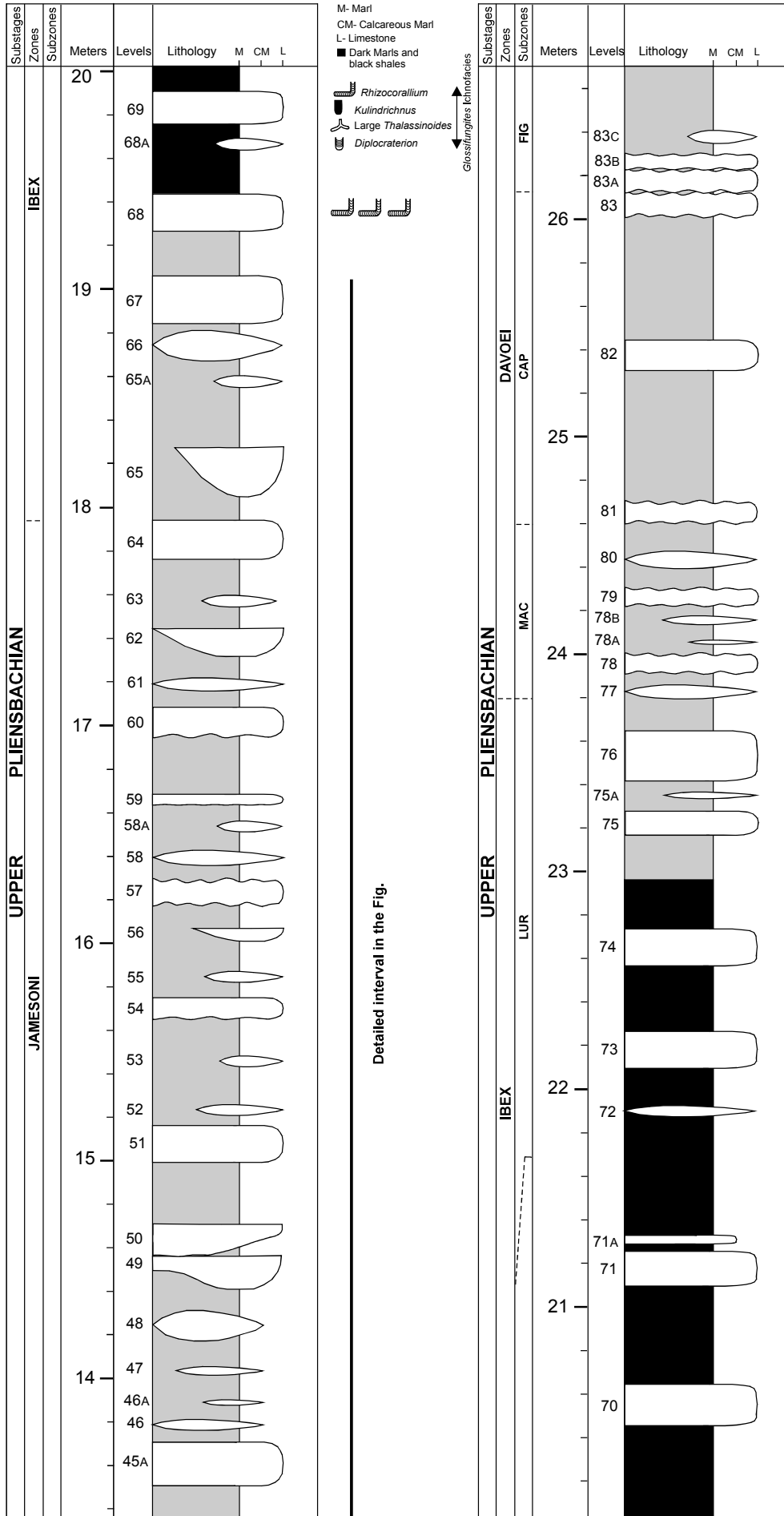
Figure 7. Continuation

Figure 7. Cotinuation

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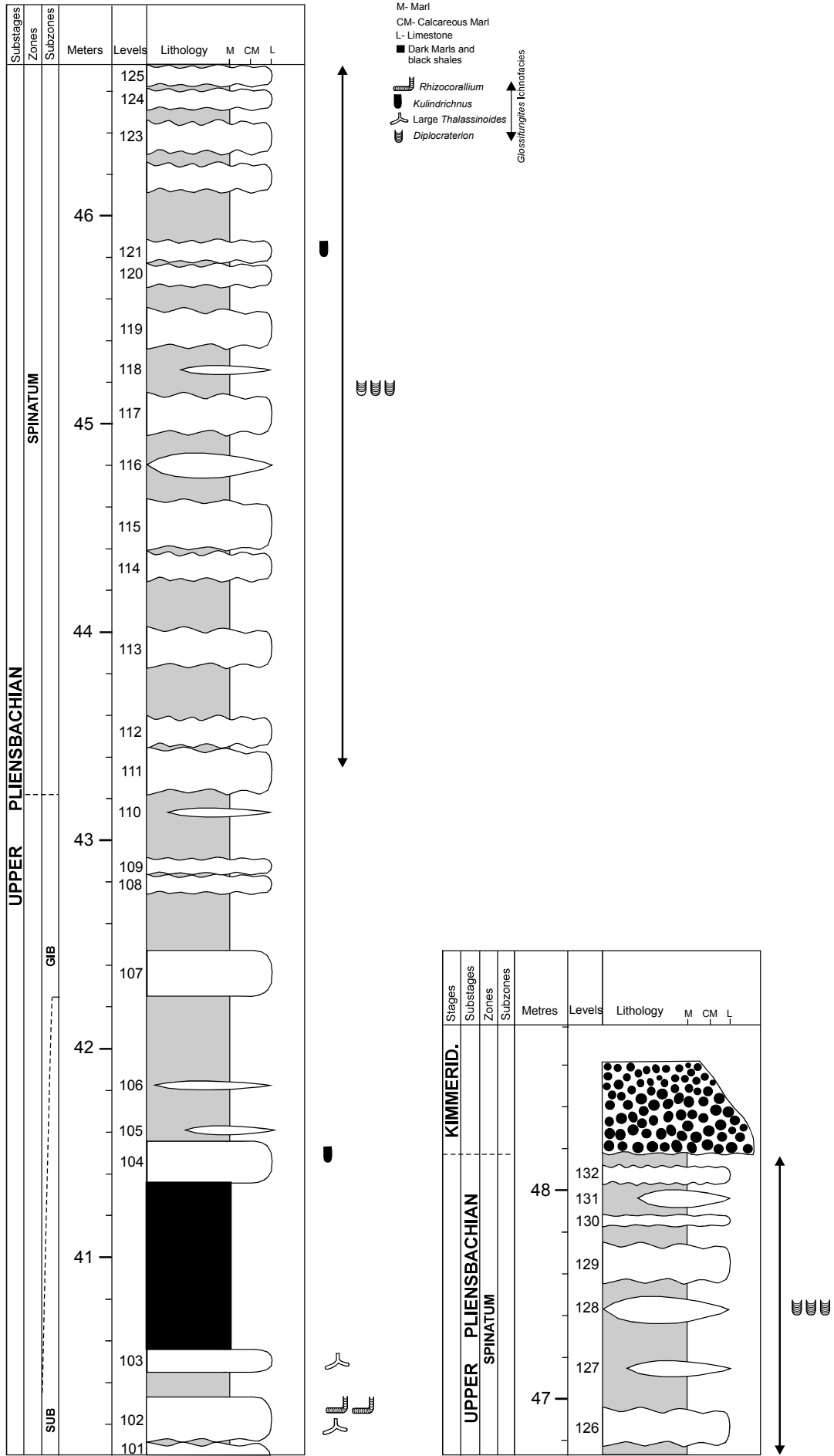
Figure 8. Shallowing episode belonging to the Jamesoni Zone, composed of five amalgamated and scoured limestone beds (levels 41 to 45 in the Fig.7). Vega beach section

Figure 9. Lenticular bed of limestone filling a scour excavated by storms. Note the lateral transition between the lenticular limestone and the light calcareous marl as well as the scours truncation of bioturbated level. Jamesoni Zone. Punta La Llastra section

Figure 10. *Chondrites* traversing interval moulds of body fossils. The process evidence that shell dissolution was produced during the early diagenesis in a phase of very shallow burial.
(A) Phosphatized internal mould of an abraded fragment of ammonites. Vega beach section.
(B) Internal mould of an ammonites in a marl interbed. Punta La Llastra section.
(C) Internal mould of a bivalve in a limestone bed. Jamesoni Zone. Vega beach section

Many examples of interval moulds of burrowing infaunal bivalves in life position occur in the limestone beds and the marl interbeds, resulting from the early diagenetic dissolution of their aragonitic shells (Westphal *et al.*, 2004; Arzani, 2006; Cherns *et al.*, 2008; Wheeley *et al.*, 2008). Some of them show escape structures to the sedimentation and another lateral displacement structures (*Protovirgularia*, Fig.11)

Among the vertebrate fauna, only a few skeletal remains have been found, still pending of identification, in the base of a limestone layer belonging to the Jamesoni Zone. Other marine successions from the Pliensbachian of Asturias have yielded diverse skeletal material including several plesiosaur (Bardet *et al.*, 2008) and ichthyosaur bones and theeth.

Figure 11. Horizontal locomotion structures due to infaunal bivalves (*Protovirgularia*) on the top surface of a limestone bed. Jamesoni Zone. Vega beach section

Cyclical organization

The major cycles in the rhythmic series at Vega beach are measured in metres, with an asymmetrical character (i.e.: small-scale R cycles after Zecchin, 2007), and can be easily identified in coeval outcrops of the region.

Usually these cycles begin with laterally extensive alternations of laminated limestone beds and dark grey to black marls. Progressing upwards, alternations of the same lithologies are seen, but with the limestone beds predominating and becoming richer in bioclasts and whole or disarticulated fauna and showing an irregular stratification (nodular, lenticular or wavy bedding) as well as frequently erosive amalgamations. The dark grey colouration of the marls also disappears towards the top within each shallowing cycle, which represents also a progressive decrease in the sedimentation rate.

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One of the most evident shallowing cycles in this succession occurs in its upper part just below the contact with the terrigenous series of the Late Jurassic (Vega Formation), starting in the Margaritatus Zone with *black shales* and extending into the Spinatum Zone (Fig.12) The most characteristic traits of the upper portion of this cycle are the abundance of bioclasts and articulated or disarticulated macrofauna, mostly bivalves, belemnites, brachiopods and ammonites. Particularly abundant are the large pectinids and the belemnite concentrations, which reach a maximum for the whole Jurassic succession. A high density of U-shaped biogenic structures, mostly *Arenicolites* and protrusive *Diplocraterion* (equilibrichnia), are present here which correspond to a high energy environment affected by common omission surfaces due to erosion. The limestone beds often have a packstone texture, are laterally irregular, have wavy boundaries, and contain scours and other erosive features.

Figure 12. Shallowing interval (Spinatum Zone) characterized by the dominance of the *Diplocraterion-Arenicolites* ichnofabric. Note the uncorformable contact with the overlying fluvial strata of the Vega Fm.Vega beach section

This shallowing episode, characterized by a typical increase in the degree of oxygenation of the sea floor, is recognizable throughout W and NW Europe just as in the occidental realm of the Tethys. In some of those locations, such as the Northern Sea Basin or the Yorkshire region in England, the regressive succession includes siliciclastic deposits with structures indicative of subaerial exposure. The process should be related to a doming process previous to the opening of the central sector of the North Atlantic (Hallam, 1984; Steiner *et al.*, 1998).

The geochemical data based on stable isotopes appear to also indicate a strong cooling during the Spinatum Zone just as in relatively more saline waters, favouring the migration of boreal faunas towards the South (Bailey *et al.*, 2003; Morard *et al.*, 2003; Rosales *et al.*, 2004a,b, 2006; van de Schootbrugge *et al.*, 2005; Tremolada *et al.*, 2005; Arias, 2007, 2009; Gomez *et al.*, 2008; Dera *et al.*, 2009 a,b; Mailliot *et al.*, 2009; Reggiani *et al.*, 2010).

For some authors, this stage of cooling may be of glacio-eustatic origin (Guex *et al.*, 2001; Morard *et al.,* 2003). In particular, Price (1999) demonstrated the presence of dropstones and glendonites at high palaeolatitudes.

Other minor shallowing cycles can be seen in Figs. 7,8 between the tops of levels 40 and 45 within the Jamesoni Zone, between the tops of levels 68 and 76 within the Ibex Zone, or between the tops of levels 76 and 81 up near the end of the Ibex Zone and in the lower part of the Davoei Zone. Two new shallowing episodes are seen in turn between the layers 83 and 83 B near the boundary between the Capricornus and Figulinum subzones, within the Davoei Zone, and at the interval between the levels 85 and 86 A in the lower part of the Stokesi Subzone, within the Margaritatus Zone.

Millennial-scale elementary cycles

The excellent coastal outcrops from the Rodiles Formation allowed for the recent well-detailed study and bed-to-bed correlation (Bádenas *et al.*, 2009, 2010) of rhythmites in the Vega beach (Ribadesella) and the Punta La Llastra (Villaviciosa) series which are 18 km apart (Fig.13). These were formed in a storm-influenced ramp below the fairweather wave base.

A series of elementary cycles or rhythms can be identified in each of these two equivalent sections. These rhythms are generally a thickness of decimetres (Fig.13) and consist of alternations of limestones or light grey calcareous marls, and medium to dark grey laminated marls (*black shales*).

Light coloured marls or limestones

Comprise erosively based (with local scours), graded bioclastic layers, interbedded with intensely bioturbated hemipelagic layers. In addition to the allochtonous bioclasts, these deposits commonly contain autochtonous and parautochtonous articulated and disarticulated macrofauna. A small proportion of sand a silt-sized quartz grains is often present in these layers. The grade of bioturbation within the bioclastic layers is highly variable.

Dark grey marls limestones

The dark grey marls, *black shales* and rare limestones have a characteristic lamination and the proportion of bioclasts and articulated or disarticulated macrofauna is much lower and less diverse than that found in the lighter coloured layers mentioned in the previous paragraph. The darker intervals, less rich in carbonate, locally show thin erosively based lenses rich in crinoids, especially those coming from the Margaritatus Zone. Furthermore, the faunal diversity is specially low in these black shale intervals.

The ichnofauna consists of *Chondrites, Planolites* and *Phymatoderma,* penetrating from the overlying lighter coloured marls and limestones.

These light and dark elementary cycles are often included inside of limestone-marl alternations affected by diagenetic cementation (Novo, 1995). Moreover, these do not coincide exactly with those of the cementation and diagenetic dissolution zones, nor with the weathered limestone-marl boundaries in the exposed outcrop (Biernacka *et al*., 2005).

