

MICROPALAEONTOLOGÍA

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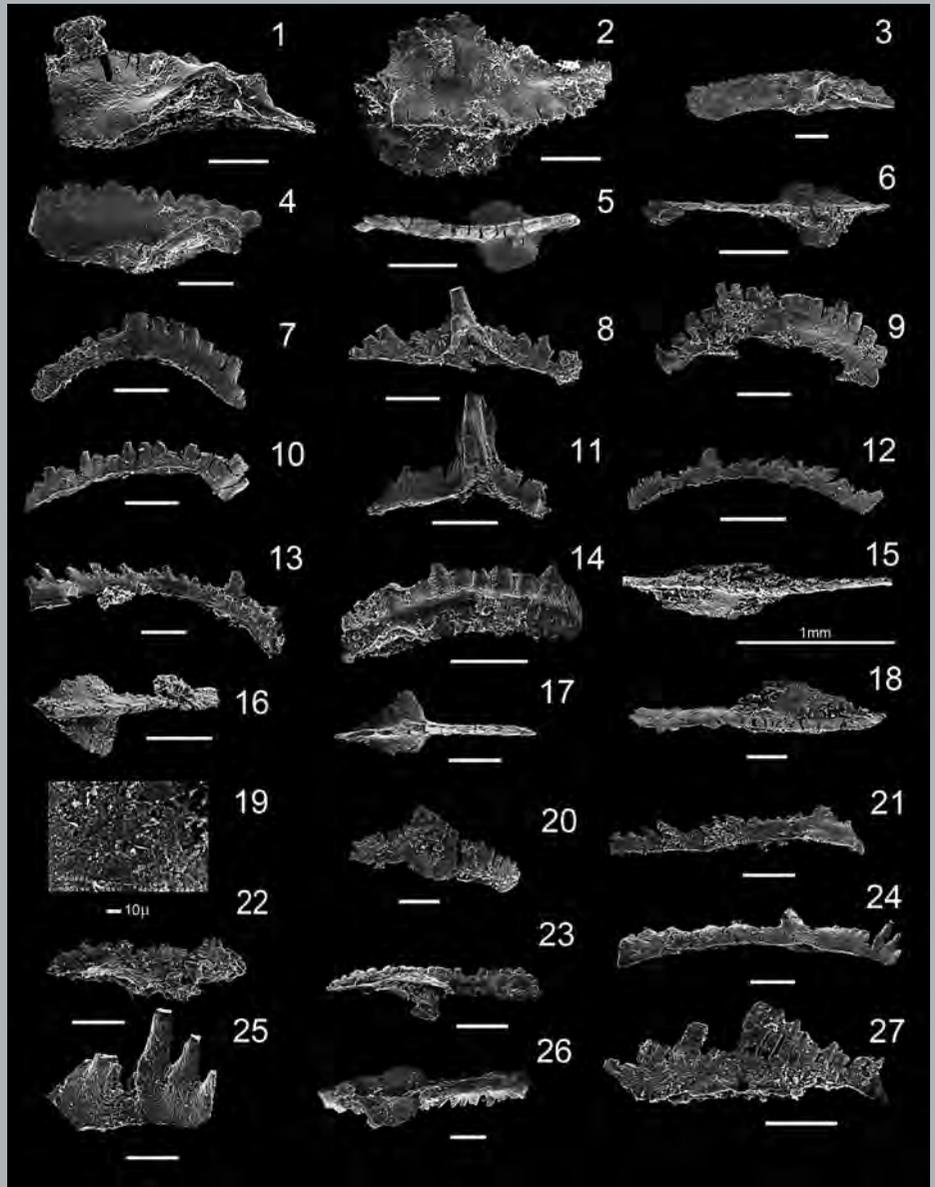
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Cover: Some Silurian conodonts from the Sierra del Courel (province of Lugo, Spain). Microphotographies by Roselis W. Salazar Ramírez, Graciela N. Sarmiento y Juan Carlos Gutiérrez-Marco. Scale bars = 200 µm.

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Palynological dating and correlation of Early Mississippian (Tournaisian) diamictite sections, Parnaíba Basin, northeastern Brazil

Geoffrey Playford¹, Leonardo Borghi², Gustavo Lobato³, and José Henrique G. Melo⁴

¹ School of Earth Sciences, The University of Queensland, St. Lucia Campus, Brisbane, Queensland 4072, Australia.
g.playford@uq.edu.au

² Instituto de Geociências, Universidade Federal do Rio de Janeiro, Cidade Universitária, Ilha do Fundão, 21941-916 Rio de Janeiro, RJ, Brazil. lborghi@geologia.ufrj.br

³ Petrobras/E&P-Exp/labmeq/Intp, Avenida República do Chile, 330, 20031-170 Rio de Janeiro, RJ, Brazil.
gustavolobato@petrobras.com.br

⁴ Petrobras/Cenpes/Pdgeo/Bpa, Avenida Horácio Macedo, 950, Cidade Universitária, Ilha do Fundão, 21941-915 Rio de Janeiro, RJ, Brazil. jhmelo@petrobras.com.br

RESUMEN

En el subsuelo de la cuenca de Parnaíba del oeste de Brasil, las diamictitas y las lutitas asociadas que integran la Formación Longá, proporcionan un testimonio sedimentológico claro de su origen glaciogénico. La identificación taxonómica de las abundantes y diversas miosporas en las muestras del sondeo han permitido la atribución de la palinoflora a la Zona de Intervalo de miosporas *Spelaeotriletes pretiosus–Colatisporites decorus* [PD] del Misisípico Inferior (finales del Tournaisiense medio – principios del Tournaisiense superior), que se estableció en la Cuenca del Amazonas y es equivalente a la Zona PC del Oeste de Europa. La aparición de palinomorfos reelaborados del Devónico Medio y Superior indica procesos de erosión glacial. Este evento glacial tournaisiense que está representado también en las cuencas del Amazonas y Solimões en el oeste, es cronológicamente intermedio entre las edades de hielo del Devónico final (Estruniense) y del principio del Viseense superior (del Holkeriense superior al Asbense), los cuales están documentados en el oeste de Gondwana. Su señal en paleolatitudes bajas (p.e. Norteamérica, Europa occidental y oriental, Asia Central, Siberia y China) consistió en un enfriamiento climático acompañado con regresiones forzadas que produjeron extensas superficies de exposición subaérea y anomalías de los isótopos estables en todo el mundo. El descubrimiento de diamictitas tournaisienses en el noreste de Brasil apoya las evidencias geoquímicas y estratigráficas de una glaciación durante el límite entre el Tournaisiense medio y el superior.

Palabras clave: Palinología, estratigrafía, diamictitas, glaciación, Misisípico (Tournaisiense), Brasil, Cuenca de Parnaíba.

ABSTRACT

In the subsurface of the Parnaíba Basin of northeastern Brazil, diamictites and associated shaly sediments comprising the upper Longá Formation provide clear sedimentological testimony of their glaciogenic origin. Taxonomic identification of abundant and diverse plant microfossils (miospores) in the core samples enables attribution of the palynoflora to the *Spelaeotriletes pretiosus–Colatisporites decorus* [PD] miospore Interval Zone of Early Mississippian (late middle to early late Tournaisian) age, which was established in the Amazonas Basin and is equivalent to the PC Zone of Western Europe. Recycled palynomorphs point to glacial-erosional contributions from Middle and Late Devonian provenances. This Tournaisian glacial event, represented also in the Amazonas and Solimões basins to the west, is chronologically intermediate between the latest Devonian (Strunian) and early late Visean (late Holkerian to Asbian) ice

ages (both of which are documented in Western Gondwana). Its signature in lower palaeolatitudes (e.g., North America, Western and Eastern Europe, Middle East, Siberia, and China) consists of climatic cooling coupled with forced regressions, causing widespread sub-aerial exposure surfaces and worldwide stable isotope anomalies. The discovery of Tournaisian diamictites in northeastern Brazil lends support to geochemical and sequence-stratigraphic evidence for a glaciation at the middle/late Tournaisian boundary.

Keywords: Palynology, stratigraphy, diamictites, glaciation, Mississippian (Tournaisian), Brazil, Parnaíba Basin.

1. INTRODUCTION

Glacial events of latest Devonian and Mississippian age in South America have attracted considerable attention, and their effects and lithological records have been documented from there and elsewhere in an expanding literature.

Of these events, the end-Famennian (Strunian) glacial episode is undoubtedly the best documented, and also the most widespread geographically. It has been recognized on the basis of varied lithological evidence, which includes tillites and polymictic diamictites with angular, often polished/striated clasts; dropstones within varve-like laminites; glacially deformed strata; and (in outcrop) striated pavements and erratic boulders. This end-Devonian event is particularly well characterized in Palaeozoic basins of northern and northeastern Brazil, viz., the Solimões, Amazonas, Parnaíba, and Jatobá basins (Kegel, 1953; Malzahn, 1957; Carozzi *et al.*, 1975; Caputo, 1984, 1985; Streel *et al.*, 2000, 2001, 2011; Melo & Loboziak, 2003; Caputo *et al.*, 2008; Caputo & Ponciano, 2010). Moreover, it is possibly represented in the Paraná Basin to the south (Ortigueira diamictite: Loboziak, Melo, Steemans & Barrilari, 1995; Melo & Loboziak, 2003; Milani *et al.*, 2007; Caputo *et al.*, 2008). Outside Brazil, glaciogenic (or glacially-influenced) deposits of Strunian age occur in Andean/Subandean regions of Peru, Bolivia, and Argentina (Díaz-Martínez & Isaacson, 1994; Díaz-Martínez *et al.*, 1999; Isaacson *et al.*, 1999; Carlotto *et al.*, 2006; Streel, 2006); in eastern North America (Suter, 1991; Cecil *et al.*, 2002, 2004; Brezinski *et al.*, 2008, 2010; Etensohn *et al.*, 2009; Clayton *et al.*, 2010); and in some African countries, notably in Libya, Niger, Ghana, and South Africa (Bär & Riegel, 1974; Hambrey & Kluyver, 1981; Caputo, 1984, 1985; Streel *et al.*, 2000).

Diamictites and other glacially influenced strata of Tournaisian age have previously been reported from the Solimões and Amazonas basins of northern Brazil (Ca-

puto, 1984; Loboziak, Melo, Quadros *et al.*, 1995; Caputo *et al.*, 2008), and less surely from other South American basins (Streel, 2006; Caputo *et al.*, 2008; J.H.G. Melo and collaborators, unpublished data) and South Africa (Streel & Theron, 1999).

Early late Visean (late Holkerian to Asbian) glaciogenic sediments are known from northern and northeastern Brazil (Solimões and Parnaíba basins: Caputo *et al.*, 2008), the Altiplano and Eastern Cordilleran regions of Peru and Bolivia (Caputo *et al.*, 2008), and northwestern Argentina (Río Blanco Basin: Gulbranson *et al.*, 2010; Perez Loinaze *et al.*, 2010). Diamictite dykes of the Falkland Islands/Islas Malvinas (offshore Argentina) – palynologically dated as “late Visean-early Namurian” by Hyam *et al.* (1997) – contain saccate pollen associated with late Visean and younger trilete spores, and are thus assignable to the early Serpukhovian (Melo & Playford, in press).

Objectives of the present study are to detail the sedimentological characteristics of Tournaisian diamictites and associated sediments from cored drillholes penetrating part of the Mississippian succession of the Parnaíba Basin (northeastern Brazil); and to precisely date and correlate the strata on the basis of palynofloral evidence with reference to palynozonations established for the region, in particular the adjacent Amazonas Basin.

For details and additional literature pertaining to the lithostratigraphy of Brazilian Palaeozoic basins discussed in this paper, the reader is referred to Wanderley Filho *et al.* (2007: Solimões Basin), Cunha *et al.* (2007: Amazonas Basin), Milani *et al.* (2007: Paraná Basin), and Costa *et al.* (2007: Jatobá Basin). Caputo (1984) and Caputo *et al.* (2008) have provided additional information on the Palaeozoic stratigraphy and glacial records in those basins. The Devonian–Mississippian miospore zonation of Brazilian basins utilized herein is that of Melo & Loboziak (2003) and Melo & Playford (in press).

2. GEOLOGY

The Parnaíba Basin is a predominantly Palaeozoic (Ordovician?, Silurian–Permian) sag basin situated in northeastern Brazil, ca. 3,500 m deep at its depocentre, with an area of ca. 600,000 km². The Longá Formation belongs to the Canindé Group, which is considered a 2nd-order sequence of late Early Devonian–Middle Mississippian age (Vaz *et al.*, 2007).

The lower and upper contacts of the Longá Formation, with the underlying Cabeças Formation and the overlying Poti Formation, are considered conformable and unconformable respectively (Caputo, 1984). This author subdivided the formation into three informal lithostratigraphic units: (a) lower greenish-greyish shales with interbedded conglomerates and conglomeratic sandstones; (b) an intermediate unit composed mainly of sandstones and conglomeratic sandstones; and (c) an upper unit, comprising black shales with rare sandstone interbeds and subordinate dropstones. The Longá palaeoenvironment is generally regarded as a transgressive, storm-dominated, shallow sea (Caputo, 1984; Góes & Feijó, 1994; Vaz *et al.*, 2007). The lower part of the formation contains an impoverished benthic marine invertebrate fauna of latest Devonian age and Malvinokaffric affinity (Santos & Carvalho, 2009), including bivalves, brachiopods, rare small trilobites, and trace fossils (Fernandes *et al.*, 2002).

The Cabeças Formation, composed mainly of sandstones and subordinate siltstones of Late Devonian (Famennian) age, was deposited in storm- to tide-dominated, shallow marine, deltaic, and fluvial environmental settings (Góes & Feijó, 1994; Young & Borghi, 2006; Vaz *et al.*, 2007; Santos & Carvalho, 2009). Glaciogenic rhythmites, mudstones, slumped sandstones, diamictites, and tillites, locally associated with striated pavements and erratic boulders, occur in the upper part of the formation (dated as Strunian with miospores; Grahn *et al.*, 2001; Streel *et al.*, 2001; Melo & Loboziak, 2003; Grahn *et al.*, 2006; Caputo *et al.*, 2008). Moderately fossiliferous sandstones with shaly interbeds, formerly attributed to the lowest part of the formation and constituting the so-called “Passagem Member” of Kegel (1953), are limited to the eastern outcrop belt of the basin (Piauí State). Fossil assemblages include bivalves, gastropods, tentaculitids, brachiopods, crinoids, trilobites, land plant debris, and trace fossils, to-

gether with palynomorphs of early to early middle Givetian age (Kegel, 1953; Grahn *et al.*, 2008; Breuer & Grahn, 2011a; Ponciano *et al.*, 2012). According to recent interpretations (Ponciano & Della Favera, 2009; Ponciano *et al.*, 2012), the Passagem Member comprises distal (fine-grained) mouth-bar deposits interbedded with delta-front, hummocky cross-stratified, graded sandstone lobes, collectively representative of a flood-dominated, fluvio-deltaic system. The member is unconformably overlain by Famennian sandstones of the Cabeças Formation *sensu stricto*, and grades westerly (basinward) into Givetian marine strata of the Pimenteira Formation (Beurlen, 1965; Beurlen & Mabesoone, 1969; Grahn *et al.*, 2008; Breuer & Grahn, 2011b; Ponciano *et al.*, 2012), the total age span of which is late Eifelian through early Famennian in the subsurface (Grahn & Melo, 2005; Grahn *et al.*, 2006).

The Poti Formation, succeeding the Longá Formation, consists mainly of sandstones with subordinate mudstones, in part fossiliferous, containing plant remains, bivalves, and miospores of early late Visean age (Melo & Loboziak, 2000, 2003; Iannuzzi & Pfefferkorn, 2002; Santos & Carvalho, 2009). The formation accumulated in fluvial to marine environmental settings (Santos & Carvalho, 2009; Góes & Feijó, 1994; Vaz *et al.*, 2007).

In his study of the upper Longá Formation (i.e., the interval analyzed herein) and the lower Poti Formation, Lobato (2010) described facies, depositional systems, and forced regressive stratigraphic surfaces (unconformities). These, in turn, resulted in the designation of two 3rd-order stratigraphic depositional sequences, divisible into sequences of higher (5th/6th) orders. Lobato (2010) further suggested that the boundary between the Longá and Poti formations represents a subaerial erosional surface separating two 3rd-order sequences. Accordingly, palynological studies by Grahn *et al.* (2001), Melo & Loboziak (2000), and the present authors point to a hiatus of considerable magnitude (i.e., ca. 10–11 Ma, after the timescale of Ogg *et al.*, 2008), separating late Holkerian–Asbian from early late Tournaisian (= early Ivorian, early Osagean) strata of the Poti and upper Longá formations, respectively. Additional palynological evidence (Grahn *et al.*, 2001; S. Loboziak and J.H.G. Melo, unpublished data; G. Playford and J.H.G. Melo, ongoing studies) permits the recognition of at least two 3rd-order sequences within the Longá Formation; viz., an end-Famennian (latest Strunian) sequence (which

also encompasses the underlying, uppermost Cabeças strata) and a middle to early late Tournaisian sequence. These are separated by an intra-formational hiatus of earliest Tournaisian or longer duration, depending on whether pre-PD strata are locally present or absent in the Longá Formation due to Tournaisian coastal onlap at the basin's margin (cf. Figure 4). For instance, in Petrobras well 1-TM-1-MA (situated in the southern part of the basin; see Andrade & Daemon, 1974; Grahn *et al.*, 2001), the Longá Formation is entirely of PD age and lies disconformably on latest Famennian (LVa-age) sandstones and siltstones of the Cabeças Formation (Loboziak, Melo, Quadros *et al.*, 1995; S. Loboziak and collaborators, unpublished data).

Evidence of glacial conditions within the Canindé Group was first observed by Malzahn (1957), and later corroborated by Carozzi *et al.* (1975) and Caputo (1984, 1985). Caputo *et al.* (2008) recognized three glacial episodes within the Upper Devonian–Mississippian of South America. In the Parnaíba Basin, the oldest and youngest of these (latest Famennian and early late Viséan) have been reported from the Cabeças and Poti formations, respectively (Caputo, 1984; StreeL *et al.*, 2000, 2001, 2011; Caputo *et al.*, 2008). A further (stratigraphically intermediate) ice age, at the middle/late Tournaisian transition, is now confirmed herein, within the upper Longá Formation. Caputo *et al.* (2008, p.167) pointed to the difficulty in palynologically dating the Tournaisian glaciation in South America, because of massive reworking and resedimentation of older (latest Famennian), richly palyniferous glaciogenic sediments. Moreover, two of the present authors (L.B. and G.L.) note lithostratigraphic uncertainties in the long-distance correlation of Parnaíba Basin diamictites identified in outcrops and subsurface (cores), based on 'event-beds' (or 'marker beds'). Accordingly, they maintain, as had Lobato (2010), that some diamictites previously attributed to the Poti Formation could actually belong to the Longá Formation.

3. MATERIAL AND METHODS

The samples investigated herein derive from three shallow core-drills (diameter = 4 cm) commissioned by the Departamento Nacional da Produção Mineral (DNPM) in the

Parnaíba Basin (Figure 1) during the early 1970s, in connection with Project "Carvão da Bacia do Parnaíba" (Leite *et al.*, 1975; Lima & Leite, 1978). Well codes and geographic locations are as follows: 1UN-04-PI (04°09'51.0"S, 42°56'45.0"W), near Miguel Alves town, Miguel Alves Municipality, Piauí State (PI); 1UN-24-PI [*sic*] (03°54'37.0"S, 42°56'47.0"W), near Buriti town, Buriti Municipality, Maranhão State (MA); and 1UN-27-PI [*sic*] (09°29'28.0"S, 47°14'21.0"W), Pilões village, Rio Sono Municipality, Tocantins State (TO). Note that, although the three wells include the suffix "PI" (indicating Piauí State) in their original codes (Leite *et al.*, 1975), only 1UN-04-PI was actually drilled in that state. Henceforth in this paper, the codes for the three wells (1UN-04-PI, 1UN-24-PI, and 1UN-27-PI) are abbreviated thus: UN-04, UN-24, and UN-27, respectively.

Lobato (2010) described cores from two wells (UN-04 and UN-24) in sedimentological detail (1:40 scale), with particular emphasis on their main facies attributes (lithology, granulometry/selection, sedimentary structures, trace fossils, and colour). These facies analyses served as a prime basis for interpreting the depositional system in a sequence-stratigraphic context. Cores from the third well (UN-27) have subsequently been described by G. Lobato (not having been included in his initial study). Twenty-six samples, including shales and diamictites, were collected for the present palynological analyses from the cores of UN-04, UN-24, and UN-27 within the Longá Formation (Figure 1). Of that total, (a) eight samples belong to the diamictite facies, (b) 14 samples to the shaly facies, and (c) the remaining four samples derive from muddy interbeds within argillaceous sandy intervals of UN-24 (at 259.17 m, 283.80 m, 301.55 m, and 335.40 m).