Parallel to a change from relatively shallower zones of the ramp to the more subsident depocenters, there is a gradual change in lithology, texture, organic content and colour of the two intervals that make up each of these high frequency elementary cycles.

The relatively high initial levels of carbonate both in the bioclastic and hemipelagic intervals are the locations most favourable for the formation of limestone beds during diagenesis. The percentage of Corg in these beds is very low, varying between 0.09 and 0.87 in the Jamesoni and Ibex zones (Saavedra Fernández-Combarro, 1999). However, within some of the laminated and relatively thick *black shale* intervals, such as those that characterize one part of the Ibex and Margaritatus zones, there are some dark grey and blackish
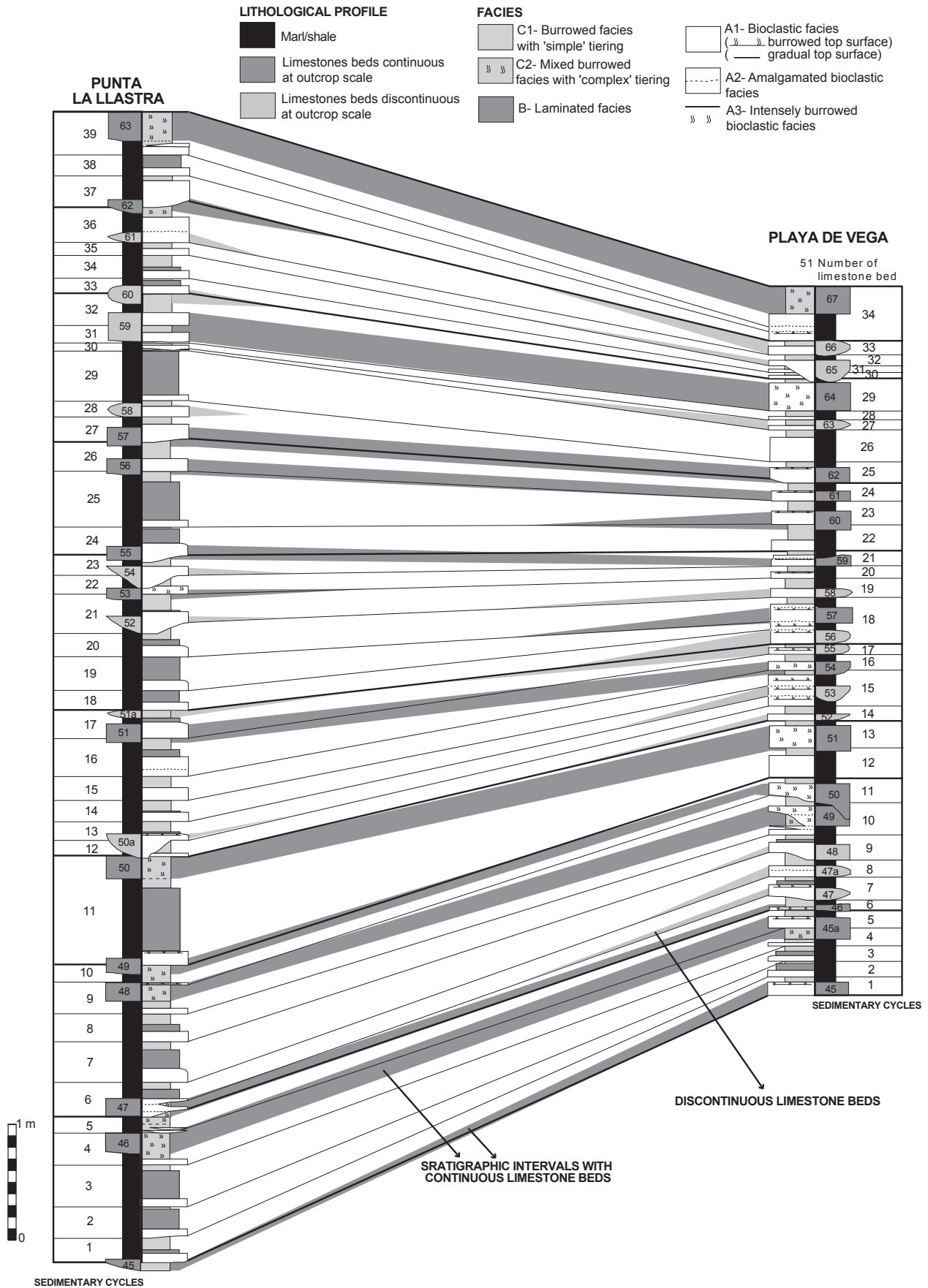
Figure 13. Lateral correlation of small-scale elementary and diagenetic cycles between the Punta La Llastra (Villaviciosa) and Vega beach (Ribadesella) sections (Bádenas *et al*., 2010). Distance between the two outcrops: 18km

parallel-laminated limestones, which have been generated by diagenetic cementation of similarly coloured marls. Examples of dark laminated limestones of a similar character have been described in Lower Jurassic rhythmic series of Great Britain (Arzani, 2006; Sheppard *et al.,* 2006; Paul *et al.,* 2008).

Many limestone beds wedge laterally, at least at outcrop scale, therefore the eventual measures of number and thickness of marl/limestone rhythms show many lateral variations depending of the place selected for these measures (Figs.7,8,14,15). Some of these limestone lenses with convex bases and flat tops, represent scours infilled by bioclastic material (Figs.8,9).

Studies aimed to determine possible orbital oscillations related to periodic climatic changes from the rhythmic carbonate series based exclusively on thickness measurements of limestone and marl alternations may be subject to considerable errors if one overlook early diagenetic process (Böhm *et al.* , 2003; Westphal *et al,.* 2004; 2008, 2010) overprinted to the primary bedding.

Moreover the presence of numerous and recurrent indicators of repeated minor sedimentary interruptions (e.g. erosionally based storm beds truncating underlying tiered ichnofossil communities suggests that it is necessary to exercise caution when using the carbonate rhythms to obtain chronostratigraphic data, such as already indicated by Westphal *et al.* (2004, 2008). This is especially true in the case of rhythmic successions in outcrops that aren't exceptionally well preserved.

Figure 14. Limestone-marl alternations including higher frequency elementary cycles. Upper part of the Jamesoni Zone. Vega beach section

Figure 15. Rhythmic marl-limestone succession and elementary cycles. Note the lateral transit between limestones and light marls. Jamesoni Zone. Vega beach section

Recent data based on ammonites (Comas-Rengifo & Goy, 2010) are taken from the section studied at Vega beach within the upper third of the Jamesoni Zone and the basal portion of the Ibex Zone. This has allowed for good lateral correlation to be established between elementary cycles in both successions. Because the most distal section (Punta La Llastra) shows less erosive amalgamations and therefore the sedimentary record is more continuous, it has been chosen to determine the number of elementary cycles (39 in total), as well as the limestone-marl couplets (21).

Assuming that the Jamesoni Zone lasted for 1.15 my according to the Geologic Time Scale (Ogg *et al.*, 2008), it seems that the period of recurrence for each elementary cycle is approximately 9829 years. Taking into account that some of these still show evidence of erosive amalgamations of two or more successive bioclastic elements and that the sedimentary record shows multiple signs of storm erosion and episodes of bypass, their periodicity can be considered much lower.

Millennial-scale cycles similar to these, also constituted by rich and poor carbonate beds, have been described by Elrick *et al.* (1991) in ramp slope deposits of the Mississippian of Montana, as well as by Elrick & Hinnov (2007) in various deep-water rhythmic series of the Palaeozoic of North America.

Given the lateral wedging of some of the limestone beds, following the chosen section they tend to intersect in the rhythmic section of the Punta La Llastra outcrop (Fig.16), between 15 and 18 limestone/ marl cycles of the 21 that can be recognized visually if the section is examined laterally to a distance of approximately 150 m.

If we achieve the same calculation as before, now considering the thickness of the preserved limestone-marl cycles, instead of the elementary cycles, the periodicity obtained for the couplets would be between 18253 and 21277 years, depending on the chosen location. This is consistent with Milankovitch orbital precession values, although a causal connection cannot be firmly established.

Figure 16. General aspect of the marl-limestone alternations. Jamesoni Zone and lower part of the Ibex Zone. Punta La Llastra section (Villaviciosa). Note also the primary vertical repetition of light (carbonate-rich) and dark (carbonate-poor) marls representing millennial-scale (climatic?) cycles.

The most likely origin of the lighter bioclastic levels appears to be related to strong and frequent storm events which spread the mostly fragmented biological material supplied from littoral bars all the way to the outer part of the ramp, together with parautochtonous articulated and disarticulated shells (Figs.9,17,18,19,20).

The more calcareous levels, often represented by a light grey, intensely bioturbated mudstone to wackestone limestone, maybe represent stages of low sedimentation, mostly hemipelagic, the sedimentary record of which is situated immediately above and/or in lateral areas to bioclastic scours of storm origin associated with abundant episodes of by pass. The storms that generated these erosive structures supplied oxygen to the sea floor and at the same time transport an immigrant fauna dominated by decapod crustaceans from shallower-water ramp areas. The ichnofauna includes mainly *Asterosoma, Thalassinoides, Gyrolithes, and Rhizocorallium*. Decapods carried by gravitational flows to relatively deeper-water zones, usually less oxygenated, were commonly known as doomed pioneers, according to the interpretations of previous authors, such as Föllmi & Grimm (1990) and Grimm and Föllmi (1994).

To the contrary, the levels of dark grey to black laminated marls preferently accumulated in deeper areas of the ramp can represent calm stages of the sea, sporadically affected by minor and less frequency storm episodes that put a large quantity of mud in suspension together with dispersed bioclastic material. A similar hypothesis was proposed by Van Buchem *et al.* (1994) for the rhythmic series of the Pliensbachian of Yorkshire (UK), by Mattioli & Pittet (2004) for rhythmic series of the Lower Jurassic in central Italy and by Biernacka *et al.* (2005) for the Late Permian of the Nord-Sudetic Basin in Poland. During these episodes, the sea floor was likely relatively poor in oxygen, especially towards the deeper areas of the ramp especially favourable for the *black shale* accumulation. In the upper portion of these intervals, there is habitually a well developed vertical zonation of ichnofauna (simple tiering, Taylor *et al.,* 2003).

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Figure 17. Graded marly bioclastic level (tempestite) with planar erosive base and parauthoctonous brachiopds in the top. Above laminated dark marls. Punta La Llastra section

Figure 18. Graded marly bioclastic level similar to preceding image. Note in this case a highly bioturbated level in the upper part of the marl interbed. Jamesoni Zone.Vega beach section

Figure 19. Scoured limestone bed (49) with fossil concentrations in the base of scour. Jamesoni Zone. Vega beach section

Figure 20. Bioclastic storm bed (lighter colour) showing normal grading between darker marls. Note the bioclastic infill in flattened burrows of *Thalassinoides* and the underlying bioturbation by *Chondrites*. Jamesoni Zone. Punta La Llastra section

Thickness of these light colour bioclastic-hemipelagic levels is in direct relation to the persistence of relatively oxygenated conditions of the bottom (Savrda & Bottjer, 1989; Savrda, 2007), depending at the same time on the frequency and/or intensity of the storms that affect the carbonate ramp. Complex tiering patterns (sensu Taylor *et al.*, 2003) occurred under slow sedimentation (Figs.21,22,23,24,25).

Figure 21. Complex tiering ichnofabric dominated by *Chondrites*, *Rhizocorallium* and *Thalassinoides* in a limestone bed. Vega beach section

Figure 22. Limestone bed showing a gradual transit between the Light *Chondrites-Phymatoderma* ichnofabric (lower part) to *Asterosoma-Thalassinoides* ichnofabric (middle part) and to Diffuse mottling ichnofabric (upper part)

Figure 23. Light *Chondrites-Phymatoderma* ichnofabric underlying an *Asterosoma-Thalassionides* ichnofabric (lighter bioclastic marl)
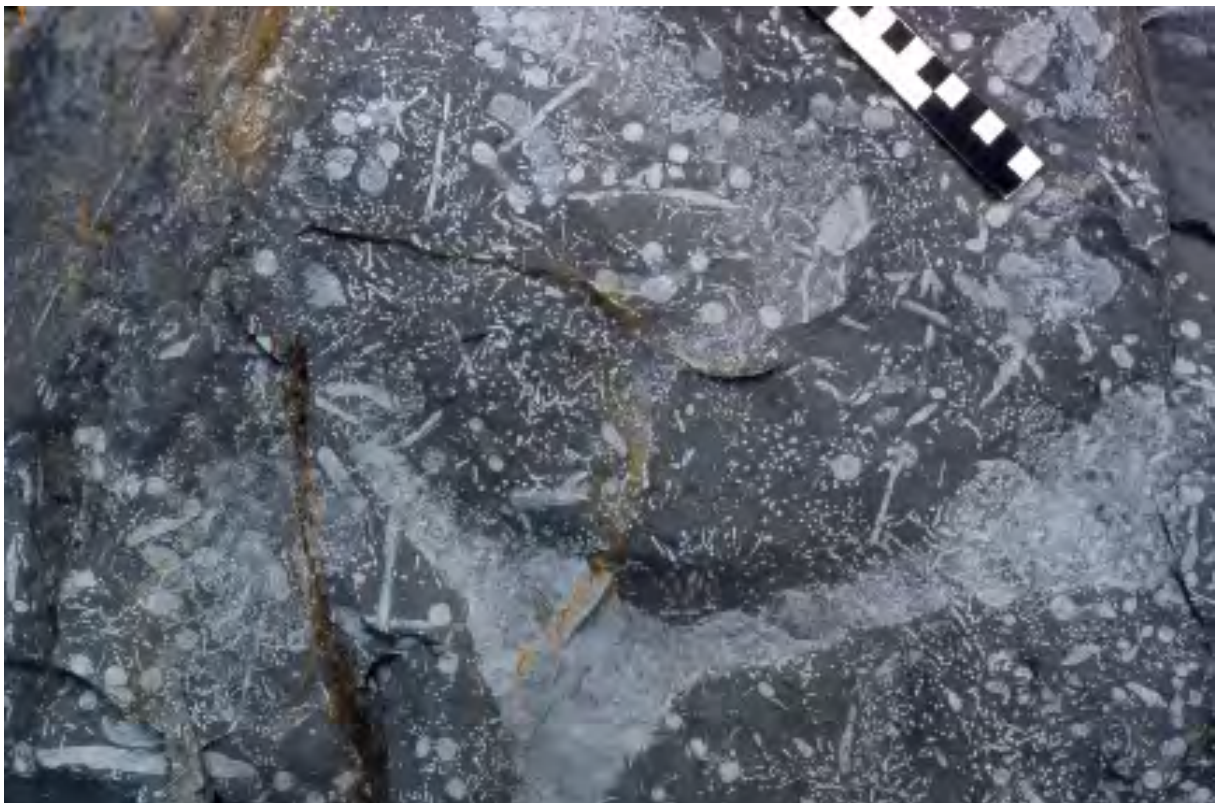