All samples were processed in the Petrobras/Cenpes/Bpa laboratories using conventional physico-chemical techniques for extraction and concentration of the palynomorphs (e.g., Playford & Dino, 2000, pp. 10, 12). Yields were generally satisfactory, but quality of palynomorph preservation proved fair to poor in many samples, due to the inimical effects of thermal baking and corrosion of thin exines resulting from regional Mesozoic diabase intrusions. Such effects tend to obscure or distort taxonomically significant sculptural/structural features; nevertheless, many species could be reliably identified, including certain defining indices of Melo & Loboziak's

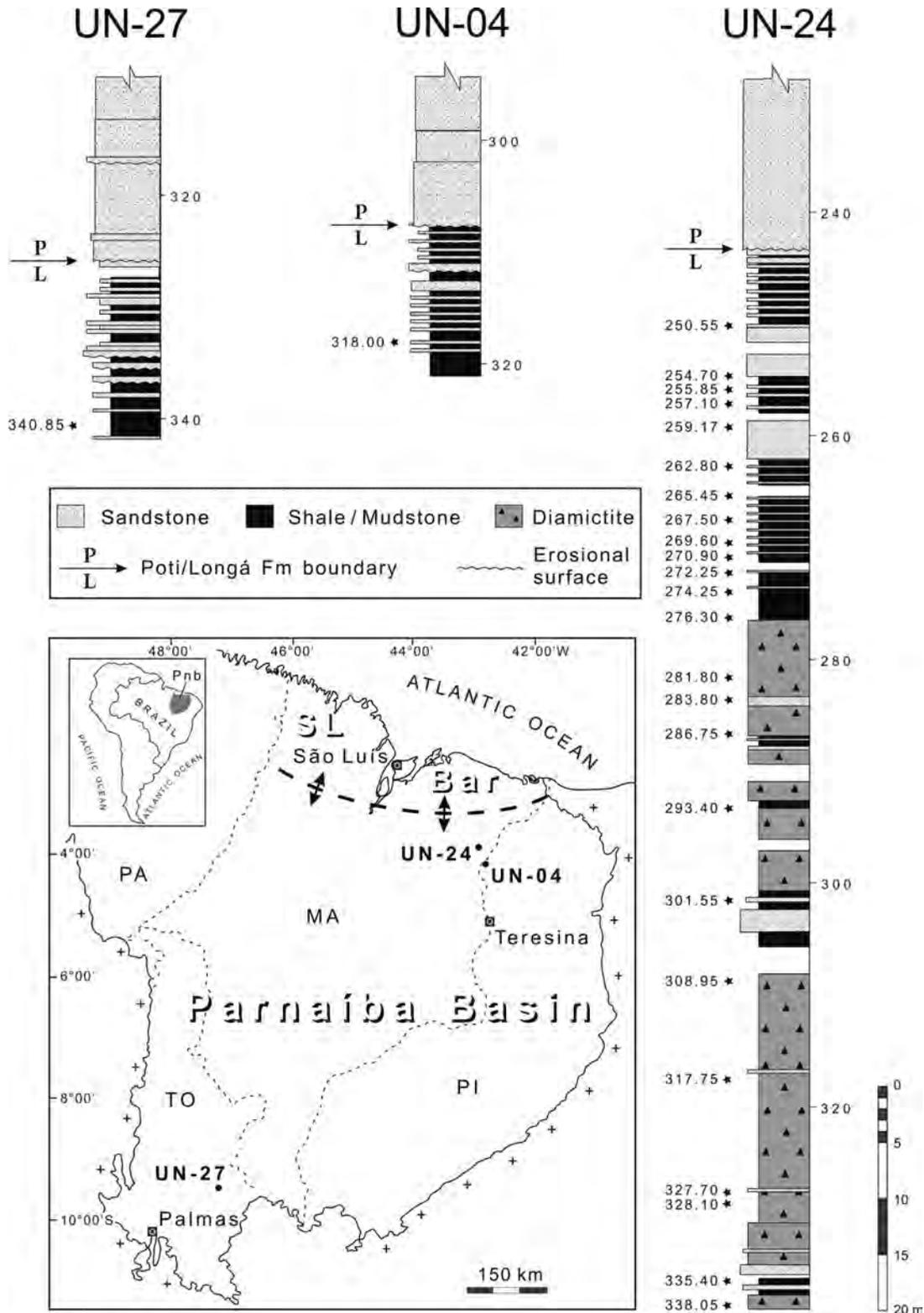


Figure 1. Location map of the Parnaíba Basin showing the three core-drills studied (UN-04, UN-24, and UN-27) with their lithological logs and sampling depths. Note: (a) Grey area of inset map indicates Parnaíba Basin (Pnb). (b) Thick dashed line with transverse double arrows represents tectonic hinge (Ferrer–Urbano Santos Arch) separating Parnaíba Basin, to the south, from coastal São Luís (SL) and Barreirinhas (Bar) basins, to the north. (c) Thin dashed lines indicate Brazilian state boundaries. (d) States designated thus: MA, Maranhão; PA, Pará; PI, Piauí; TO, Tocantins. (e) Sharp boundary, in the three core-drills, between the lower interval (shales, diamictites, and thin sandstone interbeds), belonging to the Longá Formation, and the succeeding, thicker sandstone package of the Poti Formation. (f) Blank intervals in the lithological columns indicating non-recovery of cores.

(2003) miospore zones of the Amazonas Basin. Strew slides were examined at Petrobras/Cenpes/Bpa by means of a Zeiss Imager.A1 binocular microscope equipped with a Zeiss AxioCam MRc/AxioVision 4.6.3 SP1 facility for transmitted-light photomicrography through AxioVision Rel. 4.8 software.

Slides containing specimens illustrated herein (Plate 1, Figs.1-12; Plate 2, Figs. 1-12; Plate 3, Figs. 1-16) are housed at Biostratigraphy and Palaeoecology Management of Petróleo Brasileiro S.A. (Petrobras/Cenpes/Pdgeo/Bpa), Ilha do Fundão, Rio de Janeiro, RJ, Brazil. Curatorial data are incorporated in the plate explanations.

4. SEDIMENTOLOGY

Lobato (2010) characterized sedimentologically the upper interval of the Longá Formation and the succeeding lower part of the Poti Formation, based on 180 m of cores from two wells (UN-04 and UN-24). Altogether, 13 sedimentary facies (10 litho- and three ichnofacies) have been identified and grouped into nine facies associations (Lobato & Borghi, 2007; Lobato, 2010). These enabled recognition of three major depositional systems; viz., (a) glaciomarine, (b) wave-dominated shallow marine, and (c) wave-dominated fluvio-deltaic (delta front and prodelta). Of particular interest for the present study are the diamictite and shaly facies of the upper Longá Formation (Figure 2E).

Diamictite facies are characterized by an overwhelming dominance of extremely poorly sorted, clayey-sandy matrix, dark to medium grey (N4-N5) or greyish red (5R-4/2), in which numerous shale pebbles float (Figure 2A, C). Petrographically, they are classifiable as pebbly mudstones (“muddy diamictites”, Figure 3A, B) or sandstones (“sandy diamictites”, Figure 3C). These diamictite facies show (a) flow alignment of the clasts, produced by shearing gravity flow (Figure 3A); (b) rounded and angular quartz grains (Figure 3A, C), the latter produced by crushing; (c) till pellets (clasts), actually small diamictite clots coated with clay (cf. Ovenshine, 1970, p. 893), produced by the reworking of preexisting diamictites (Figure 3B); and (d) highly deformed strata, characterized by steeply dipping

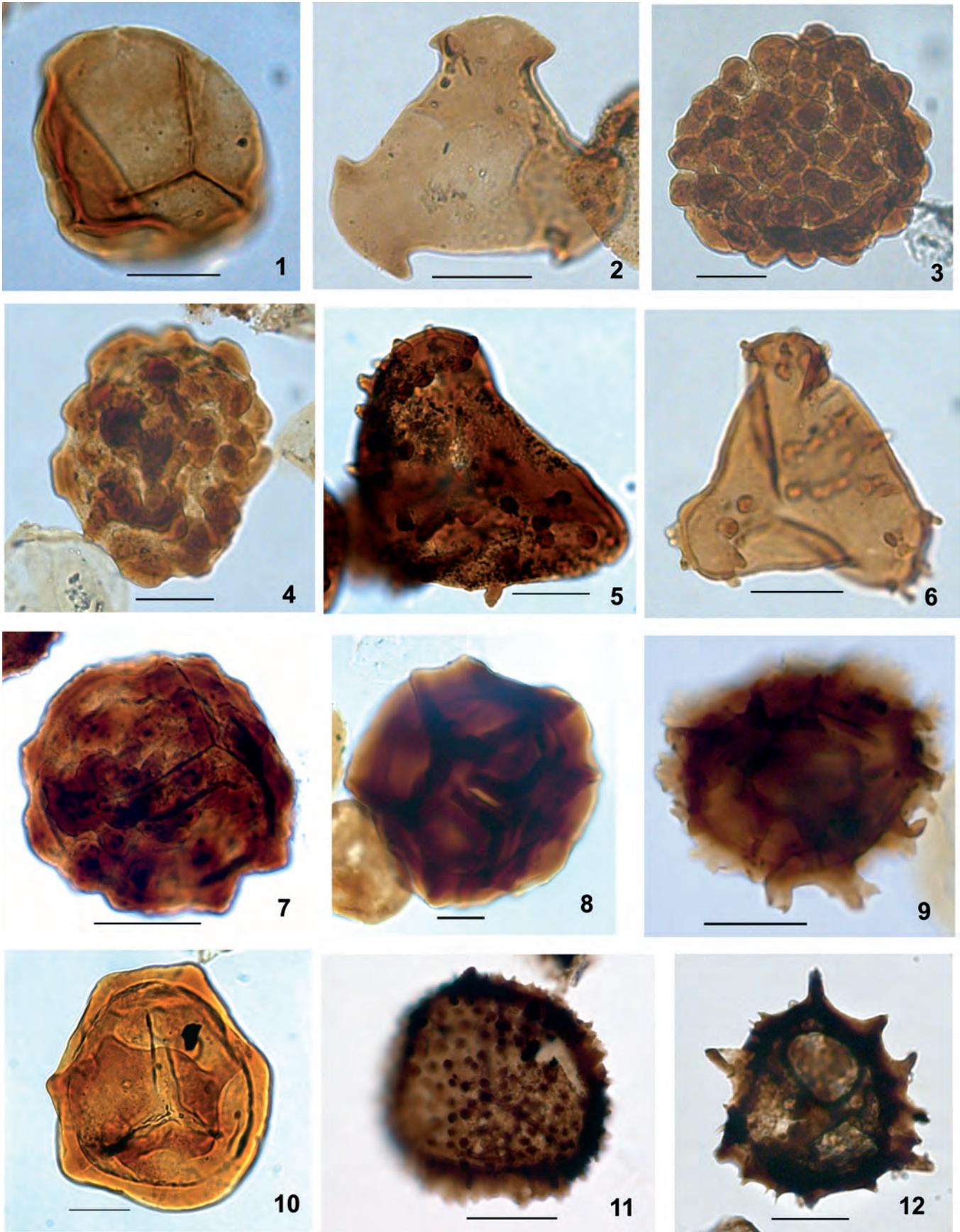
(Figure 2B) and overturned (“swiss-roll”) beds (Figure 2C), produced by slumping or sliding. These features of the diamictites are interpreted as the result of rain-out, re-sedimentation, and bottom dragging of unconsolidated sediments by icebergs or glaciers. They characterize the lower part of the studied interval (Figure 1), interpreted as part of a glaciomarine depositional system (Lobato, 2010).

Shaly facies comprise wavy- to linsen-bedded shales and sandstones. The shales are pale to dark grey (N8-N3), displaying synaeresis cracks (Figure 2B). These heterolithic shaly facies characterize the upper part of the studied interval. In association with subordinate sandstone beds (Figure 1), they can be interpreted as a glacially-influenced (?) prodelta/delta within a wave-dominated deltaic depositional system (Lobato, 2010).

5. PALYNOFLORAL COMPOSITION AND ZONAL ATTRIBUTION

All 26 samples examined yielded palynomorphs – consisting of miospores and, commonly much subordinately, of organic-walled microphytoplankton (acritarchs and prasinophyte phycmata) – in varying concentrations and states of preservation. The best-preserved and most palyniferous sample is that from UN-27 at 340.85 m and, to a somewhat lesser degree, the sample from UN-24 at 317.75 m. Although the palynomorphs retrieved from the other 24 samples are mostly in lower concentrations and are more or less carbonized, with dark brown to black

Plate 1. Selected miospores from the Longá Formation from DNPM core-drills UN-27 (Figs. 1-10) and UN-24 (Figs. 11, 12). Information provided for each figured specimen comprises sample code and depth, preparation/slide no., and England Finder coordinates. Scale bars 20 µm in length. 1, *Calamospora* sp., P3-340.85 m, 200607773, S40/2. 2, *Waltzisporea lanzonii* Daemon, 1974, P3-340.85 m, 200607773, P47/1. 3, *Verrucosporites nitidus* Playford, 1964, P3-340.85 m, 200607773, N42/1. 4, 7, *Convolutispora stigmoidea* Bharadwaj & Venkatachala, 1962, P3-340.85 m, 200607773, K40/2 (4) and T65 (7). 5, 6, *Neoraistrickia loganensis* (Winslow, 1962) emend. Coleman & Clayton, 1987, P3-340.85 m, 200607773, G43 (5) and F59/2 (6). 8, *Corbulispora cancellata* (Waltz in Luber & Waltz, 1938) Playford & Melo, 2012, P3-340.85 m, 200607773, P58/3. 9, *Cordylosporites absimilis* (Daemon, 1974) emend. Playford & Melo, 2012, P3-340.85 m, 200607773, G41/2. 10, *Knoxisporites hederatus* (Ishchenko, 1956) Playford, 1963, P3-340.85 m, 200607773, Q65. 11, *Cymbosporites loboziakii* Melo & Playford, 2012, P19-317.75 m, CB200909881-1, Q44/4. 12, *Densosporites infacetus* Daemon, 1974, P19-317.75 m, CB200909881-1, D51/1.



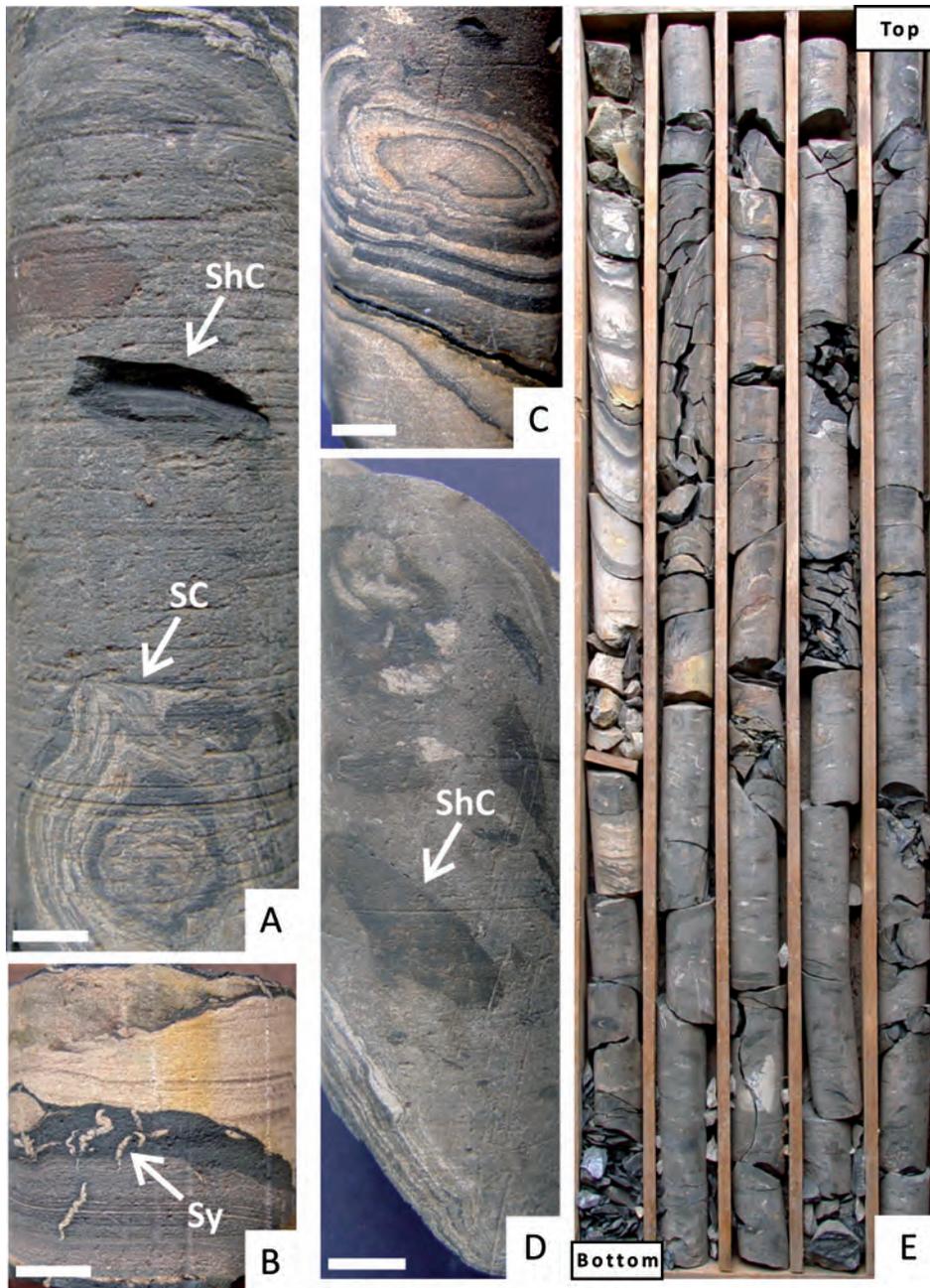
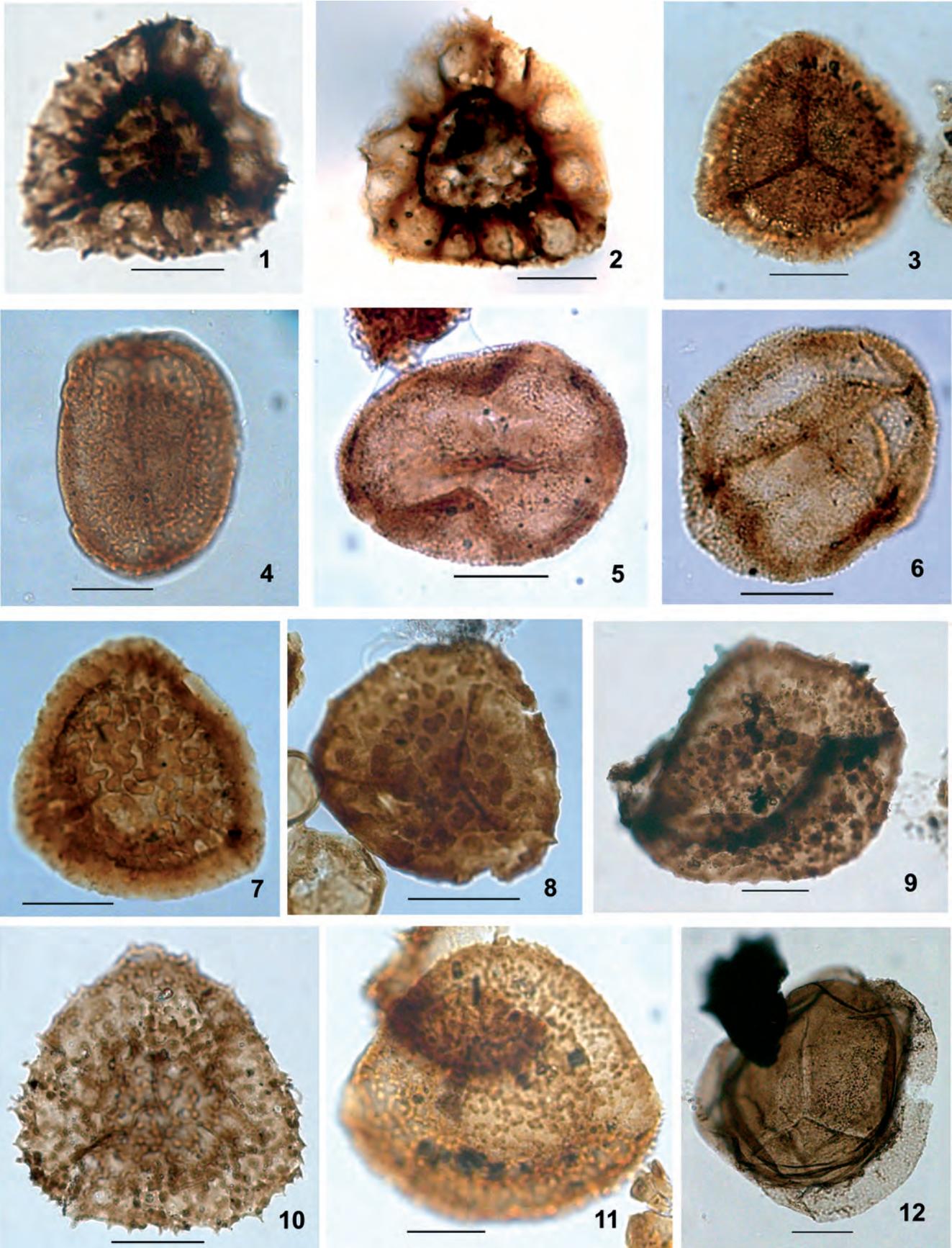


Figure 2. Megascopic aspect of the upper Longá Formation facies studied by Lobato (2010) in vertical core-drill UN-24 (A–D, scale bar = 1 cm; E, core diameter = 4 cm, box length = 1 m). A, Massive, matrix-supported diamictite with shale clasts (ShC) and a sheath-folded (= “condom-folded”) laminated siltstone/sandstone clast (SC); depth 292.00 m. B, Synaeresis cracks (Sy) in cross-section within black mudstone, overlain by very fine sandstone with wave cross bedding (wavy bedding), showing a deeply incised erosional lower bed contact (gutter cast); depth 257.35 m. C, “Swiss-roll”-folded laminated siltstone/sandstone clast within diamictite; depth 285.00 m. D, Sheared, matrix-supported diamictite with inclined (slumped) shale clasts (ShC), as evidenced by high-angle clast alignment and matrix fissility; depth 285.10 m. E, Diamictite rich in shale clasts, and slumped (folded/contorted) bedding at top left; core box #49, depth interval 297.70–302.60 m.