Figure 24. Bedding plane view of a bioturbated marly interbed with *Thalassionides, Chondrites, Rhizocorallium* and *Planolites.* Complex tiering ichnofabric. Punta La Llastra section
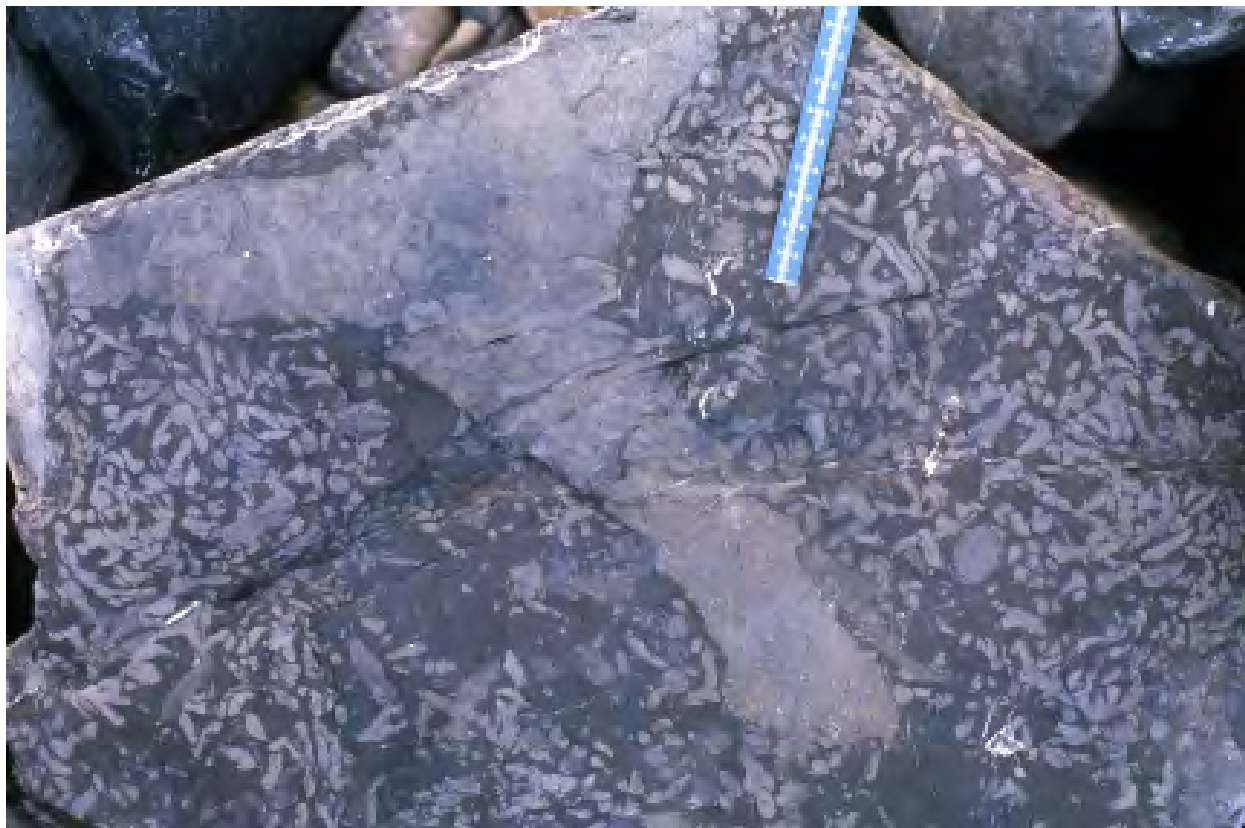
31

Figure 25. Bedding plane view of a marly interbed bioturbated by *Thalassionides* and *Phymatoderma.* Complex tiering ichnofabric. Punta La Llastra section

Deeply penetrating (up to 15 cm), up to 8 cm wide *Thalassinoides* (*Spongeliomorpha*?) occurs locally within the *black shales* of the Margaritatus Zone in the Vega beach section, cross-cutting the background ichnocoenosis (Figs.26,27,28). This illustrates the firmground *Glossifungites* ichnofacies.

It still needs to be confirmed if these alternating episodes of high and low energy, of different colour and lithological composition are related with climatic cycles of millennial scale which may in turn represent a part of other larger-scale precession cycles inside the Milankovitch band.

Preliminary analysis suggests that the intensely bioturbated limestone most likely represents more arid climate, whereas the dark and laminated marl and *black shale* intervals correspond to more humid periods. A similar interpretation has been proposed previously for rhythmic marl-limestone series of different ages (Pratt, 1984; Bottjer *et al*., 1986; Sethi & Leithold, 1994; Pittet *et al*., 2000; Mader *et al*., 2004; Sagasti, 2005; Meyers, 2006; Mailliot *et al*., 2009; Mutterlose *et al*., 2009; Tucker *et al*., 2009; Bonis *et al*., 2010). In any case, the general climatic conditions may have been controlled by a monsoonal system, which is to be expected in a relatively low latitude zone (roughly 27-30º N) and nearly to the coast, within a Pangea that was starting to break apart at that time (Wang, 2009).

The recurrent intervals of thin laminated *black shales* probably represent episodes of warm and humid climate with more frequent precipitation and great content of silt, clay and plant material. As a result, a brackish surface layer may have been formed, leading to density stratification and oxygen depletion on the sea floor. Extensive colonization of the sea bottom may have occurred only after storm events that may have disturbed this water stratification. A large productivity of phytoplankton should be added to this, related with the sources of nutrients coming in from the emerged zones. Moreover, during these stages with higher sedimentation rates the muddy sediments of the seafloor were probably soupy (Schmid-Röhl & Röhl, 2003).

Figure 26. Large *Thalassinoides* ichnofabric (*Glossifungites* ichnofacies) bed 98. Margaritatus Zone. Vega beach section

Figure 27. Large *Thalassinoides* ichnofabric (*Glossifungites* ichnofacies) bed 99. Margaritatus Zone. Vega beach section
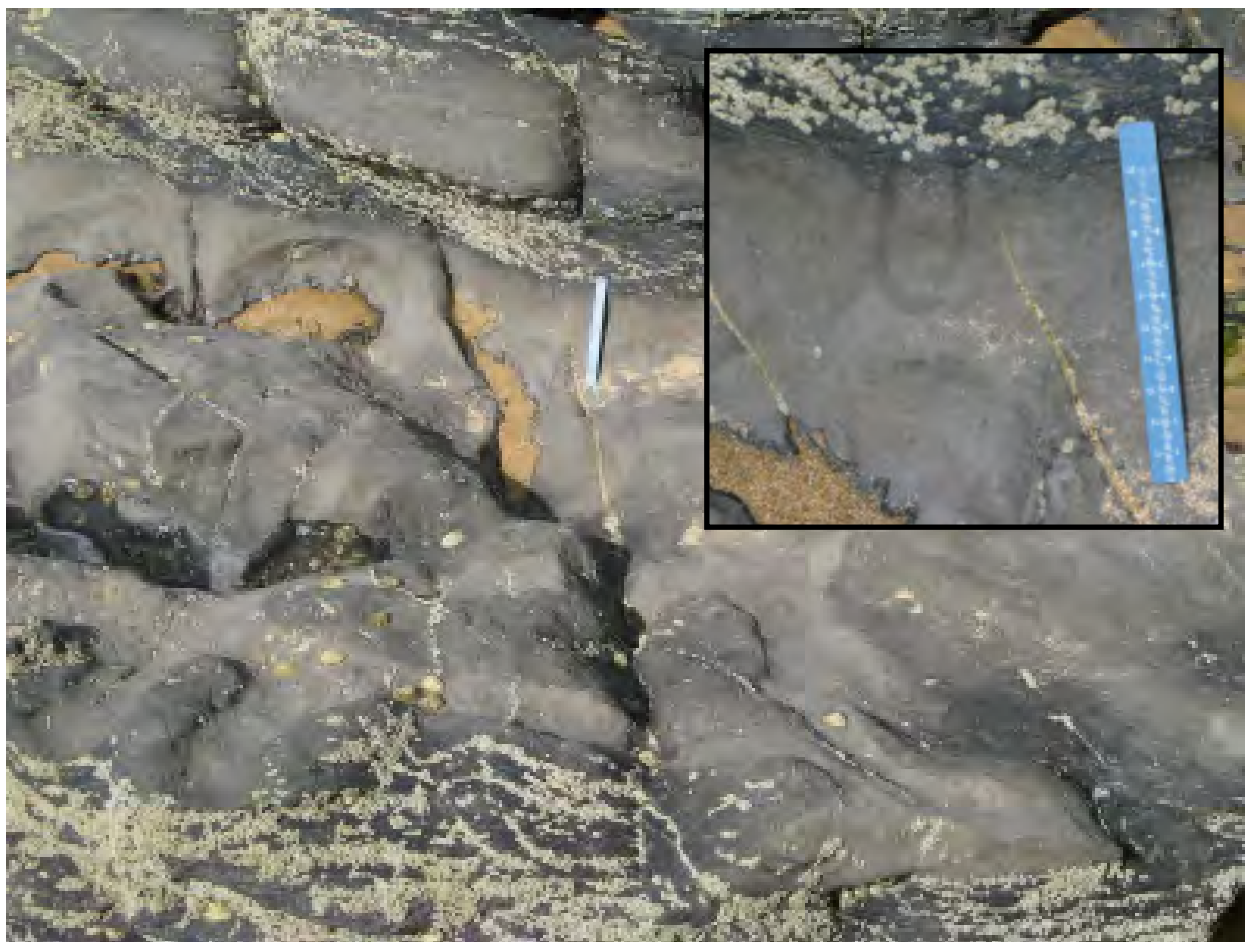
Figure 28. Large *Thalassinoides* ichnofabric in a Light calcareous marl. Note a dark marl infilled *Arenicolites* burrow truncated by a erosion surface (*Glossifungites* ichnofacies). Margaritaus Zone. Vega beach section

Furthermore, the limited tidal range in this extensive epicontinental sea may have also promoted density stratification (Allison & Wright, 2005; Wells *et al*., 2008). Additionally in these circumstances, the water cloudiness will be greater, which hinders the carbonate productivity (Tucker *et al.,* 2009). This implies that, during these mud-rich episodes from continental provenance, the development of bioclastic bars in the most shallow part of the ramp, would likely be very limited or even completely halted (Hallam, 1984). Isotopic studies of C and O from belemnite rostra show a temperature increase associated with these episodes of *black shale* accumulation (Armendáriz *et al.,* 2010).

On the contrary, the main bioclastic levels and the light grey and intensely bioturbated hemipelagics, may have formed in a well oxygenated environment maybe during stages of aridity with somewhat lower temperatures. For the most part, the dilution with fine terrigenous continental material in these intervals is minimal, which was favourable to a greater production of carbonates, such that a greater development of the benthic fauna and deposition of bioclastic bars took place in the shallower zones of the ramp. The partial destruction of those bars during storms would supply the necessary material for the formation of the bioclastic levels of these elementary cycles.

In the particular case of the Early Jurassic of Asturias, the influence of upwelling processes as a possible mechanism for the deposition of the *black shales* is unlikely. This is indicated by the large distance of deep-water areas, as well as the virtual absence of limestone-marl rhythmites in determined areas of active margins affected by those processes during the Jurassic. Waters originated at great depths are also unfavourable to the production and accumulation of aragonite, which is an essential mineral for the formation of carbonate rhythmites (Westphal & Munnecke, 2003). Röhl & Schmid-Röhl (2005) also reject a model of upwelling for the origin of the Posidonia Shale of the Lower Toarcian in SW Germany. The early dissolution of

fossil aragonitic shells constitutes a local source of carbonate which favoured the processes of cementation, leading to the development of limestone beds (Munnecke & Samtleben, 1996; Biernacka *et al.*, 2005; Wheeley *et al.*, 2008).

In the shallower areas or submarine highs of the epicontinental sea in which these series were forming, the precipitation of calcium carbonate which would later give way to the limestone beds, can be partially or completely controlled by the variations of the initial composition of sediment and by sedimentary structures (e.g. scours, ripples, ichnofossils, etc.), and is also early to diagenetic, as was discussed earlier. Nonetheless, in the deeper and/or more subsiding areas, having a greater amount of organic matter, the importation and precipitation of calcium carbonate is somewhat lowered resulting in the formation of marly limestones or calcareous marls. Additionally, this precipitation took place in a later stage of diagenesis which is demonstrated, for example, by the internal moulds of ammonites or by ichnofossils (e.g. *Thalassinoides* o *Chondrites)* which show a certain degree of compaction although always less than that in the adjacent laminated marls (Figs.20,23,29,30,31). A similar phenomenon was mentioned by Scotchman (1989) in the Kimmeridge Clay of the UK.

The absence of *Zoophycos* in these rhythmic series of the Pliensbachian can be explained by several interrelated factors, including the relatively shallow water depths, frequently high water cloudiness, and limited connection with the open sea.

The rhythmic succession of the Pliensbachian of Asturias is therefore influenced by redox, dilution, productivity and energetic cycles.

Figure 29. Light *Chondrites-Phymatoderma* ichnofabric showing a gradual transit to the *Asterosoma-Thalassinoides* ichnofabric. Note the laminated dark marls in the lower part. Punta La Llastra section

Figure 30. Bioclastic storm deposit showing normal grading and an upwards transition (Dark *Chondrites-Phymatoderma* ichnofabric) to laminated dark shales. Note the erosive base of the tempestite bed cutting a bioturbated level dominated by the deep tier *Phymatoderma*. Jamesoni Zone. El Puntal section

Figure 31. *Phymatoderma* and *Chondrites* specimens in the boundary between an highly bioturbated limestone and a laminated to bioturbated marly interbed. Note the differential compaction of the ichnofossils in both lithologies, demostrating the early diagenetic cementation of the limestones. Jamesoni Zone. El Puntal section

According to these approaches, the light grey bioclastic and/or bioturbated beds (limestones or marls) represent stages with well-oxygenated sea bottoms affected, alternative or simultaneously, by low accumulation rates (hemipelagites) and high accumulation rates (tempestites). Nevertheless the whole take place during intervals of relatively lower sedimentation rate according to nomenclature of Gómez & Fernández-López (1994).

To the contrary, the medium to dark grey, laminated and poorly bioturbated beds represent stages or environments with poorly-oxygenated to anoxic sea bottoms characterized by relatively higher sedimentation rate and more elevated proportions of organic matter.