Plate 2. Selected miospores from the Longá Formation from DNPM core-drills UN-24 (Figures 1, 2, 9–12) and UN-27 (Figures 3–8). Information provided for each figured specimen comprises sample code and depth, preparation/slide no., and England Finder coordinates. Scale bars 20 μ m in length. 1, *Indotriradites multivolus* (Daemon, 1974) emend. Melo & Playford, 2012, 317.75 m, CB200909881-1, N45/2. 2, *Radiizonates arcuatus* Loboziak, Playford & Melo, 2000, P16-286.75 m, 200607772, V47. 3, *Vallatisporites vallatus* Hacquebard, 1957, P3-340.85 m, 200607773, E57/3. 4, *Ara-trisporites saharaensis* Loboziak, Clayton & Owens, 1986, P3-340.85 m, 200607773, J68/1. 5, *Colatisporites decorus* (Bharadwaj & Venkatachala, 1962) Williams in Neves, Gueinn, Clayton, Ioannides, Neville & Kruszezwska, 1973, P3-340.85 m, 200607773, M68/4. 6, *Colatisporites denticulatus* Neville in Neves, Gueinn, Clayton, Ioannides, Neville & Kruszezwska, 1973, P3-340.85 m, 200607773, K69/1. 7, *Vallatisporites splendens* Staplin & Jansonius, 1964, P3-340.85 m, 200607773, J64/2. 8, 9, *Spelaeotriletes pretiosus* (Playford, 1964) Utting, 1987, h, P3-340.85 m, 200607773, G49/4. i, R10-283.80 m, CB20110212219-1, S49. 10, 11, *Spelaeotriletes balteatus* (Playford, 1963) Higgs, 1996. 10, P16-286.75 m, CB20110212261-1, O52/2. 11, P3-340.85 m, 200607773, O38. 12, *Auroraspora corporiga* Higgs, Clayton & Keegan, 1988, P15, 281.80 m, CB20110212260-1, Q68.



colouration, they nevertheless proved mainly identifiable at specific level. Representative palynomorphs are illustrated by transmitted-light photomicrographs comprising Plates 1-3.

The considerable diversity of *in situ* miospores identified during this study is documented in Table 1, as is the presence of miospores and organic-walled microphytoplankton reworked from subjacent Devonian strata (Table 2). From the known distribution of the autochthonous species – as listed in the former table's right-hand column with prime reference to Melo & Loboziak's (2003) palynozonal scheme – it is clear that the Longá samples are assignable within the Mississippian interval spanned successively by the AL (*Radiizonates arcuatus*–*Waltzispora lanzonii*), BL (*Spelaeotriletes balteatus*–*Neoraistrickia loganensis*), and PD (*Spelaeotriletes pretiosus*–*Colatisporites decorus*) miospore interval zones. More precise zonal attribution, hence precise dating of the Longá assemblage, is facilitated by the joint occurrence, in all or many of the samples, of the following species: *Waltzispora lanzonii* Daemon, 1974; *Neoraistrickia loganensis* (Winslow, 1962) emend. Coleman & Clayton, 1987; *Raistrickia clavata* Hacquebard, 1957 emend. Playford, 1964; *Colatisporites decorus* (Bharadwaj & Venkatachala, 1962) Williams in Neves *et al.*, 1973; *C. denticulatus* Neville in Neves *et al.*, 1973; *Indotriradites multivolus* (Daemon, 1974) emend. Melo & Playford, 2012 [syn. *I. mitratus* (Higgs, 1975) Higgs, 1996]; *Spelaeotriletes balteatus* (Playford, 1963) Higgs, 1996; and *S. pretiosus* (Playford, 1964) Utting, 1987. Collectively and with reference to Table 1, the association of these species attests to correlation with the PD Interval Zone, which was characterized and dated by Melo & Loboziak (2003, pp. 191-192) as late middle to early late Tournaisian. On a putatively contra-indicative note, however, it should be pointed out that (a) a single specimen of *Neoraistrickia cymosa* Higgs, Clayton & Keegan, 1988 was identified (in one sample only: UN-24, 250.55 m); and (b) several specimens of *Auroraspora corporiga* Higgs, Clayton & Keegan, 1988 were found in four samples from the same well (at 255.85 m, 269.60 m, 283.80 m, and 301.55 m). These species have not previously been encountered in northern Brazilian strata younger than the BL Interval Zone (*N. cymosa*) or the AL Interval Zone (*A. corporiga*): S. Loboziak, J.H.G. Melo, and G. Playford (unpublished data). However, in Western Europe the range of both species reportedly extends into

age-equivalents of the PD Interval Zone; i.e., PC Zone of Ireland (Higgs *et al.*, 1988) and Belgium (Higgs, 1996). Hence the occurrence of *N. cymosa* and *A. corporiga* in the assemblage reported herein is not signified as conflicting with the PD correlation.

As noted above, the majority of the organic-walled microphytoplankton comprise recycled Devonian elements. However, all slides additionally contain some small, simple, and better preserved veryhachiids and morphologically simple acanthomorphic vesicles, which are not age-diagnostic and are believed to be autochthonous, hence indicative of marine depositional conditions for the Tournaisian Longá strata.

6. AGE, CORRELATION, AND GLOBAL SIGNIFICANCE OF THE LONGÁ FORMATION DIAMICTITES

As stated previously (see Introduction), the occurrence of end-Famennian (Strunian) and early late Visean (late Holkerian to Asbian) glaciogenic sediments in Palaeozoic basins of northern Brazil and other parts of the world is well documented in the literature (see Caputo *et al.*, 2008; Clayton *et al.*, 2010; Gulbranson *et al.*, 2010; Perez Loinaze *et al.*, 2010; Streele *et al.*, 2000, 2001, 2011; and references therein). Some Strunian, Tournaisian, and Visean incised-valley sandstones reported from North Africa (Morocco and Algeria: Lubeseder *et al.*, 2010) could reflect marine lowstands partly related to Gondwanan glaciations to the south. Our new discovery of late middle to early late Tournaisian, glacio-marine diamictites in the Parnaíba Basin (Lobato & Borghi, 2007; Lobato, 2010; and present paper) adds to previous records of glacially derived Tournaisian strata in the Solimões and Amazonas basins of northern Brazil, and possibly other South American basins (Loboziak, Melo, Quadros *et al.*, 1995; Caputo, 1984; Caputo *et al.*, 2008; Melo & Playford, in press; J.H.G. Melo and collaborators, unpublished data). Collectively, these records provide unmistakable evidence of an important Tournaisian ice age in Western Gondwana.

The effects of the Early Mississippian glaciation in extra-Gondwanan regions, mainly in the U.S.A. but also in Eu-

Miospore species ↓	Samples →	1	2	3	4	5	6	7	8	Zonal occurrence
<i>Anapiculatisporites hystricosus</i> Playford, 1964			✓							PD
<i>Anapiculatisporites redactus</i> Playford, 1978								✓		PD
<i>Anapiculatisporites semicuspidatus</i> Playford & Melo, 2012			✓	✓			✓			AL–PD
<i>Apiculatasporites quadrosii</i> Playford & Melo, 2012				✓	✓	✓	✓	✓	✓	AL–PD
<i>Apiculiretusispora multiseta</i> (Luber in Luber & Waltz, 1938) Butterworth & Spinner, 1967			✓							AL–Mag
<i>Apiculiretusispora</i> sp.				✓						PD
<i>Aratrisporites saharaensis</i> Loboziak, Clayton & Owens, 1986		✓	✓	✓		✓	✓	✓		Rle–Mag
<i>Auroraspora corporiga</i> Higgs, Clayton & Keegan, 1988					✓					LVa, AL, PD
<i>Calamospora</i> spp.			✓		✓			✓	✓	AL–Mag
<i>Colatisporites decorus</i> (Bharadwaj & Venkatachala, 1962) Williams in Neves <i>et al.</i> , 1973		✓	✓		✓	✓		✓		PD, Mag
<i>Colatisporites denticulatus</i> Neville in Neves <i>et al.</i> , 1973		✓	✓	✓	✓	✓	✓	✓		PD, Mag
<i>Convolutispora major</i> (Kedo, 1963) Turnau, 1978			✓							Rle–PD
<i>Convolutispora stigmoidea</i> Bharadwaj & Venkatachala, 1962								✓		AL–PD
<i>Corbulispora cancellata</i> (Waltz in Luber & Waltz, 1938) emend. Playford & Melo, 2012								✓		Rle–Mag
<i>Cordylosporites absimilis</i> (Daemon, 1974) emend. Playford & Melo, 2012		✓	✓	✓				✓		Rle–PD
<i>Cordylosporites glumaceus</i> (Byvsheva, 1972) emend. Playford & Melo, 2012		✓	✓	✓						Rle–PD
<i>Cristatisporites peruvianus</i> Azcuy & di Pasquo, 2005		✓		✓						Rle–Mag
<i>Cymbosporites loboziakii</i> Melo & Playford, 2012			✓	✓		✓			✓	AL–PD
<i>Cyrtospora cristifera</i> (Luber in Luber & Waltz, 1941) emend. Van der Zwan, 1979			✓	✓		✓			✓	Rad, Rle–Mag
<i>Densosporites annulatus</i> (Loose, 1932) Smith & Butterworth, 1967			✓			✓				Rle–Mag
<i>Densosporites infacetus</i> Daemon, 1974		✓	✓	✓		✓	✓		✓	LVa–PD
<i>Densosporites pseudoannulatus</i> Butterworth & Williams, 1958							✓	✓		Rle–Mag
<i>Diatomozonotriletes</i> sp.								✓		PD
<i>Dictyotriletes trivialis</i> Kedo, 1963			✓							Rle–PD
<i>Endoculeospora frumentifera</i> Melo & Playford, 2012				✓						AL–PD
<i>Endosporites micromanifestus</i> Hacquebard, 1957					✓					LVa?, AL–PD
<i>Grandispora senticosa</i> (Ishchenko, 1956) Byvsheva, 1985			✓							LVa–PD
<i>Indotriradites multivolus</i> (Daemon, 1974) emend. Melo & Playford, 2012		✓	✓		✓	✓		✓	✓	PD, Mag
<i>Knoxisporites hederatus</i> (Ishchenko, 1956) Playford, 1963								✓		Rad–PD
<i>Neoraistrickia cymosa</i> Higgs, Clayton & Keegan, 1988							✓			AL–PD
<i>Neoraistrickia loganensis</i> (Winslow, 1962) emend. Coleman & Clayton, 1987		✓	✓	✓	✓	✓	✓	✓	✓	BL, PD
<i>Punctatisporites</i> spp.		✓	✓	✓	✓	✓	✓	✓	✓	AL–Mag
<i>Radiizonates arcuatus</i> Loboziak, Playford & Melo, 2000		✓	✓	✓	✓	✓	✓	✓	✓	LVa–Mag
<i>Raistrickia baculosa</i> Hacquebard, 1957			✓	✓						AL–Mag
<i>Raistrickia clavata</i> Hacquebard, 1957 emend. Playford, 1964			✓							PD, Mag
<i>Raistrickia strumosa</i> Playford, 1976			✓	✓						BL, PD
<i>Spelaeotriletes balteatus</i> (Playford, 1963) Higgs, 1996		✓	✓	✓	✓		✓	✓		BL–Mag
<i>Spelaeotriletes pretiosus</i> (Playford, 1964) Utting, 1987		✓	✓		✓			✓		PD, Mag
<i>Tumulispora rarituberculata</i> (Luber in Luber & Waltz, 1941) Playford, 1991							✓	✓		Rle–PD
<i>Vallatisporites banffensis</i> Staplin & Jansonius, 1964			✓							Rle–Mag
<i>Vallatisporites splendens</i> Staplin & Jansonius, 1964		✓	✓	✓				✓		AL–Mag
<i>Vallatisporites vallatus</i> Hacquebard, 1957		✓	✓	✓				✓		LVa–Mag
<i>Vallatisporites verrucosus</i> Hacquebard, 1957				✓			✓	✓		Rle–Mag
<i>Verrucosisporites congestus</i> Playford, 1964			✓			✓	✓	✓		AL–Mag
<i>Verrucosisporites depressus</i> Winslow, 1962						✓	✓			AL–Mag
<i>Verrucosisporites mesogrumosus</i> (Kedo, 1963) Byvsheva, 1985			✓							Rle–Mag
<i>Verrucosisporites nitidus</i> Playford, 1964			✓	✓		✓	✓	✓		LVa–Mag
<i>Waltzisporea lanzonii</i> Daemon, 1974							✓	✓	✓	AL–PD

Table 1. Checklist of miospores identified as autochthonous in eight representative samples studied from the Longá Formation, with summary of established zonal occurrences of the individual species in northern Brazilian basins based principally on Loboziak *et al.* (1992, 1993, 2000), Melo & Loboziak (2003), Melo & Playford (in press), and S. Loboziak, J.H.G. Melo, and G. Playford (unpublished data). Key to samples: 1, UN-24, 338.05 m; 2, UN-24, 317.75 m; 3, UN-24, 286.75 m; 4, UN-24, 283.80 m; 5, UN-24, 272.25 m; 6, UN-24, 250.55 m; 7, UN-27, 340.85 m; 8, UN-04, 318.00 m. Acronyms for miospore zones in right-hand column are after Melo & Loboziak (2003).

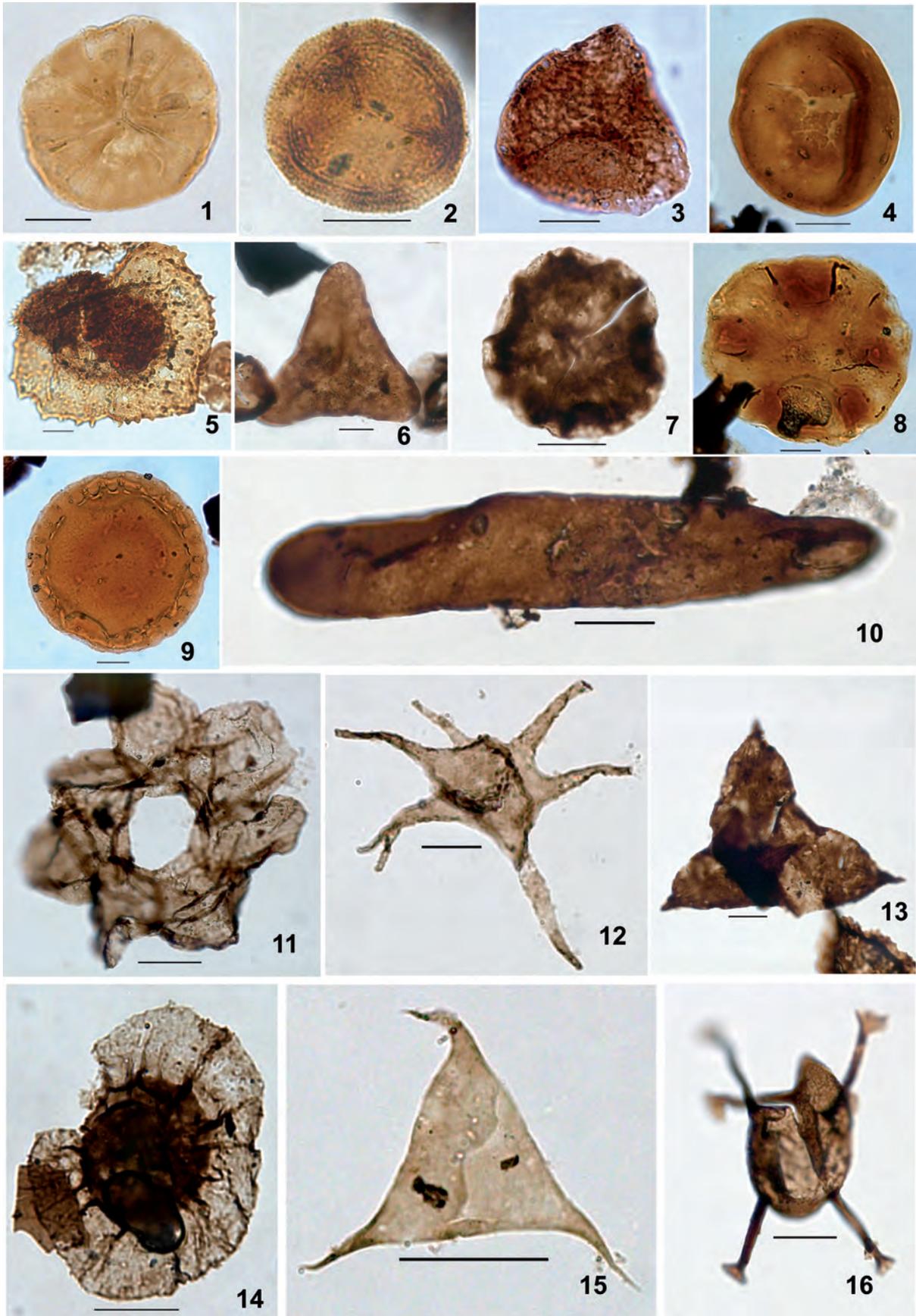
Recycled palynomorphs↓	Samples→	1	2	3	4	5	6	7	8	Time range
<i>Ancyrospora</i> sp.				✓						Middle–Late Devonian
<i>Chelinospora concinna</i> Allen, 1965							✓			late Givetian–early Frasnian
<i>Emphanisporites rotatus</i> McGregor, 1961		✓		✓	✓			✓		Devonian
<i>Geminospora lemurata</i> Balme, 1962 emend. Playford, 1983								✓		early Givetian–late Frasnian
<i>Geminospora</i> sp.								✓		Middle–Late Devonian
<i>Gneudnaspota divellomedia</i> (Chibrikova, 1959) Balme, 1988							✓	✓		Silurian–Devonian
<i>Grandispora</i> sp. cf. <i>G. protea</i> (Naumova, 1953) Moreau-Benoit, 1980			✓					✓		latest Emsian–earliest Frasnian
<i>Hystricosporites</i> spp.							✓			Middle–Late Devonian
<i>Leiotriletes struniensis</i> Moreau-Benoit, 1979				✓						mid–latest Famennian
<i>Retispora lepidophyta</i> (Kedo, 1957) Playford, 1976			✓	✓	✓	✓		✓		latest Famennian (Strunian)
<i>Diexallophasis remota</i> (Deunff, 1955) emend. Playford, 1977				✓	✓					Silurian–Devonian
<i>Duvernaysphaera angelae</i> Deunff, 1964					✓					Middle–Late Devonian
<i>Duvernaysphaera radiata</i> Brito, 1967			✓	✓		✓	✓	✓	✓	Late Devonian
<i>Estiastra rhytidia</i> Wicander & Wood, 1981				✓						Givetian–Frasnian
<i>Maranhites brasiliensis</i> Brito, 1965					✓		✓	✓		Givetian–Famennian
<i>Maranhites lobulatus</i> Burjack & Oliveira, 1989					✓					Late Devonian
<i>Maranhites mosesii</i> (Sommer, 1956) Brito, 1967						✓	✓	✓		Late Devonian
<i>Multiplicisphaeridium ramusculosum</i> (Deflandre, 1945) Lister, 1970		✓								Silurian–Late Devonian
<i>Navifusa bacilla</i> (Deunff, 1955) Playford, 1977					✓			✓		Emsian–Famennian
<i>Petrovina connata</i> Oliveira & Burjack, 1996		✓								(?)early–late Frasnian
<i>Stellinium micropolygonale</i> (Stockmans & Willièrè, 1960) Playford, 1977				✓		✓				Middle–Late Devonian
<i>Umbellasphaeridium saharicum</i> Jardiné <i>et al.</i> , 1972		✓	✓	✓	✓	✓			✓	Late Devonian
<i>Veryhachium trispinosum</i> (Eisenack, 1938) Stockmans & Willièrè, 1962 <i>sensu</i> Servais <i>et al.</i> (2007)			✓		✓		✓			Ordovician–Permian (and younger?)

Table 2. Checklist of palynomorphs identified as recycled in eight representative samples studied from the Longá Formation, with summary of their known time ranges. Key to samples: 1, UN-24, 338.05 m; 2, UN-24, 317.75 m; 3, UN-24, 286.75 m; 4, UN-24, 283.80 m; 5, UN-24, 272.25 m; 6, UN-24, 250.55 m; 7, UN-27, 340.85 m; 8, UN-04, 318.00 m. Acronyms for miospore zones in right-hand column are after Melo & Loboziak (2003).

rope, Asia, and elsewhere, have long been detected on the basis of stable isotope geochemical, eustatic, and sequence-stratigraphic proxies (Bruckschen *et al.*, 1999; Crowell, 1999; Mii *et al.*, 1999; Saltzman, 2002; Richardson, 2003; Matchen, 2004; Matchen & Kammer, 2006; Katz *et al.*, 2007; Kammer & Matchen, 2008; Qie *et al.*, 2011; and references therein). Biostratigraphic (palynological and microfaunal) data further strengthen the inter-regional correlations and constrain the age span of the Tournaisian glacial episode.

Stable isotopes – primarily $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (often coupled with $^{87}\text{Sr}/^{86}\text{Sr}$) – are used to identify climatic patterns (sea water palaeotemperatures), and to estimate rates of organic matter productivity and burial. Positive shifts in the fluctuation curves of both isotopes, detected in coeval sedimentary sections at different locations, commonly characterize cooler climates with concomitant depletion in atmospheric CO_2 . If these factors coincide with forced regressions and consequential extensive subaerial erosion, then glaciations may be implicated as the triggering mechanism. Such conditions are met, e.g., for several Kinderhookian/Osagean (= Tn2/Tn3) transition sections of

Plate 3. Selected recycled miospores (Figures 1-6), prasinophyte phycocmata (Figures 7-9), and acritarchs (Figures 10-16) from the Longá Formation from DNPM core-drills UN-27 (Figures 1-5, 8, 9) and UN-24 (Figures 6, 7, 10-16). Information provided for each figured specimen comprises sample code and depth, preparation/slide no., and England Finder coordinates. Scale bars 20 μm in length. 1, *Emphanisporites rotatus* McGregor, 1961, P3-340.85 m, 200607773, O37. 2, *Geminospora lemurata* Balme, 1962 emend. Playford, 1983, P3-340.85 m, 200607773, U52/1. 3, *Retispora lepidophyta* (Kedo, 1957) Playford, 1976, P3-340.85 m, 200607773, M66. 4, *Gneudnaspota divellomedia* (Chibrikova, 1959) Balme, 1988, P3-340.85 m, T43. 5, *Grandispora* sp. cf. *G. protea* (Naumova, 1953) Moreau-Benoit, 1980, P3-340.85 m, 200607773, Y57. 6, *Leiotriletes struniensis* Moreau-Benoit, 1979, P16-286.75 m, CB20110212261-1, V58/2. 7, *Maranhites lobulatus* Burjack & Oliveira, 1989, R10-283.80 m, CB20110212219-1, M38/1. 8, *Maranhites mosesii* (Sommer, 1956) Brito, 1967, P3-340.85 m, 200607773, W55/3. 9, *Maranhites brasiliensis* Brito, 1965, P3-340.85 m, 200607773, H57. 10, *Navifusa bacilla* (Deunff, 1955) Playford, 1977, R13-327.70 m, CB20110212221-1, T56. 11, *Petrovina connata* Oliveira & Burjack, 1996, P21-338.05 m, CB20110212266-1, R56. 12, *Diexallophasis remota* (Deunff, 1955) emend. Playford, 1977, P17-293.40 m, CB20110212262-1, H41. 13, *Estiastra rhytidia* Wicander & Wood, 1981, P16-286.75 m, CB20110212261-1, V52. 14, *Duvernaysphaera radiata* Brito, 1967, P19-317.75 m, CB200909881-1, Q44/4. 15, *Veryhachium trispinosum* (Eisenack, 1938) Stockmans & Willièrè, 1962 “group” (*sensu* Servais *et al.*, 2007), R10-283.80 m, CB20110212219-1, W48/2. 16, *Umbellasphaeridium saharicum* Jardiné, Combaz, Magloire, Peniguel & Vachey, 1972, P16-286.75 m, CB20110212261-1, N53/1.