The pronounced differences in the Bioturbation Index (BI) between the lighter and darker beds are therefore related with: a) the oxygen content in the sea bottom controlled by the intensity and frequency of storms, and b) the relative sedimentation rates.

To the preceding processes it is necessary to consider the local and regional tectonics controlling the temporary apparition of relatively shallow water subsident depocentres favourable for the thickest *black shale* episodes.

Major episodes of accumulation of dark marls and *black shales*

They are represented by larger scale laminated dark marl and *black shale* intervals dm and m in thickness and therefore more prolonged in the time that the others to bed-scale previously analized.

These intervals include thin lighter grey cm-thick bioturbated marly layers which likely represent sporadic events of greater oxygenation in the sea floor. The main *black shale* episodes in the Vega beach succession are located, according to the biostratigraphic data of Comas-Rengifo & Goy (2010), in the Upper Sinemurian (part of the Raricostratum Zone), in the Lower Pliensbachian (part of the Jamesoni Zone and almost the whole of the Ibex Zone, Fig.32) and in the Upper Pliensbachian (Subnodosus Subzone within the Margaritatus Zone Fig.33). These three main episodes of *black shales* are practically synchronous with those mentioned by Quesada *et al.* (2005) in the Basque Cantabrian Basin, althogh also show a partial equivalence with those of the Upper Sinemurian and Pliensbachian of the Lusitanian Basin in Portugal (Duarte *et al.,*2010).

These major *black shale* deposits are coincident roughly with the onset of transgressive episodes and peaked in the subsequent maximum flooding zones. Soft and soupy substrates and high sedimentation rates are envisaged, particularly in the more subsiding depocenters, although in seafloors relatively shallows. Van Buchem *et al.* (1994) and Fernández-López *et al.* (2000) observed the same phenomena in the upper Lias of Yorkshire (UK) and Lower Pliensbachian of Portugal, respectively. On the other hand, the presence of euxinic conditions within the photic zone of the water column has been previously suggested by Pancost *et al.* (2004), Van de Schootbrugge *et al.* (2005) and Prauss (2007).

The fact that these Pliensbachian *black shale* facies are not found in other areas of the Iberian Peninsula (other than the two previously mentioned), such as the Iberian Range, the Algarve or the Betic Zone suggests that their development is conditioned in great part by a tecto-sedimentary context within a local or a more restricted regional scope. The poorly oxygenated conditions would be a result of the irregularities of the epicontinental sea floor and its relatively limited depth, which would prevent water circulation, particularly in the most subsiding areas of the basin (Tyson, 1996).

An additional circumstance that would contribute to this frequent water-mass stratification is its palaeogeographic location typical of low gradient epicontinental shallow seas between emerged areas and isolated from the oceanic tidal waves and with microtidal to a-tidal characteristics (Allison & Wright, 2005).

Specifically for this palaeogeographic context, the phenomenon of upwelling as a generator of anoxia doesn't appear plausible in this case, given that good water circulation is required for its development.

Figure 32. Laminated *black shales* belonging to the Ibex Zone. Vega beach section

Figure 33. Repeated *black shale* episodes alternating with light grey marls and limestones. Note the smaller-scale elementary cycles. Vega beach section

Therefore, the main source of nutrients that facilitated the anoxia would be controlled by sources of water and fine sediment coming from nearby emerged areas. In any case, the amount of poorly oxygenated marine water would have been scant and in close proximity to the sea floor, in such a manner that during significant storms, the sediment-water interface would be well aerated, allowing for the sporadic colonization of a particular type of opportunistic fauna capable of bioturbating the sea floor.

The total concentration in organic carbon (TOC) in the *black shales* from the Jamesoni and Ibex zones (apdated from biostratigraphic data based on Comas-Rengifo & Goy, 2010), reach up to 7.48 % (Suárez Ruiz, 1988; Borrego *et al,*. 1997).

Apparently, not all of the marly levels rich in organic matter which appear in the succession have the same significance. For example, a given interval within the Jamesoni Zone is made up of many brachiopod-shells, many of which are infilled with oil, something that does not occur in the shells of the Ibex and Margaritatus Zones. Their origin is probably related to storm currents, rich in mud and faecal pellets, and capable of picking up a large number of these individuals, which were then buried alive. The interior of these shells was filled much later with oil, after of the deep burial stage.

One of the most typical characteristics of the *black shales* from the Ibex and Margaritatus Zones is their bioclastic concentrations, frequently lenticular, formed almost exclusively by crinoids and with an erosive base.

With respect to the range or degree of evolution of the organic material in these *black shales*, according to Suárez Ruiz (1988) and Suárez-Ruiz & Prado (1995), gradually increases from W to E along the coastal outcrops, progressing from a stage of diagenesis (close to Avilés and Gijón) to another of catagenesis (from Villaviciosa to Ribadesella), the last of which containing highly evolved organic matter and within which the outcrop at Vega beach is located. This increase pattern in the evolution to the east may have resulted from an increase of sediment load towards the East as a consequence of the rifting process that took place in the Late Jurassic-Early Cretaceous, with a direction of extension NE-SW (Lepvrier & Martínez García, 1990).

Trace fossils and ichnofabrics

Pliensbachian marl/limestone rhythmites of Asturias show a relatively high diversity of trace fossils, most of them belonging to the *Skolithos*, *Cruziana* and *Glossifungites* ichnofacies.

The most representative ichnogenera in the succession, although represented in variable proportions are: *Arenicolites*, *Asterosoma*, *Chondrites*, *Conichnus*, *Diplocraterion*, *Gyrolithes*, *Kulindrichnus*, *Lockeia*, *Palaeophycus*, *Phymatoderma*, *Phycosiphon*?, *Planolites*, *Protovirgularia*, *Rhizocorallium*, *Skolithos*, *Taenidium*, *Teichicnus*, *Thalassinoides* and *Trichichnus*.

Ichnological analysis, focusing on trace fossil composition and ichnofabrics, was conducted in a total of eleven Pliensbachian outcrops located in the sea cliffs between Gijón and Ribadesella localities.

Eight ichnofabrics have been differentiated, based on the most abundant and/or visible ichnogenera in the outcrops or the most characteristic texture (Fig.34). However, ichnofabric differentiation is variable depending on lithofacies, being easier in marls and marly limestones intervals than in limestone beds. Strong diagenetic obliteration affecting the last ones difficults a clear identification directly in the outcrop; in these cases ichnofabrics have been studied on cutting surfaces and thin sections.

A. *Light Chondrites-Phymatoderma ichnofabric (Figs.23,29,30,31,35,36,37,38)*

Characterized by the presence of light carbonate-rich filled burrows on a darker and often laminated sediment background. This ichnofabric is mainly recognized into the marly intervals.

Ichnofabrics

E
C
B
A

Ichnofabrics

C
D
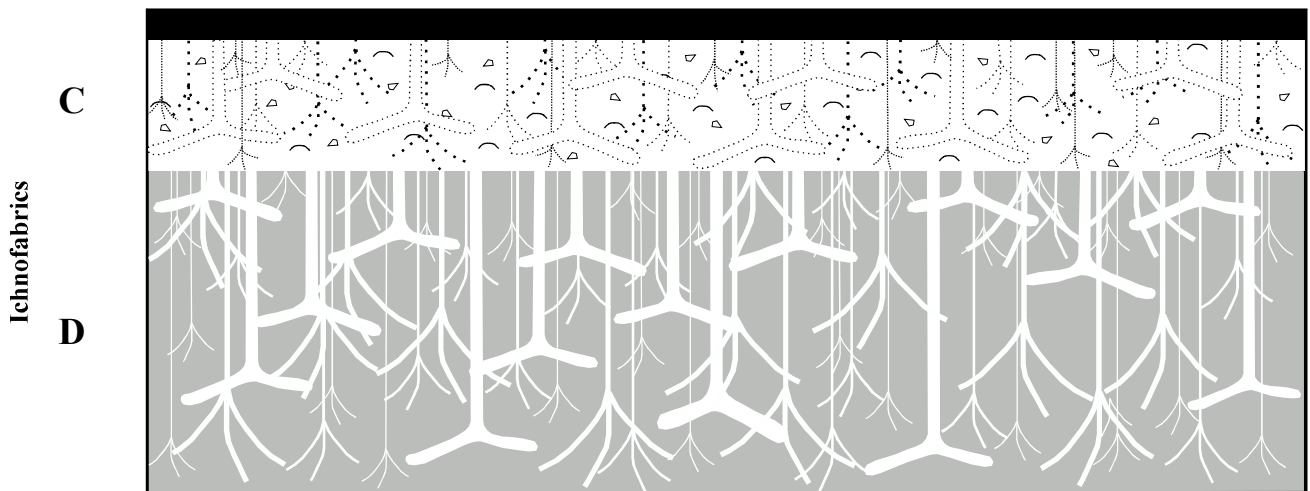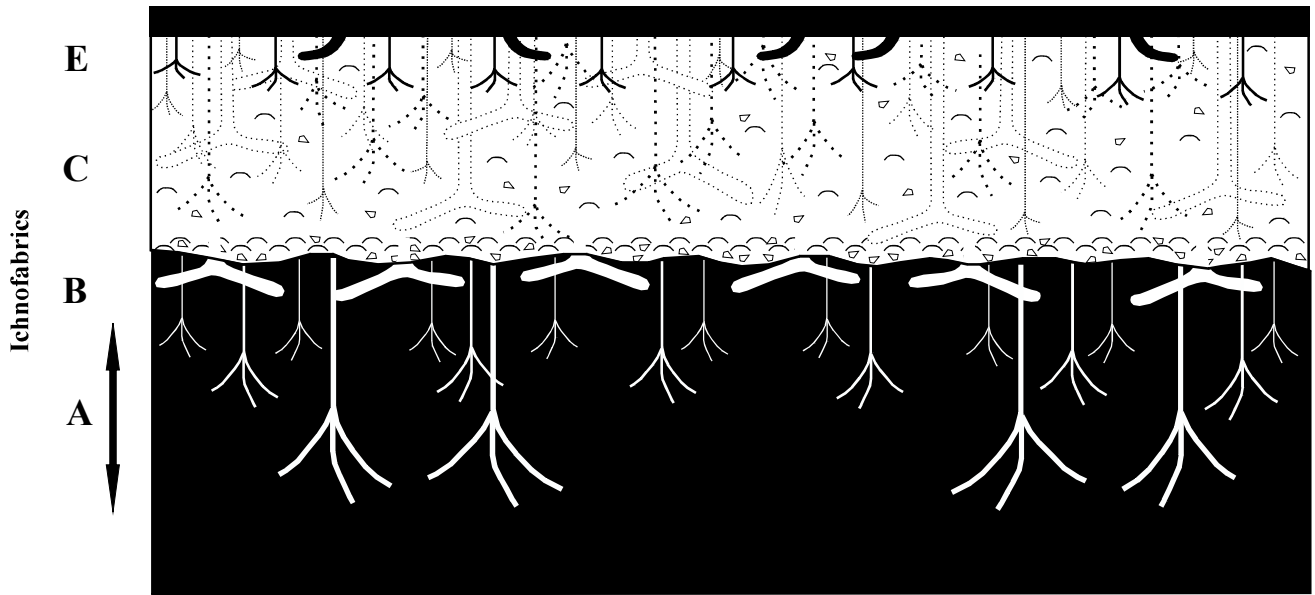
∪ *Arenicolites*   ⊌ *Diplocraterion*   ◣ *Planolites*   ⌒ Shells

⅄ *Chondrites*   ■ *Kulindrichnus*   ⊔ *Rhizocorallium*   △ Bioclasts

▼ *Conichnus*   ⅄ *Phymatoderma*   Y *Thalassinoides*

Figure 34. Main ichnofabrics identified in the Pliensbachian marl-limestone rhythmic successions