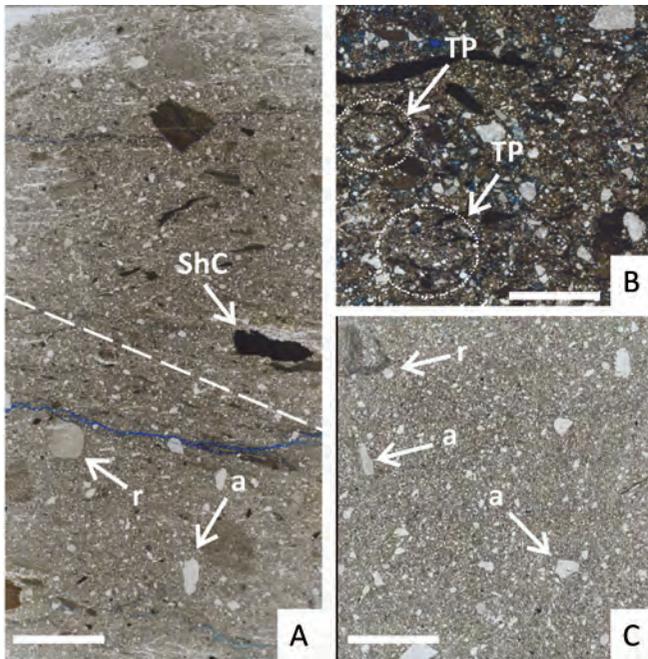


Figure 3. Photomicrographs of diamictites in thin section (plane-polarized light; scale bar = 5 mm). Core-drill samples: A, UN-24, 327.70 m; B, UN-24, 283.80 m; C, well UN-24, 335.40 m. Diamictites are, in general, extremely poorly sorted, corresponding petrographically to pebbly, fine-grained sandstones and mudstones. Note: A, small shale clasts (ShC) and subtle fabric alignment (marked by dashed line); B, till pellets (TP); C, crushed nature of the quartz sand matrix; A, C, rounded (r) and angular (a) quartz grains.

western, Midcontinental, and eastern U.S.A. (see discussions in Saltzman, 2002; Matchen, 2004; Matchen & Kammer, 2006; Katz *et al.*, 2007; Kammer & Matchen, 2008). These isotopic fluctuations can be correlated with the PD-age glacial episode documented in the Parnaíba Basin (this paper; cf. Figure 4), in other Palaeozoic basins of northern Brazil, and possibly elsewhere in South America (Caputo, 1984; Loboziak, Melo, Quadros *et al.*, 1995; Caputo *et al.*, 2008).

Parallel positive anomalies of both oxygen and carbon stable isotopes – recorded in Tn2/Tn3 boundary marine limestones, conodont apatite, and brachiopod shell calcite from Euramerica – have been cited as evidence for global cooling in the early late Tournaisian within the upper *Siphonodella crenulata*–*S. isosticha* conodont Zone (Bruckschen *et al.*, 1999; Mii *et al.*, 1999; Saltzman, 2002; Saltzman *et al.*, 2004; Buggisch & Joachimski, 2006; Buggisch *et al.*, 2008). Coeval hiatuses and incised valley fills, implying a major sea-level fall, have been documented in

the Midcontinent and Appalachian basins of eastern U.S.A. (Matchen, 2004; Matchen & Kammer, 2006; Kammer & Matchen, 2008). Outside North America, a global sequence boundary associated with widespread regression and unconformity at the Tn2/Tn3 boundary is also identified in Saudi Arabia, Belgium, Czech Republic, Germany, western Russia, Siberia, and China (see Kammer & Matchen, 2008; Qie *et al.*, 2011; and references therein). Collectively, these various lines of evidence lend robust support to the recognition of a late middle to early late Tournaisian glaciation in Gondwana.

A positive carbon isotope ($\delta^{13}\text{C}$) excursion has been recognized in late Kinderhookian and early Osagean carbonates of western U.S.A. (southeastern Idaho and Nevada: Saltzman, 2002, 2003). The oldest $\delta^{13}\text{C}$ peak (+7‰) is within the *Siphonodella isosticha* conodont Zone (uppermost Kinderhookian), and a younger peak occurs in the lower *Gnathodus typicus* conodont Zone (lowest Osagean). These values, reported from widely separated basins, display a consistent trend that can be correlated with Early Mississippian curves generated from brachiopod calcite in Western Europe and the U.S. Midcontinent. They are interpreted to reflect primary, glacially-driven seawater fluctuations, albeit partly masked by the tectonic effects of the Antler Orogeny in western U.S.A. (Saltzman *et al.*, 2000; Saltzman, 2002; Saltzman *et al.*, 2004). Sea-level changes occurred near the beginning of the late Kinderhookian $\delta^{13}\text{C}$ shift (upper to middle parts of the *S. isosticha* Zone) and within the peak of the $\delta^{13}\text{C}$ excursion (Kinderhookian/Osagean boundary). Likewise, $\delta^{18}\text{O}$ values, measured mainly from brachiopod shells and conodont apatite, become more positive up-section, thus paralleling the positive trend in $\delta^{13}\text{C}$ during the late Kinderhookian, not only in western U.S.A. (Saltzman, 2002) but also in coeval (late Hastarian) sections of Western and Eastern Europe (Saltzman *et al.*, 2004; Buggisch *et al.*, 2008). This suggests declining sea water temperatures and increasing ice buildup in higher palaeolatitude regions of Gondwana during the middle/late Tournaisian transition.

No subaerial exposure surfaces have been recognized in the southeastern Idaho and Nevada sections investigated by Saltzman (2002), but they do occur farther to the east, in coeval sedimentary intervals of the North American Midcontinent and the central Appalachian and Michigan

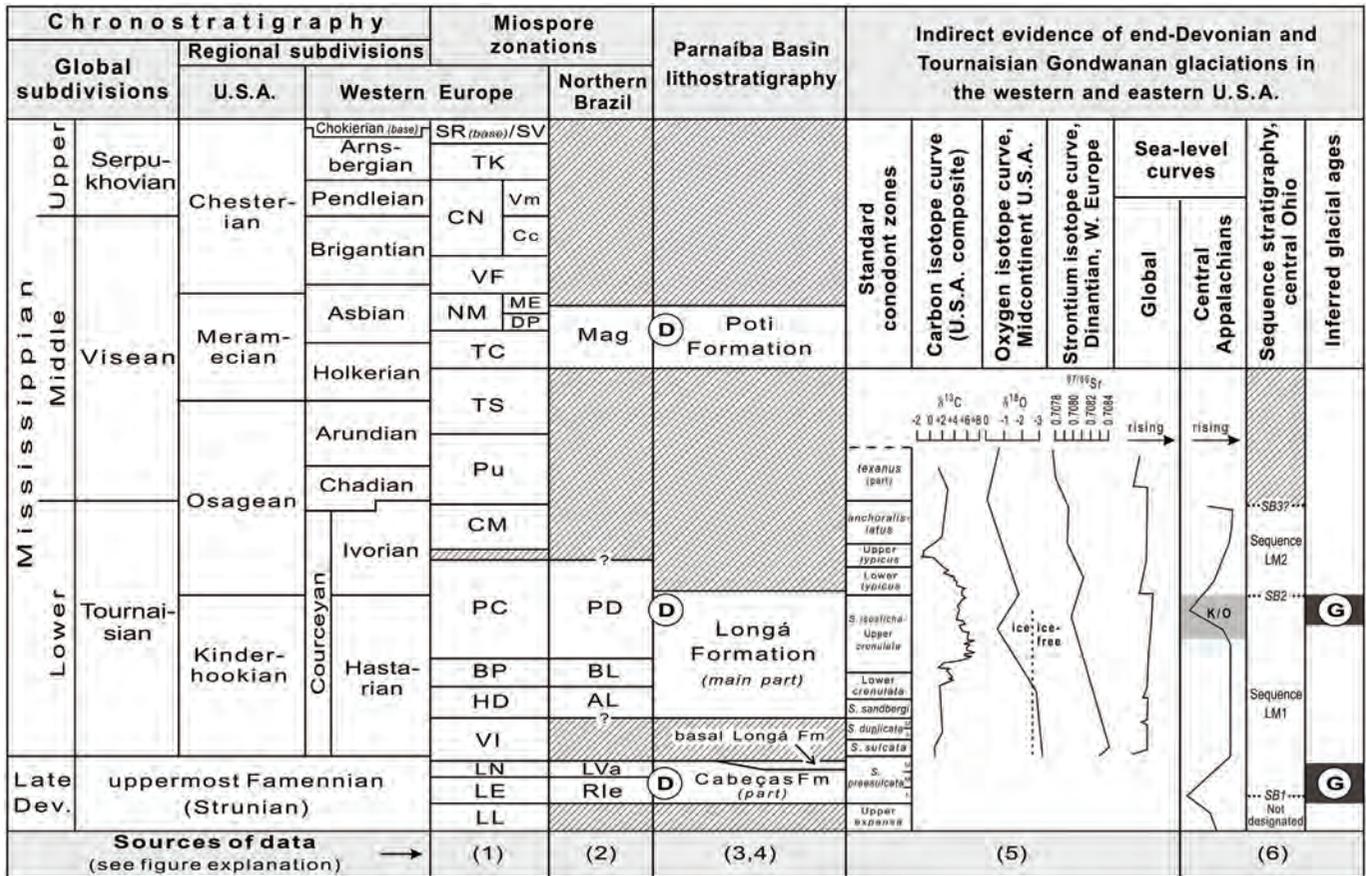


Figure 4. Palynological (miospore-based) correlation of uppermost Devonian–Mississippian lithostratigraphic units of the Parnaíba Basin with the standard global, North American, and Western European chronostratigraphic subdivisions (Menning *et al.*, 2006). Other sources of data, enumerated in bottom line of figure, are as follows: 1, miospore zonation for Western Europe according to Higgs *et al.* (2004, 2005) and Owens *et al.* (1992) and Higgs *et al.* (2004, 2005); 2, northern Brazilian miospore zones after Melo & Loboziak (2000, 2003) and Melo & Playford (in press); 3, Parnaíba Basin lithostratigraphy after Vaz *et al.* (2007), with age ranges of rock units modified by Melo & Loboziak (2000, 2003) and the present authors. 4, encircled letters “D” denote stratigraphic occurrences of diamicrites and other glaciogenic rocks in the Parnaíba Basin, as recognized by Caputo *et al.* (2008; Cabeças and Poti formations) and the present authors (Longá Formation); 5, conodont zonation, isotope data, and global sea-level curves after Saltzman (2002) and references cited therein; 6, Central Appalachians sea-level curve, grey-shaded “K/O” band (highlighting forced regression near the Kinderhookian/Osagean boundary), central Ohio sequence stratigraphy including sequence boundaries (SBs) 1-3, and inferred end-Devonian and Tournaisian glacial ages (indicated by black bands with encircled letters “G”), based on Matchen (2004), Matchen & Kammer (2006), Kammer & Matchen (2008), and references cited therein. See text for discussion.