Ichnofabrics F
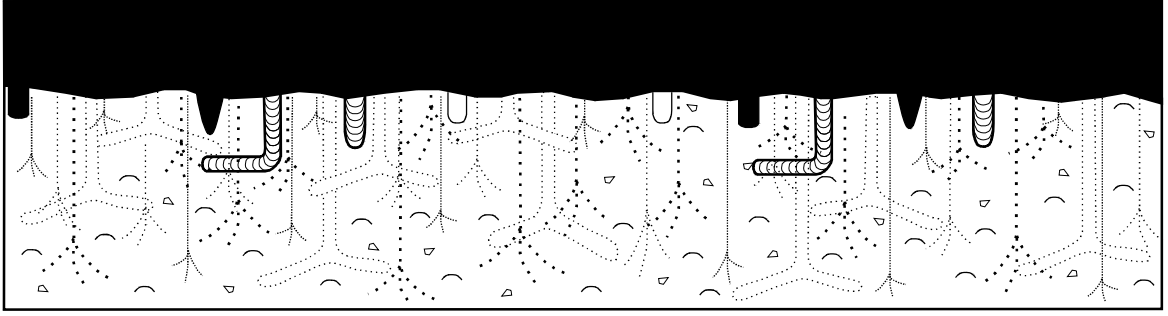
Ichnofabrics G

Ichnofabrics H

41
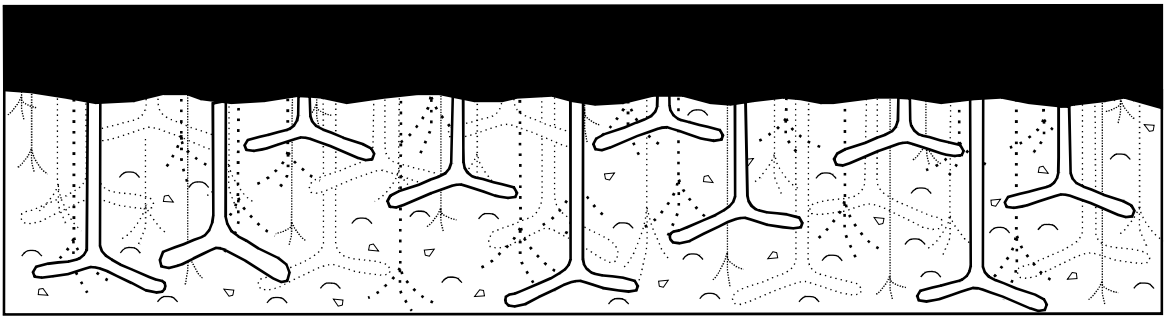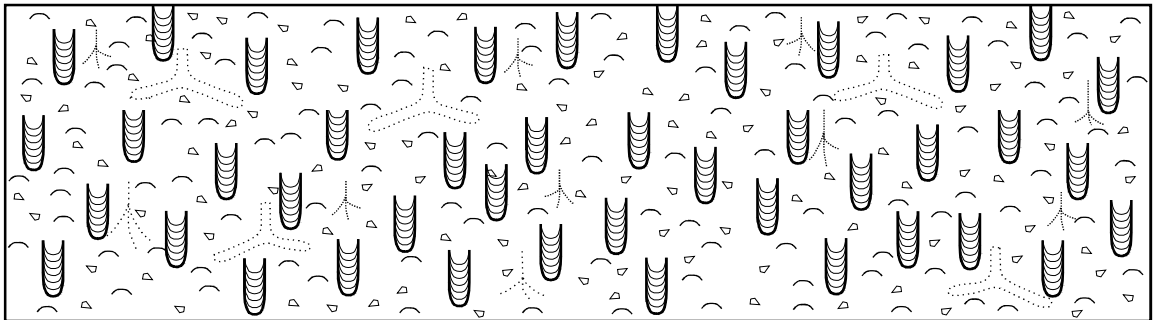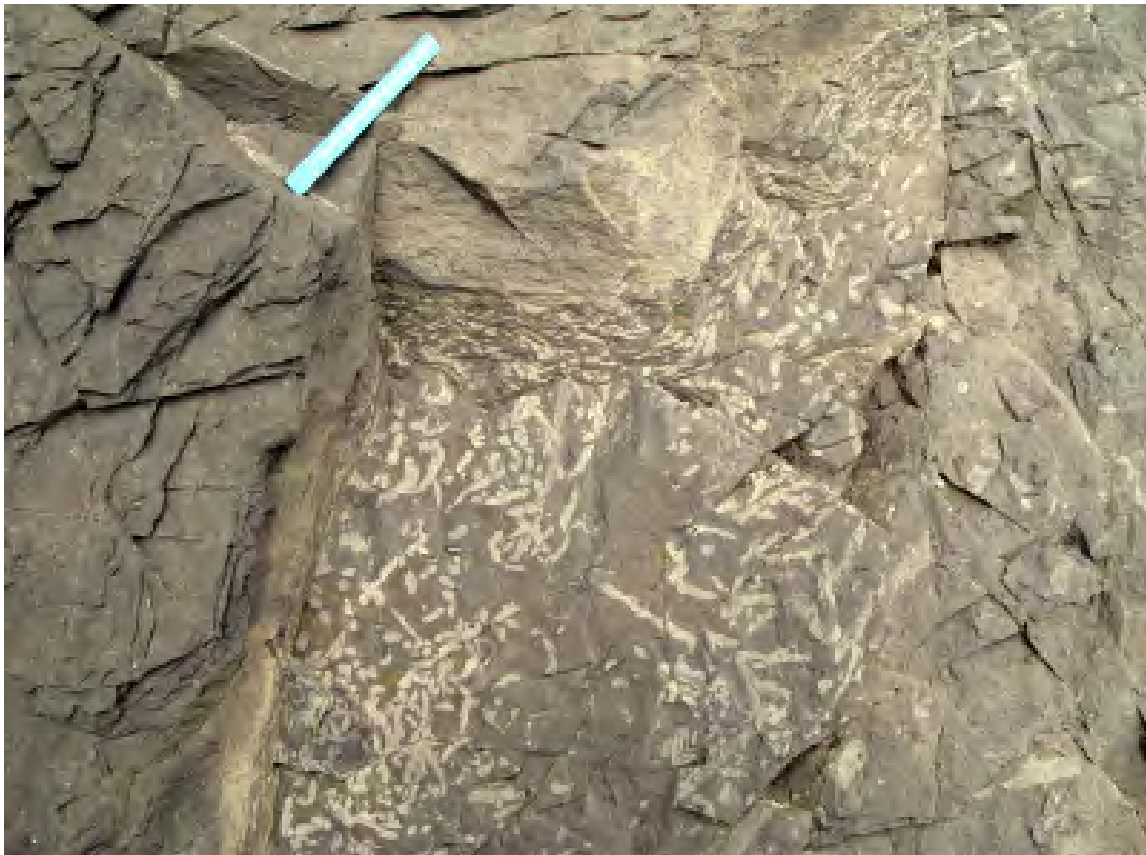
Figure 35. *Phymatoderma* specimens emplaced in a dark marl of the Jamesoni Zone. Vega beach section

Figure 36. Detailed view of *Phymatoderma* specimens showing the pelletoidal infil and the patehy distribution in the marl interbed. Vega beach section

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Figure 37. *S*imple tiering in a marl interbed. Light *Chondrites-Phymatoderma* ichnofabric. Jamesoni Zone. Vega beach section

Figure 38. Marly interval showing three different ichnofabrics. Lower level (Dark *Chondrites-Planolites* ichnofabric). Middle level (Light *Chondrites-Phymatoderma* ichnofabric). Upper level (*Asterosoma-Thalassionides* ichnofabric). Punta la Llastra section

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The overall trace fossil diversity of this ichnofabric is low, with two dominant, near exclusive, ichnogenera. Two forms of *Chondrites* have been differentiated based on size: smaller forms with less than 1 mm in diameter, and larger forms between 1 and 2 mm. Small *Chondrites* are registered in a comparatively shallow tier respect to the larger forms, always emplaced in a deeper tier. The deepest tier in this ichnofabric is colonized by *Phymatoderma* (*Chondrites granulatus* after Simpson, 1957 and Savrda *et al.*, 1991), showing tubes of 3-6 mm in diameter infilled by faecal pellets.

Distribution of ichnotaxa is variable, with *Chondrites* extended uniformly and densely into the substrate while *Phymatoderma* shows a characteristic patchy distribution. Furthermore *Chondrites* systems normally do not cross-cut one another, probably due to phobotaxis.

The Bioturbation Index (BI) after Droser & Bottjer (1989) range between 3 and 4.

Following Taylor et al. (2003), this ichnofabric corresponds to a simple tiering, and was emplaced in a soft to stiffground transition.

Occasionally, *Thalassinoides*, *Planolites*, *Trichichnus*, *Palaeophycus* and *Rhizocorallium*, are present.

The ichnofabric reveals the work of deep tracemakers colonizing a poorly oxygenated substrate from a upper level better oxygenated. This ichnofabric is similar to the short-term oxygenation events of Savrda & Bottjer (1989).These organic structures record the emplacement of transition-layer burrows (piped zones) into laminated sediments that were far enough below the sediment-water surface to prevent mixed layer bioturbation during the oxygenation episodes (Savrda, 2007). The piped zones represent also the initial phases of these storm-related oxygenation episodes.

The latest and deeper burrows such as *Chondrites* and *Phymatoderma* were emplaced into a substrate of stiff to firm consistency (see Schieber, 2003).

B. ***Asterosoma-Thalassinoides ichnofabric (Figs.9,20,22,23,29,38,39,40)***

As the previous ichnofabric, this one is characterized by light carbonate-rich filled burrows on a darker sediment background. Although this ichnofabric is easily differentiated in the marl and marly limestone intervals, its predominance at the base or into the limestone beds, difficult their recognition due to diagenetic obliteration.

Diversity of this ichnofabric is high, with the most abundant ichnogenera being *Asterosoma-Thalassinoides*, and associated *Chondrites*, *Lockeia*, *Protovirgularia*, *Planolites*, *Rhizocorallium*, *Phycosiphon*?, *Conichnus*, *Gyrolithes* and *Teichichnus*.

The usual BI is 5. This ichnofabric is usually registered in a shallower tier than the aforementioned Light *Chondrites-Phymatoderma* ichnofabric as a frozen profile of a simple tiering produced in a softground.

C. *Diffuse mottling ichofabric (Figs.22,31,41)*

Its nearly exclusive presence into the light carbonate-rich limestones, strongly affected by diagenesis, together with the intense background bioturbation, and the absence of color differentiation, difficult the identification of discrete trace fossils inthe outcrops. Even so, *Thalassinoides*, *Asterosoma* and *Chondrites* with poorly outlined contours are registered. Represent one of the different varieties of the complex tiering after Taylor *el al.* (2003).

This ichnofabric represents colonization of a softground probably in a well oxygenated hemipelagic environment with a low sedimentation rate.

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Figure 39. Simple tiering in a marl interbed belong to Light *Chondrites-Phymatoderma* ichnofabric showing a gradual transit to the *Asterosoma-Thalassinoides* ichnofabric (lower part of the limestone bed). Villar section

Figure 40. Small-scale elementary cycles in a marly interbed. The limestone bed at the top include an *Asterosoma-Thalassinoides* ichnofabric, poorly recognized due to early diagenetic carbonate cementation. Jamesoni Zone. Punta La Llastra section

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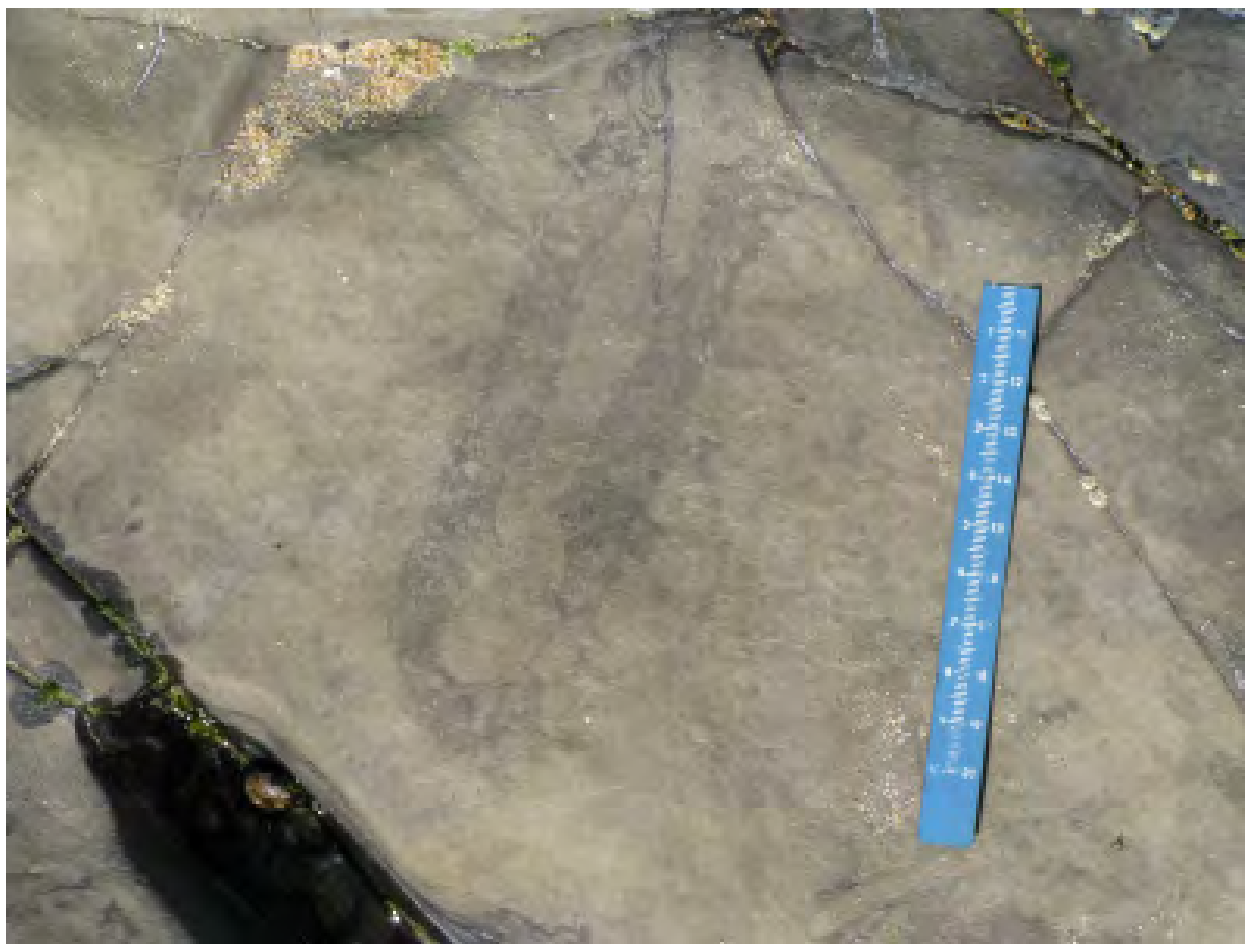
Figure 41. *Rhizocorallium* specimen with dark fill reborrowed by *Chondrites* in a Diffusse mottling ichnofabric. Margaritaus Zone. Vega beach section

D. *Complex tiering ichnofabric (Figs.21,24,25)*

This ichnofabric is mainly registered into the most carbonate-rich beds (limestones), being their outcrop differentiation difficult except when a partial late diagenetic carbonate dissolution is produced. Represent another variety of complex tiering after Taylor *el al.* (2003).

The overall trace fossil diversity is between low and moderate, being the dominant ichnotaxa *Thalassinoides*, *Rhizocorallium*, *Chondrites* and *Phymatoderma*, but locally *Planolites*, *Lockeia*, *Protovirgularia*, *Gyrolithes*, *Asterosoma* and *Conichnus* can be also recognized. Minimum deformation of ichnofossils is favoured by a early diagenesis of the limestone beds.

Bioturbation Index (BI) is between 4 and 5. This ichnofabric reveals the colonization of a multi-tiered substrate in soft to stiff conditions.

This ichnofabrics can be associated to repetitive processes of erosion and bioclastic deposition due to distal tempestite events, separated by short intervals of omission or hemipelagic deposition. Erosive processes determine a taphonomic filter with the frequent elimination of the uppermost tiers and the potential preservation of the intermedious and deeper ones bounded in the upper part by minor erosive surfaces. These surfaces are usually partially blurred due to the following colonization by tracemakers, determining the amalgamation of different phases of colonization. This complex tiering ichofabric is relatively similar to the

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extended oxygenation events of Savrda & Bottjer (1989). They are recognized by extensive cross-cutting relationships between ichnofossil varieties in the piped zones as a consequence of upward migration of the tiered infauna through time (Savrda & Bottjer, 1989).

This ichnofabric is associated, in some cases, to transgressive surfaces.

E. *Dark Chondrites-Planolites ichnofabric (Figs.30,38,42)*

This ichnofabric, generally with small vertical expansion, is characterized by dark filled burrows on lighter carbonate-rich and highly bioturbated sediment background. It is usually registered in the upper parts of both limestone and bioclastic calcareous marls of distal tempestitic origin, located below dark gray laminated marls.

Overall trace fossil diversity is low and the Bioturbation Index (BI) is 2.

This ichnofabric could be related to the episodes of "oxygen deterioration" in Taylor *et al*. (2003) or to the "gradual deoxygenation events" in Savrda & Bottjer (1989). The colonization of tracemakers occurs in soft to stiffground conditions.

The record in the case study of dark *Chondrites* in well oxygenated and light substrates prevents on the generalized use of this ichnogenera as indicative of anoxic conditions into sediments according to proposal from Bromley & Ekdale (1984) as well as revealing a chemiosymbiotic behavior as indicated by Savrda *et al*. (1991).
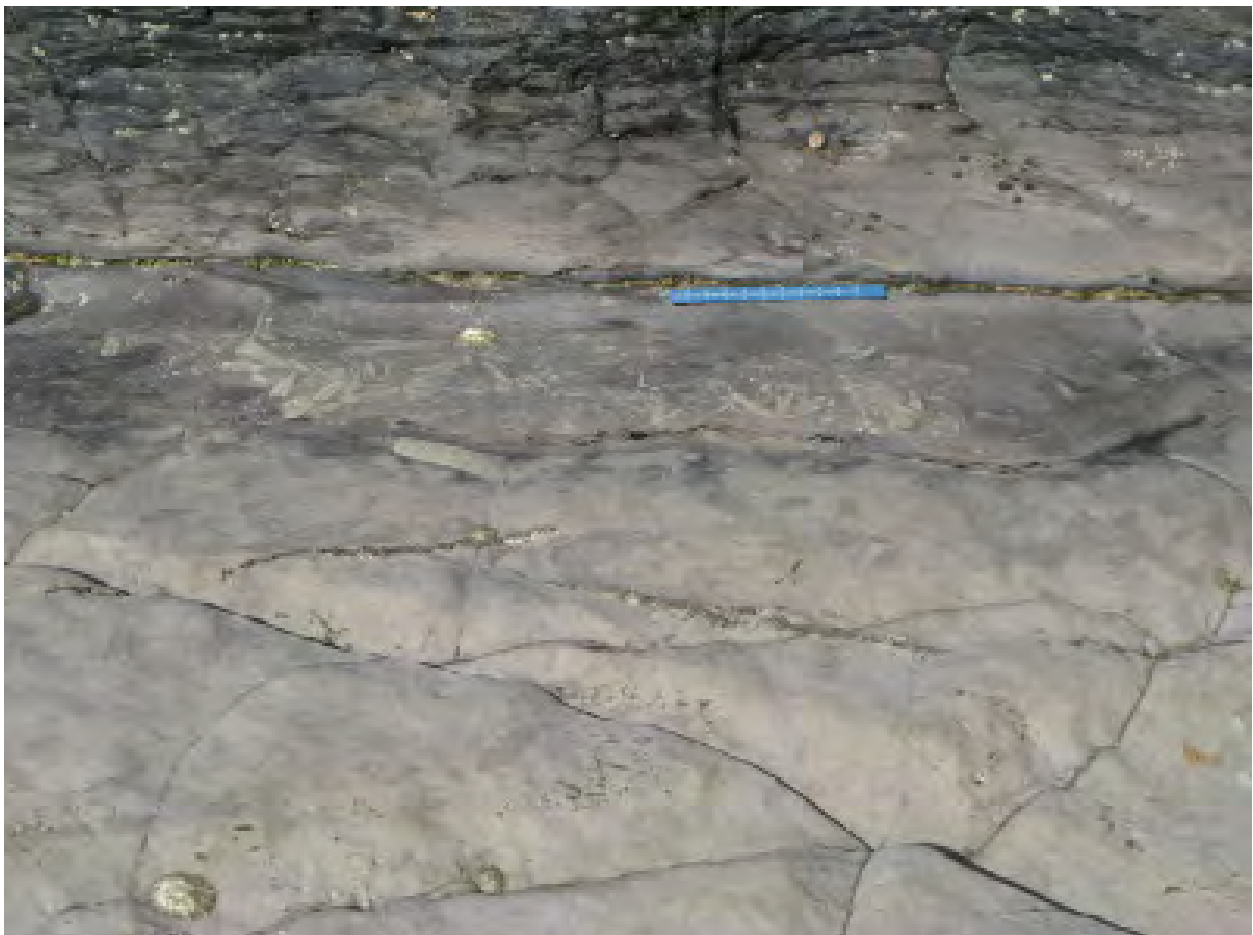

Figure 42. Belemnite concentration in a scoured storm bed cutting an highly bioturbated marly level. Margaritatus Zone. Vega beach section

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F.   *Kulindrichnus-Rhizocorallium ichnofabric (Fig.43)*

This ichnofabric is near exclusively registered at upper part of the limestone beds associated there to planar erosive surfaces just below the usually laminated dark marly intervals. Thus, the ichnofabric represents the emplacement of dark filled burrows in lighter carbonate-rich and highly bioturbated host sediments.

The overall trace fossil diversity is low and the Bioturbation Index (BI) is 2.

Several features allow the attribution to stiffground to firmground conditions, as: the ichnofabric composition, including the presence of *Conichnus*, *Diplocraterion* and *Arenicolites*, the undeformed burrows structures, their sharp boundaries and the truncation of previous soft bottom burrows. Thus, probably a superposition of the *Glossifungites* ichnofacies to a previous *Cruziana* ichnofacies can be interpreted.
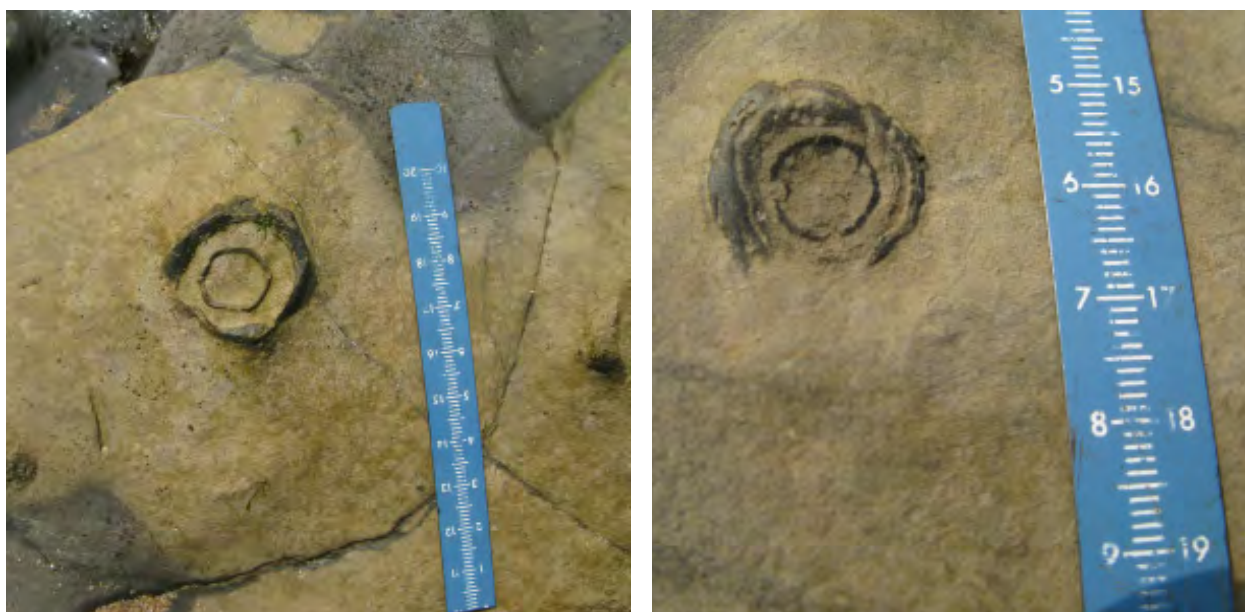
Figure 43. Phosphatized *Kulindrichnus* in the erosive top surfaces of limestone beds. *Kulindrichnus-Rhizocorallium* ichnofabric

G.   *Large Thalassinoides ichnofabric (Figs.26,27,28,44)*

This ichnofabric is registered in calcareous marls and limestone beds, being the best examples those from beds 98, 99, 101, 102, and 103 (Margaritatus Zone).

The overall trace fossil diversity is low and the Bioturbation Index (BI) is between 3 and 4. *Thalassinoides*-like structures are prominent together with some *Diplocraterion* and *Arenicolites*.

In the base of the limestone beds, very large *Thalassinoides* (*Spongeliomorpha*?) are prominent, with tube diameters between 4 and 10 cm, while into the marly levels the horizontal branches show a slight flattening by compaction. The *Thalassinoides* structures are robust, sharp-walled and unlined, commonly penetrating up to 15-20 cm below the associated erosive discontinuities. The latter are located in the boundary between the hosting highly bioturbated light calcareous marl or limestone beds and the overlying dark gray laminated marls.

Figure 44. Bioturbated marly interbed (light colour) between laminated dark marls. Note the planar erosive top of the light marl. Large *Thalassionides* ichnofabric (*Glossifungites* ichnofacies). Margaritatus Zones. Vega beach section

The size of structures could be related with adult producers (crustaceans) that survived transport by storm currents (allochtonous tracemakers) and renew their burrowing activities in the new deeper environment. They are the so-called "doomed pioneers" of Föllmi & Grimm (1990) and Grimm & Föllmi (1994). These biogenic structures produced by allochtonous crustaceans have been recognized also by Fürsich (1998), Jensen & Atkinson (2001), Pemberton *et al.* (2001), Savrda & Nanson (2003), and Savrda *et al.* (2010).

The sharp-walled biogenic structures created by these crustaceans (probably thalassinid decapods), cross-cut an exhumed previous community, reflecting the new colonization of a firm substrate (*Glossifungites* ichnofacies).

The origin of these marine firmgrounds with *Thalassinoides* may be related to: a) incipient synsedimentary cementation of carbonates during periods of sediment bypass and/ or non deposition, or b) erosional exhumation of compacted muds (Savrda, 2007). In these Pliensbachian examples here studied the fill of *Thalassinoides* is cast by carbonate-rich light sediments located below a storm-generated erosion surface associated to bypass processes and covered finally with laminated dark marls.

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H. *Diplocraterion-Arenicolites ichnofabric (Fig.12)*

This ichnofabric is mainly located in the upper part of the rhythmic succession, between beds 111 and 132, into the *Spinatum* Zone, in coincidence with a well established shallowing phase, recognized in all of West European domain.

The overall trace fossil diversity is very low and Bioturbation Index (BI) is between 3 and 4. Trace fossil assemblage is mainly represented by protrusive *Diplocraterion*, showing a high density in both marly and limestone bioclastic beds (wackstone to packstone). Variable colour infilling is recognized, with dark specimens into light-surrounding material and viceversa. Other scarce ichnotaxa are *Thalassinoides* and *Chondrites*, the latter sometimes penetrating the marginal limbs of *Diplocraterion*.

Shell concentrations dominated by belemnites and bivalves (pectinids) are significantly registered. These shells show frequently borings.

This ichnofabric represents a succession of repetitive firmgrounds (*Glossifungites* facies) overprinting a softground community. The firmgrounds are associated to repeated small wave-reworked erosion surfaces.

Stop 2. Villar sea cliffs (Villaviciosa). Reworked hardground ichnofabrics at the top of regressive small-scale cycles. Upper Sinemurian (Lower Jurassic) successions in a inner ramp characterized by shallow water nodular limestones

Marl-limestone rhythmic successions (Rodiles Formation) are common in the Lower and Middle Jurassic sea cliffs of the Asturian eastern coast. The Rodiles Formation consists of two members: the lower one is the Buerres Member and upper one the Santa Mera Member (Valenzuela *et al*., 1986) (Fig.4).

The Buerres Member consists of nodular limestones with thin marly interbeds and shows several semi-lithified firmgrounds and hardgrounds capping small-scale (decimeters to meters) regressive cycles (R cycles after Zecchin, 2007) in response to high-frequency sea-level changes (Fig.45). The whole accumulated in a shallow-water inner carbonate ramp.

The outcrop proposed for observation and study is located to the east of the Villaviciosa estuary (Fig.1).

The matter for discussion in this outcrop are several loosen calcareous blocks fallen of the sea cliffs and showing well preserved ichnofabrics enclosed into firmground and hardground levels (Valenzuela *et al*., 1992) (Fig.46).

These omission surfaces cover a succession of nodular and cross-stratified pelmicrites forming wave-ripples as well as hummocky beds of storm origin (Fig.47). Other characteristic structures placed below the hardground are synsedimentary deformations (convoluted beds and slumps) probably related to storm-induced wave loading processes which represent a good marker horizon for stratigraphic correlation in the Asturian Lower Jurassic series (García-Ramos *et al*., 1989, 1992).

Figure 45. Small-scale regressive cycles topped by omission surfaces and hardgrounds in an outcrop of the Upper Sinemurian rocks located 550 m to the west of the Stop 2

Figure 46. Reworked hardground at the top of a small-scale regressive cycle. Calcareous block fallen in the base of the sea cliffs

Figure 47. Cross-stratified pelmicrites forming wave-ripples located lightly below the hardground

Once formed, the firmgrounds and hardgrounds were erosively reworked and broken by waves and storms in platy limestone fragments (Fig.48) that were submitted to several and repetitive processes of biological encrustation (oysters) (Fig.48) and boring by polychaetes and bivalves (Figs.49,50).

Many of the bivalve borings (*Gastrochaenolites*) were probably emplaced in firmgrounds and are incomplete in their upper part due to erosional processes (Figs.49,50).

The platy and rounded limestone fragments are traversed by multiple cracks, eroded *Gastrochaenolites*, small vertical tubes (*Trypanites*?, *Skolithos*?) and covered by encrusting oysters indicating extended exposure on the sea bottom prior to burial.

There are also no articulated valves in the oysters attached to hardground, therefore these encrusters were dead before final burial.

The small cracks connect in an irregular manner, but no show polygonal patterns (Figs.49,50).

The origin of the cracks is problematic but their characteristics suggest, at least for many of them, an emplacement during the firmground stage (i.e.: the *Gastrochaenolites* and other vertical small biogenic structures determinate partially the orientation and termination of the cracks) (Fig.50).

Demicco & Hardie (1994) mentioned in a Lower Ordovician succession of western Maryland (USA), firmgrounds with an irregular erosional surface at the top of wave-ripple peloidal carbonates overlained by a flat pebble conglomerate. The pebbles have also rounded corners and have been interpreted as intraclasts ripped up from the firmground by storm-generated bottom currents. Demicco & Hardie (op. cit.) consider that firmgrounds can be easily broken up and transported as cohesive intraclasts.

Figure 48. Erosively reworked platy and rounded limestone fragments of the hardground. Note the borings, the cracking and the encrusting oysters

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Figure 49. Eroded *Gastrochaenolites* borings, some of them infilled by echinoderm skeletal fragments and small lithoclasts

Figure 50. Cracks controlled partially eroded by *Gastochaenolites* and other small biogenic structures

If these conclusions are correct, it is necessary to suppose that the *Gastrochaenolites* and the small vertical biogenic tubes (similar to *Trypanites*) and probably due to polychaetes, were emplaced during the firmground stage (*Skolithos*?), because many cracks postdate and displace the above-mentioned biogenic structures (Figs.49,50). An alternative interpretation, although more dubious, may be invoked, supposing that some *Trypanites* were emplaced in firmgrounds as states Goldring & Kaźmierczak (1974) and Gruszczyński (1986).

The *Thalassinoides* networks excavated in the firmground are typically infilled by small lithoclasts and bioclasts (Fig.51), the latter composed of broken shell fragments and echinoderm skeletal debris (Fig.49).

Figure 51. *Thalasinoides* networks (firmground stage) infilled by lithoclasts and bioclasts in a loosen block

Stop 3. El Puntal sea cliffs (Villaviciosa). Rhizocorallium-dominated ichnofabrics in marl-limestone alternations of an Upper-Sinemurian succession

El Puntal outcrop is located in the western margin of the Villaviciosa estuary (Fig.52). This stop is focused on several aspects related to *Rhizocorallium*-dominated ichnofabric enclosed into the marl-limestone alternations of the Buerres Member (Rodiles Formation) (Figs.4,53), dated as Upper Sinemurian, Obtusum Zone (Suárez Vega, 1974; Comas-Rengifo & Goy, 2010).

The *Rhizocorallium* ichnofabric is settled into the thin transgressive lower part of a small-scale dominant regressive cycle (R cycles of Zecchin, 2007). These cycles are asymmetrical and composed of a dominant regressive part, whereas transgressive strata are relatively thin. R cycles are similar to classical parasequences but without the interpretation problems associated to them (Walker, 1992; Arnott, 1995; Miall, 1997; Catuneanu, 2006; Zecchin, 2007).

Thinner transgressive intervals may be produced in these nearshore and low-gradient topography settings, if erosion and sediment bypass prevail, the transgression is rapid, or in sediment starved conditions (Zecchin, 2007).

The characteristic L-shaped *Rhizocorallium* specimens of this outcrop are enclosed into a marly interval placed between 35 and 50 cm above a hardground surface (Fig.54) representing the base of R cycle. The regressive part of it consists of nodular limestone beds and lens (pelmicrites) with wave-ripples and storm-related hummocky cross-stratification, alternating with marls.

Figure 52. Location of the El Puntal outcrop into the western margin of Villaviciosa estuary

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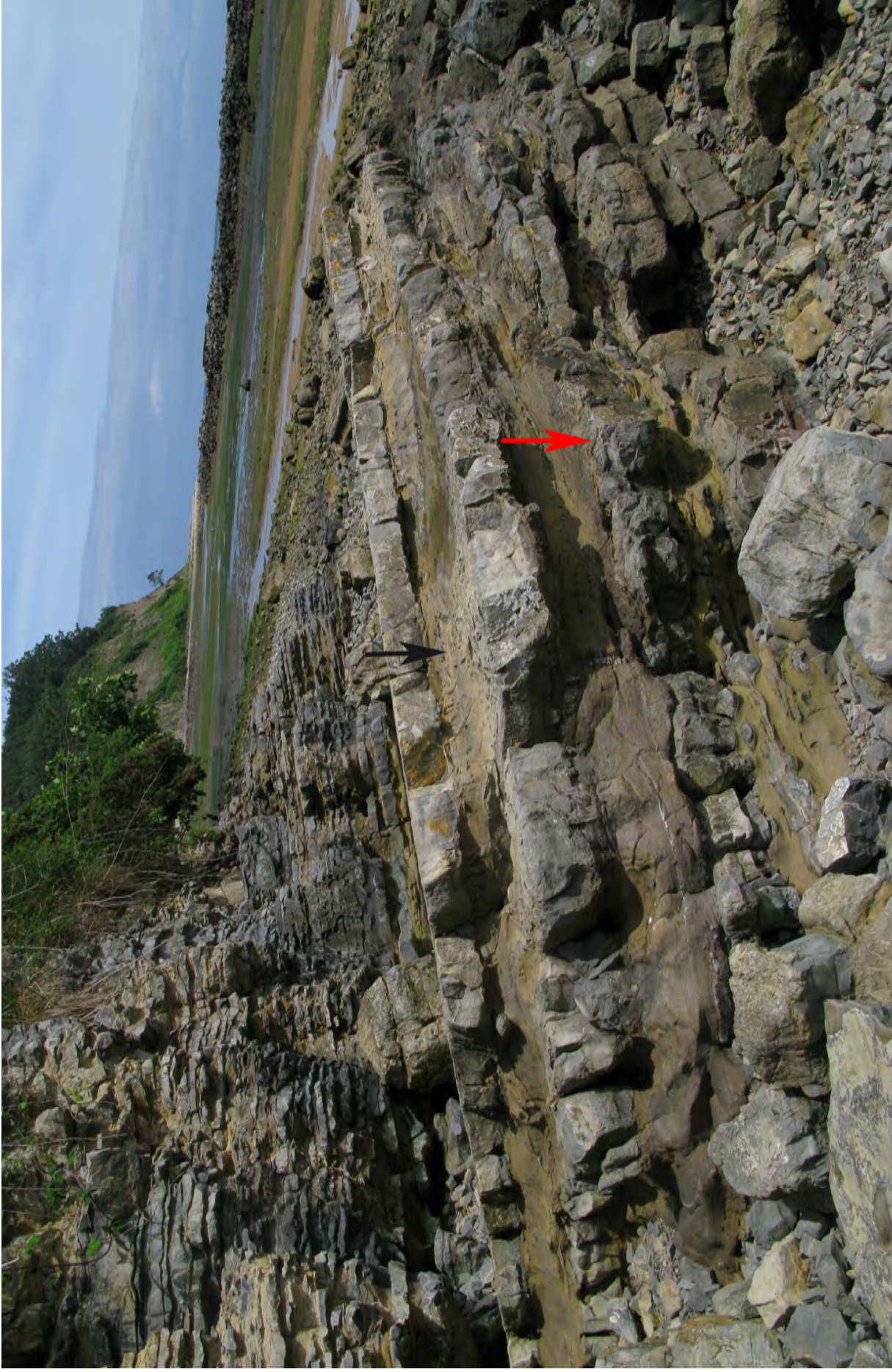
Figure 53 . Aspect of the El Puntal outcrop showing a small-scale regressive cycle with a hardground in the base (red arrow). The main level with L-shaped *Rhizocorallium* specimens (black arrow) is located into the relatively thin transgressive interval

Figure 54. View of a reworked hardground correlatable with the same level showed in the Stop 2

Some L-shaped U-tubes of *Rhizocorallium* were filled with micritic limestone and intensely reburrowed by *Chondrites* (Fig.55), but in many others, the fill is bioclastic and lithoclastic as well as the spreite (Figs.56,57,58). The sharp contract with hosted marls, local evidence of scratch-marks and nearly circular sections of tubular limbs, suggest a setting in a firmground (*Glossifungites* ichnofacies).

Figure 55. Segment of a *Rhizocorallium* limb reborrowed by *Chondrites*

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Figure 56. L-shaped U-tubes of *Rhizocorallium* filled with bioclastic and lithoclastic sediments. Note in the upper picture the flattened variety with protuberances emplaced in an upper tier together with *Chondrites*

Figure 57. Horizontal segment of a *Rhizocorallium* specimen. Notice the bioclastic and lithoclastic fillin the spreite

Figure 58. Retrusive *Rhizocorallium* above the scale showing also an increasing in size towards the distal part (*Glossifungites* ichnofacies)

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The tracemakers of *Chondrites* may have preferentially exploited *Rhizocorallium* burrow fills perhaps because, compared to hosted marls, these fills either had greater nutrient value or had sediment consistences that were more amenable for feeding and/or locomotion (Locklair & Savrda, 1998).

Some other references on *Chondrites* fills in *Rhizocorallium* limbs include to Simpson (1957), Farrow (1966), Sellwood (1970) and Moghadam & Paul (2000).

Several specimens of *Rhizocorallium* in this outcrop show an increasing in size towards the distal part (fan-shaped) (Fig.58). This is probably related to the growth of the tracemaker (Uchman *et al*., 2000). The latter was at least in many occasions a crustacean as suggested by scratch-mark impressions (Sellwood, 1970; Fürsich, 1974; Schlirf, 2003; 2011; Seilacher, 2007; Rodríguez-Tovar & Pérez Valera, 2008).

The characteristic L-shaped *Rhizocorallium* specimens (Fig. 56) of this outcrop are not included in the new classification concept for U-shaped spreite trace fossils (Schlirf, 2011).

The relatively large vertical extent of the U-limbs in the proximal part, remember the behaviour of the protrusive *Diplocraterion parallelum* tracemaker and probably was induced by erosion processes associated to transgressive episode. Apparently a single biogenic U-structure may represent in different times, the activities of a mixed sediment and suspension feeder crustacean.

In the same surface of the L-shaped U-tubes of *Rhizocorallium* it is possible to see *Rhizocorallium irregulare* (Fig. 59)

Other particular specimens of *Rhizocorallium* located into the same marls up to 2 cm below the limestone base, show horn-shaped protuberances on the external part of the flattened and limestone-filled U tubes. These acute and asymmetrical ornamentations are regularly spaced and may be produced by the appendages of the crustacean tracemakers during the digging activities (Figs.P60) (work in progress).

Figure 59. Apecimen of *Rhizocorallium irregulare*
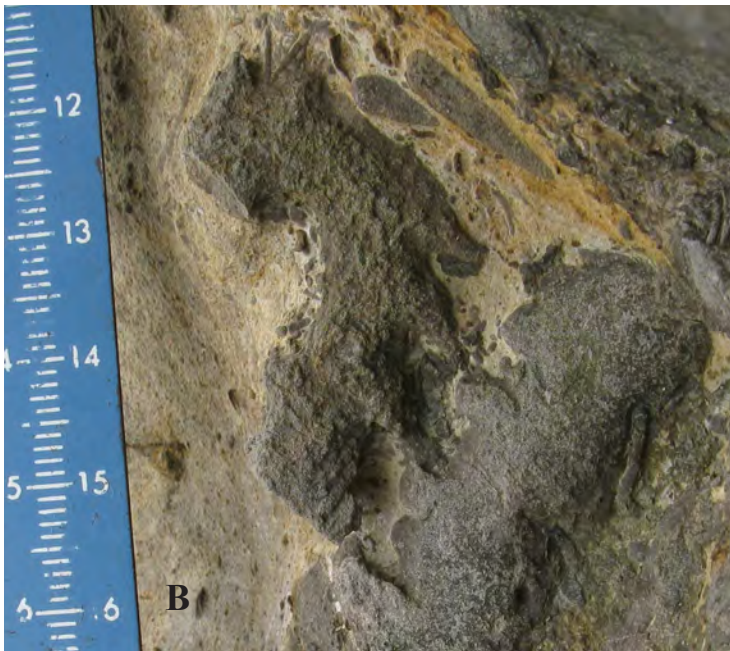
Figure 60. Horizontal segments of the flattened *Rhizocorallium* variety showing the horn-shaped protuberances in the outer part of the tubes. **(A)** and **(B)** El Puntal outcrop. **(C)** Another specimen in a correlatable level located in the sea cliffs 2 km to the east

Stop 4. La Griega beach (Colunga). Dinosaur tracks, invertebrate trace fossils and ichnofabrics in the transition between continental, littoral and restricted shelf lagoon facies (Kimmeridgian, Upper Jurassic)

La Griega beach tracksite is located in the center of the Dinosaur Coast (Figs.1, 61) and is known mainly because has one of the biggest dinosaur footprints in the Jurassic fossil record.

There are two approach roads to the tracksite, the first one, to walk by the eastern edge of the beach, about 650 m, and the second one, to walk by an indicated path which runs between the beach and the forest. This path, providing spectacular views of the beach, the Lastres bay and harbor and the Jurassic Museum of Asturias, ends in a wood staircase which goes down straight to the tracksite, even during a high tide periods. At the end of the path there is a explanatory and detailed panel.

Figure 61. Location of La Griega tracksite.

The succession represented in the Figs.62.63 starts with a highly burrowed sandstone of crevasse-splay origin accumulated in a floodplain-coastal plain transition area.

The ichnofauna of this sandstone bed is characterized by a low-diversity and dominated by meniscate burrows of the *Scoyenia* ichnofacies (Buatois *et al*., 2002). The burrows apparently do not show any preferential orientation.

Two main biogenic meniscate structures assigned to *Beaconites* (with lining) and *Taenidium* (without lining) are recognized (Fig.64). The diameter of the lumen in both ichnogenera is between 6 and 16 mm with an average of 11 mm. The wall lining of *Beaconites* is composed of a dark carbonate-cemented sandstone which is often currently wheathered by decalcification on the upper surface of the sandstone bed. This highly burrowed sandstone level is laterally recognized in several neighbouring outcrops (Figs. 65,66).

These overbank deposits agree with *Beaconites-Taenidium* ichnofabric mentioned by Buatois & Mángano (2007) and the colonized substrate was probably a looseground.

The Bioturbation Index (BI) after Drose & Bottjer (1989) is between 4 and 5.

Meniscate structures similar to *Beaconites* and *Taenidium* in continental environments have been attributed to different organisms: earthworms or crayfish (Bedatou *et al*., 2009), insects (Uchman & Alvaro, 2000; Hasiotis, 2002; Smith *et al*., 2008) and terrestrial or aquatic oligochaetes (Bown & Kraus, 1983; Uchman & Alvaro, 2000).

On the top of the sandstone bed with *Beaconites-Taenidium* it is possible to see several shallow theropod footprints with a poor preservation (Fig.67). The most interesting are two parallel tracks with metatarsal impressions due to, probably, a resting position of the theropod.

None of the dinosaur footprints are affected by invertebrate ichnofossils, so they were produced after the last.

Figure 62. View of stratigraphic succession represented in Fig. . The bed surface with big sauropod tracks is indicated by red arrows (transgressive surface)
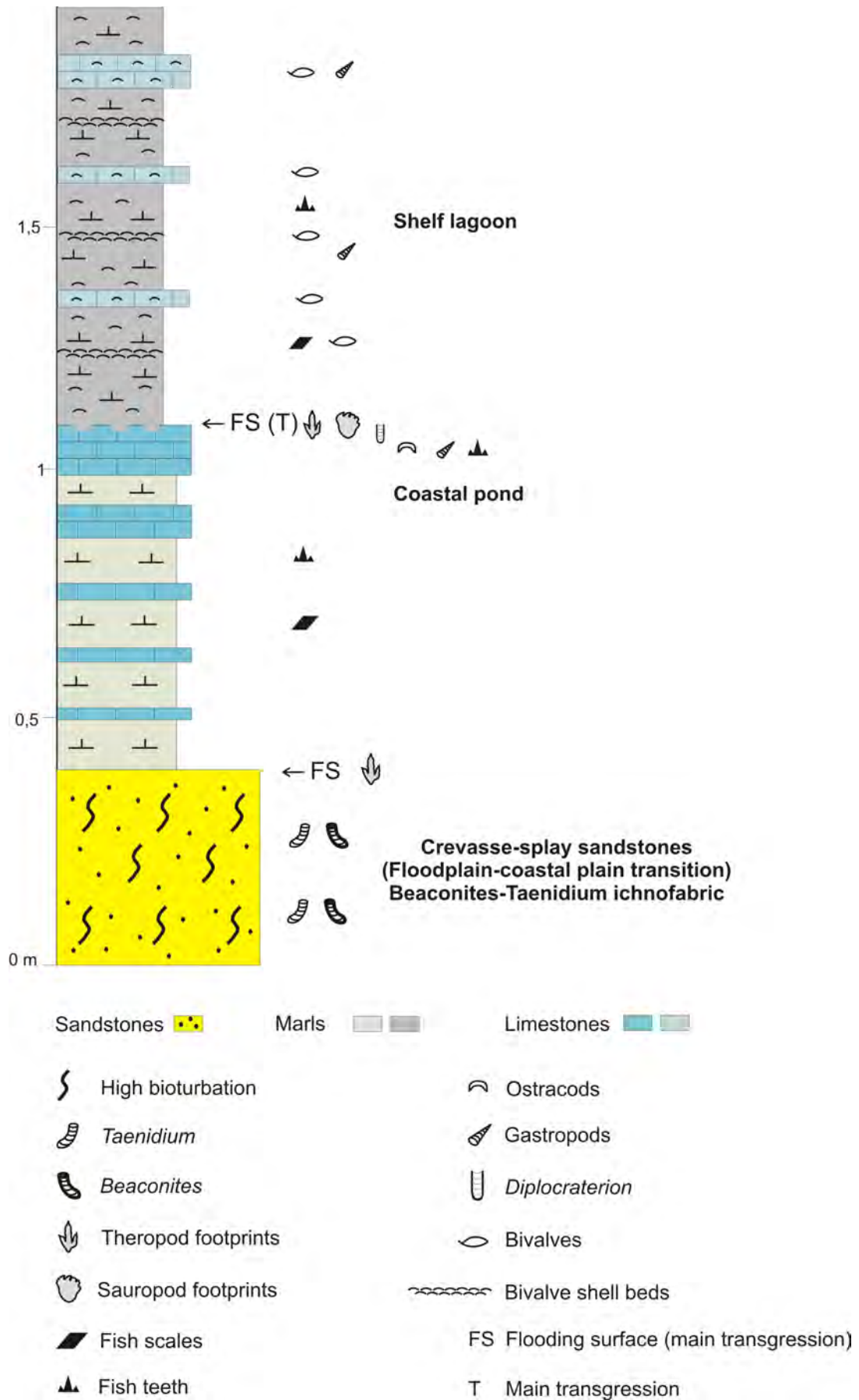
Figure 63. Studied section in the Tereñes Formation showing the levels with *Beaconites-Taenidium* ichnofabric and dinosaur footprints.

Figure 64. Verticalized tubes (*Beaconites*) on the upper surface of a highly bioturbated grey sandstone. Flooding surface on a crevasse-splay deposits

Figure 65. Aspect of the same fluvial-lagoonal transition facies in other outcrop situated 650 m to the west on the left margin of La Griega beach. Note in the lower part (pink colours) a palaeosol with vertical root traces in a floodplain deposit

Figure 66. Detail of the preceding picture (see arrow) showing the highly bioturbated *Beaconites-Taenidium* ichnofabric.

Figure 67. Trydactil dinosaur footprints in the same surface with *Beaconites*. **(A)** Two parallel theropod footprints preserving the metatarsal impressions. **(B)** Tridactyl footprint belongs to a left foot of a medium-size theropod dinosaur; notice the digit pads and claw marks.

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Above the intensely bioturbated sandstone, there is a small marl-limestone interval culminating in a micritic grey limestone with dinosaur footprints, belong to quadrupeds (sauropods) and some to bipeds (theropods) Fig. ; these last more difficult to recognize. The footprints were produced under shallow water when the dinosaurs walked in a coastal pond.

The biggest sauropod trackway (Figs.68,69) consists of four relatively complete manus-pes sets and appears to have a relatively narrow gauge, with a maximum trackway width of only 16 cm between the inner margins of the pes. However, all these measurements are tentative given the sub-optimal preservation.

The shape of pes seems to be typical of sauropods, rounded or semicircular in outline, and they range in length from 98 to 125 cm with corresponding widths of 73-109 cm (mean length and width = 110.75 and 89.5 cm, respectively). The two best preserved pes prints present a elevated mud rim. The diameter of some of them, 125 cm, place them among the biggest known up to now.

The best manus prints preserved has a crescentic shape. Manus measurements are less complete and reliable due to poor preservation and the apparent interference of the pes track. Nevertheless, width measurements are fairly consistent, between 55 and 62 cm (mean = 59 cm), with length measurements between 32 and 44 cm (mean = 38 cm).

Both manus and pes show no individual digit impressions. The ratio between the areas of manus and pes may give an approximate indication of the manus/pes size ratio: i.e., the heteropody value is between 1:6 and 1:5.

Figure 68. **(A)** Trackway attributed to a sauropod of extraordinary size. **(B)** Wide gauge small sauropod trackway (*Brontopodus*). The arrows show the direction movement.
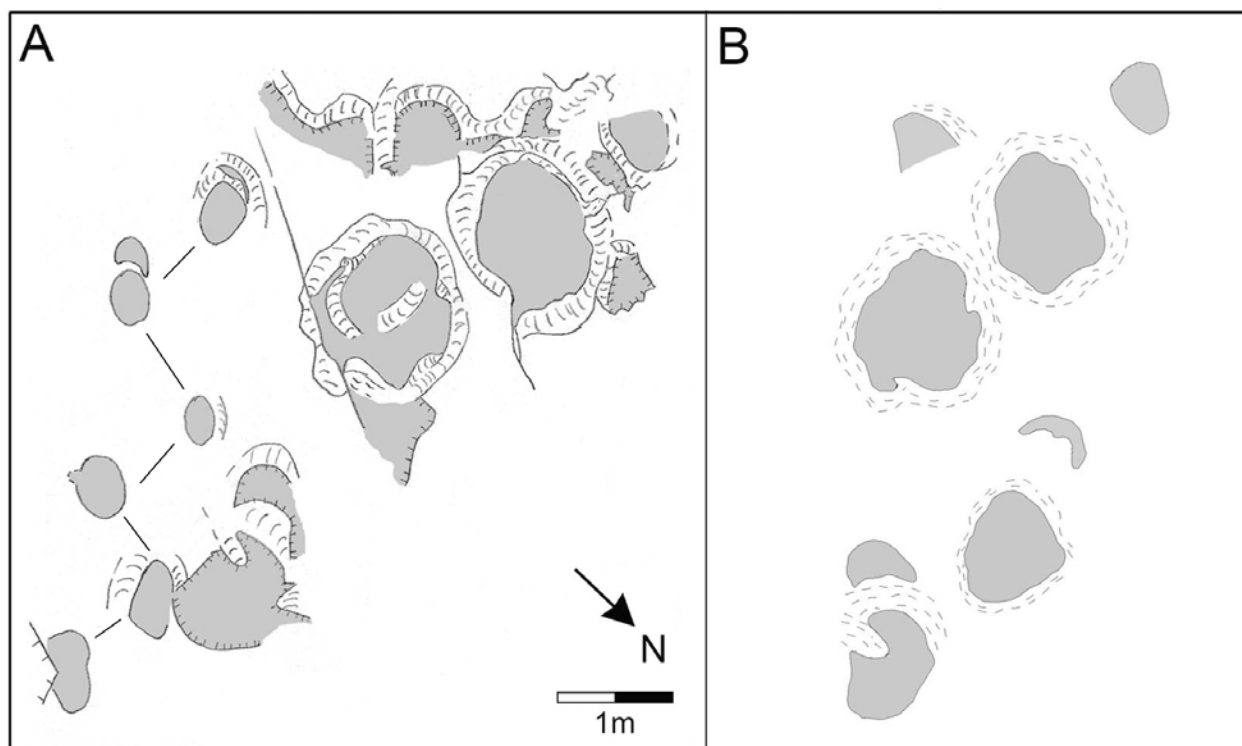
Figure 69. **(A)** Line drawings of the *Gigantosauropus asturiensis* and *Brontopodus* trackways, after Lires *et al.* (2001). **(B)** Simplified drawing of *Gigantosauropus asturiensis* after Lockley *et al.*, 2007. Dashed lines represent mud rims.

This trackway was described for the first time by Mensink & Mertmann (1984). They assigned these large round tracks, purportedly up to about 1.35 meters in maximum diameter, to the new ichnogenus and ichnospecies *Gigantosauropus asturiensis* and attributed them to a theropod dinosaur, to be precise to an megalosaur (Megalosauroidea). As noted by a number of authors (Thulborn, 1990; Lockley *et al.*, 1994; García-Ramos & Gutiérrez Claverol, 1995; Leonardi, 1997; Lires *et al.*, 2001; García-Ramos *et al.*, 2002; Lockley *et al.*, 2007; 2008), these are tracks of sauropods, not theropods.

The smallest quadrupedal trackway (Figs.68,70A,B), adjacent to *Gigantosauropus asturiensis*, consists of 8 prints, 6 pes and 2 manus, and appears to have a relatively wide gauge. The attribution to *Brontopodus* by Lires *et al.* (2001) is justified on the basis of the wide gauge and relatively large manus (Fig.70A).

The pes prints have an oval shape and they are range between 43 and 63 cm in length (this last measurement is excessive due to partial overlapping of pes above manus). One of them (H.h in the Fig. 70A,C) was attributed incorrectly to a theropod (Fig.70C) by Mensink and Mertmann (1984) and assigned as paratype of *Hispanosauropus hauboldi*, described by the same researchers on the Ribadesella sea cliffs for a tridactyl footprint.

The best manus print preserved has the typical outline in a crescent shape (Fig.70B) and is 21 cm in length and 29 cm in width. This manus print is located slightly in front of the pes prints (Fig.70B). In this case, both manus and pes do not show either digit impressions.

Leonardi (1997) illustrated the tracks of this surface indicating that it belong to an ichnoassociation that represent trampling by the passage of a large herd of animals. Lires *et al.* (2001) proposal at least two animals passed in approximately the same southwesterly direction, the larger being the *G. asturiensis* type specimen, and the smaller *Brontopodus*.

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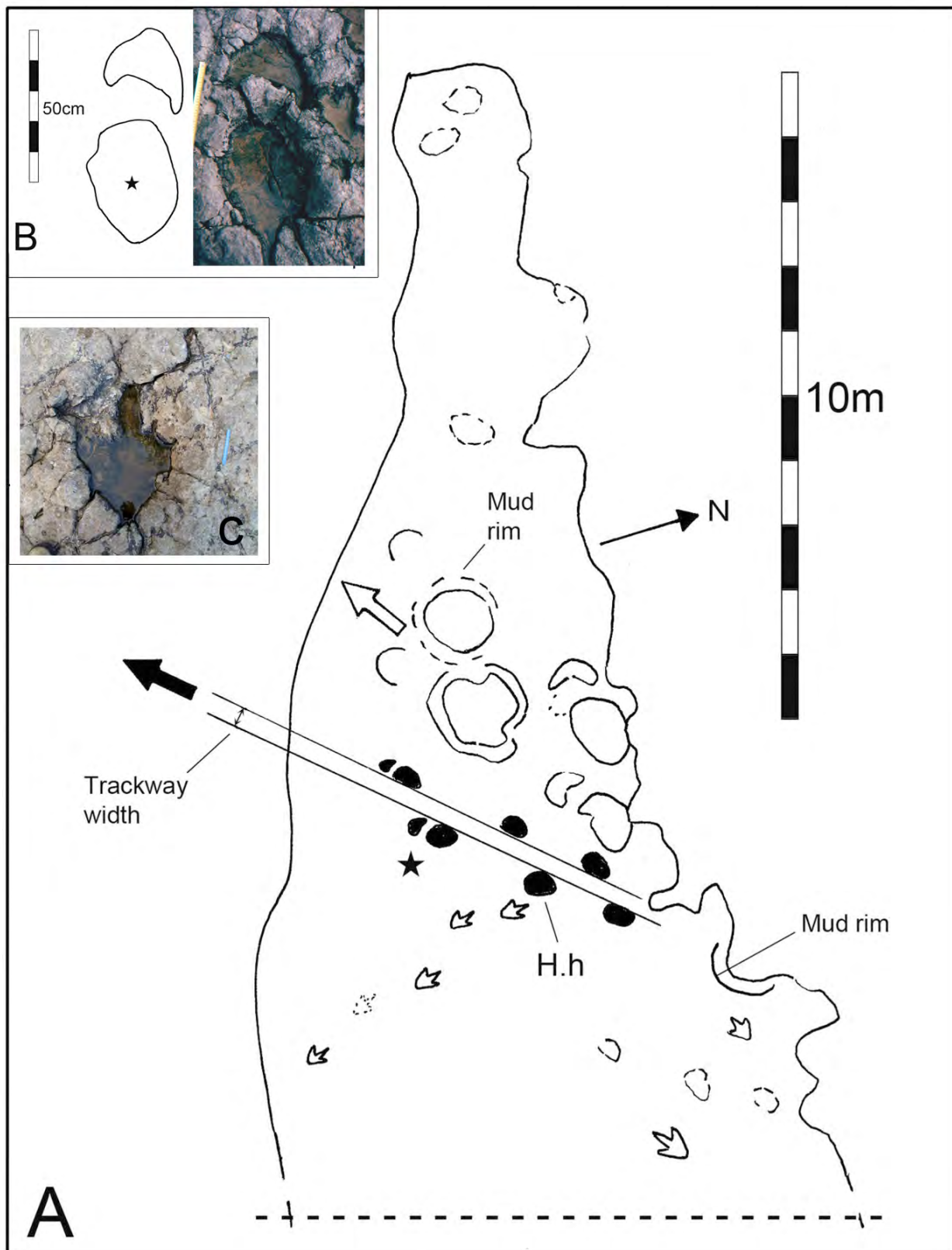**Figure** . **(A)** Map of the track-bearing surface with *Brontopodus* trackway in black showing trackway width. **(B)** The last (distal) left manus-pes set (star in the trackway). **(C)** *H. hauboldi* paratype (H.h in the trackway); notice the trydactil aspect produced by the joints and present cracking in the pes footprint. Modified from Lockley *et al*., 2007.

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