basins (Matchen, 2004; Matchen & Kammer, 2006; Kammer & Matchen, 2008). The Kinderhookian/Osagean boundary, as defined in the Mississippi Valley stratotype region (Kinderhook, Illinois) and also along the Osage River, in Missouri, is actually a physical unconformity that accounts for two missing conodont biozones (i.e., the uppermost *Siphonodella crenulata*–*S. isosticha* Zone and the *Gnathodus punctatus* Zone, *fide* Lane & Brenckle, 2005). The *G. punctatus* Zone, of earliest Osagean age, is essentially correlative with the lowest part of the *G. typicus* Zone of the standard Mississippian conodont zonation (see Buggisch *et al.*, 2008). The implied gap is estimated at about 2 Ma or slightly less, according to calculations

based independently on conodont biochronology and the duration of the $\delta^{13}\text{C}$ event at the Kinderhookian/Osagean boundary (Saltzman, 2001; Kammer & Matchen, 2008).

The same unconformity can be traced into the nearby Michigan Basin and Appalachian areas to the east/southeast (Ohio, West Virginia, and Pennsylvania). In all of these, it constitutes a sequence boundary consistently blanketed by variably thick, variously named, coarse-grained fluvial sandstones at the Kinderhookian/Osagean boundary, which are mostly interpreted as incised valley fills (Matchen, 2004; Matchen & Kammer, 2006; Kammer & Matchen, 2008). The southern/central and northeastern Ohio occurrences are of

particular interest to the present study, because of the reliable miospore-based biostratigraphic controls available for Early Mississippian sections of those regions (Winslow, 1962; Clayton *et al.*, 1998), which permit direct correlation with the Parnaíba Basin's Mississippian palynozonation.

In central Ohio, palaeovalleys incised in the Kinderhookian/Osagean unconformity (= Sequence Boundary SB2 of Matchen, 2004) are infilled with the Black Hand Sandstone, which is formally regarded as the uppermost member of the underlying (Kinderhookian) Cuyahoga Formation. The latter is part of the "Lower Mississippian Sequence 1" (LM1) of Matchen (2004) and Matchen & Kammer (2006). The Black Hand is transgressed by the sandstones and shales of the Logan Formation, which makes up "Lower Mississippian Sequence 2" (LM2) of the same authors (Figure 4). The total age span of the Logan Formation is limited to the pre-Viséan (i.e., late Tournaisian) portion of the Osagean, in view of the *Scaliognathus anchoralis*–*Doliognathus latus* Zone conodonts (Thompson *et al.*, 1971) recovered from the Logan's uppermost subunit, the Rushville Limestone, which some authors regard as an independent formation resting unconformably on the Logan Formation (*sensu stricto*; e.g., Hansen, 2001). Palynofloras diagnostic of the late middle to early late Tournaisian PC Zone (Clayton *et al.*, 1977, 1978; Higgs *et al.*, 1988) occur both below and above the Kinderhookian/Osagean unconformity in Ohio, i.e., respectively within the Cuyahoga Formation and in the bulk of the overlying Logan Formation (Winslow, 1962; Clayton *et al.*, 1998). In conterminous eastern/northeastern Kentucky, the PC–Zone and succeeding CM Zone characterize virtually the entire Borden Formation (containing Osagean ammonoids and *G. typicus* Zone conodonts; cf. Richardson & Ausich, 2004). This unit rests unconformably on early late Kinderhookian strata (basal part of the Henley Bed and underlying Sunbury Formation), which bear VI–(HD?) equivalent palynofloras, conodonts of the uppermost *S. duplicata* through lowest *S. crenulata* zones, and ammonoids (Coleman & Clayton, 1987; Sandberg *et al.*, 2002; Richardson, 2003; Heal *et al.*, 2009, and references cited therein; Work & Mason, 2009).

The PC (*Spelaeotriletes pretiosus*–*Raistrickia clavata*) miospore Zone, of late middle to early late Tournaisian

age, is the Euramerican equivalent of the PD (*Spelaeotriletes pretiosus*–*Colatisporites decorus*) Zone of Western Gondwana (Melo & Loboziak, 2003). The latter characterizes the glacio-marine diamictites of the upper Longá Formation in the Parnaíba Basin (Lobato & Borghi, 2007; present paper). In addition to diamictites noted in the Oriximiná Formation (Caputo *et al.*, 2008), the PD Zone also comprises thick sandstone bodies of early late Tournaisian age in the eastern Amazonas Basin (lower Faro Formation sandstones with subordinate diamictite interbeds: Caputo, 1984; Melo & Playford, in press; see also Loboziak *et al.*, 1998, Fig. 2, and Melo & Loboziak, 2003, Fig. 11). The same is evident in the eastern Solimões Basin (Juruá Sub-basin), where the Chibata Sandstone occurs laterally to PD-age diamictites and mudstones of the uppermost Jandiatuba Formation, which are known from the subsurface of the Juruá and Tefé river areas (Loboziak, Melo, Quadros *et al.*, 1995; J.H.G. Melo and collaborators, unpublished data). The Chibata and lower Faro sandstones of northern Brazil – both associated with diamictites – could represent regional analogues of the Black Hand Sandstone and other coeval incised-valley fills of eastern North America (e.g., Burgoon, Big Injun, and Purslane sandstones), thus testifying to a global, glacially-driven, forced regression at the middle/late Tournaisian boundary. In northern Bolivia (Madre de Dios Basin), PD palynofloras also occur in possibly glaciogenic sediments of the Toregua Formation, dated as late middle or early late Tournaisian in the Pando PND-x1 well (interval 1095.39–1119.38? m, partly associated with diamictites; Isaacson *et al.*, 1995; J.H.G. Melo and collaborators, unpublished data). Caputo *et al.* (2008) briefly reviewed similar occurrences of glacially derived rock units of doubtful Tournaisian age in Bolivia, Peru, and Argentina; some of these are hardly distinguishable from their end-Devonian equivalents (immediately subjacent) because of the very conspicuous recycling of Devonian palynomorphs.

From the above, it is concluded that upper PD glaciogenic strata (glacio-marine diamictites and more proximal, channel/?deltaic sandstones) in the Parnaíba, Amazonas, and Solimões basins of northern Brazil (and possibly elsewhere in South America) are correlative with: (a) the widespread Kinderhookian/Osagean unconformity over much of North America (associated with incised valley fills in the eastern U.S.A.); (b) unconformities at the middle/upper Tournaisian boundary in Europe, Asia, and the Middle

East; and (c) a major, positive carbon/oxygen isotope excursion of near-global recognition. The latest middle to earliest late Tournaisian ice age corresponds to two missing conodont zones in the type area of the Kinderhookian/Osagean boundary (Illinois and Missouri, U.S.A.), which together encompass no more than about 2 Ma (Kammer & Matchen, 2008). Therefore, this was seemingly shorter-lived (and possibly also far less extensive geographically) than the latest Famennian ice age, which in the Parnaíba Basin is recorded within the Cabeças Formation (Figure 4). The latter event may have comprised multiple glaciations lasting no more than ca. 100 Ka each (Streel, 1999; M. Streel, written communication, 2012); altogether they correspond to parts of the combined LE and LN miospore zones of Western Europe, i.e., to much of the *Siphonodella praesulcata* conodont Zone (Streel *et al.*, 1987, 2000, 2011; Caputo *et al.*, 2008). The implied time span is probably less than 3.8 Ma, according to the Ogg *et al.* (2008) timescale.

Another aspect to be highlighted is that no younger Tournaisian strata have been recorded above the PD-age glaciogenic sections of the Parnaíba Basin (uppermost Longá Formation), which are unconformably overlain by early late Viséan rocks of the Poti Formation containing the distinctive Mag (*Reticulatisporites magnidictyus*) zonal palynoflora (Grahm *et al.*, 2001; Melo & Loboziak, 2000, 2003). The same apparently applies to the upper PD strata of the Amazonas and Solimões basins to the west, which are also succeeded directly by early late Viséan or younger rock units (Loboziak, Melo, Quadros *et al.*, 1995; Melo & Loboziak, 2003; Caputo *et al.*, 2008). Therefore, an age not younger than earliest Osagean (= earliest Ivorian, earliest late Tournaisian) can be assumed for those highest PD intervals, based on the aforementioned correlations. On the other hand, the equivalent palynozone of Western Europe and eastern U.S.A., the PC Zone, is known to extend above the Tn2/Tn3 boundary (i.e., well into the *Pseudopolygnathus multistriatus*/*Gnathodus typicus* conodont zones), perhaps reaching the lowermost Tn3c (Clayton *et al.*, 1978; Higgs *et al.*, 1988; Riley, 1993). It follows that the upper limit of the PD Zone of Melo & Loboziak (2003) may not coincide with that of the (partly younger?) PC Zone of Euramerica (Coleman & Clayton, 1987; Higgs *et al.*, 1988, 1992; Clayton *et al.*, 1998). However, confirmation of this hypothesis will require additional palynological investigations in the Mississippian sections of

northern Brazil, and in faunally dated Tournaisian rocks assigned to the uppermost PC Zone in Western Europe and the U.S.A.

Lastly, it should be added that no Mississippian glacial episodes older than the PC Zone (or doubtfully the BP–PC zonal range, after Loboziak, Melo, Quadros *et al.*, 1995) have been identified in Palaeozoic basins of Brazil and other South American regions, contrary to the interpretation of Isaacson *et al.* (2008). These authors depicted putative diamictite occurrences in the “VI–HD” zonal range (= AL Zone of Melo & Loboziak, 2003), in addition to older Famennian (pre-Strunian) ones, none of which is palynologically proven in South American basins (Streel, 2006; J.H.G. Melo, unpublished Bolivian and Brazilian data).

7. CONCLUSIONS

1. The upper part of the Longá Formation in the Parnaíba Basin subsurface provides persuasive evidence of glacial depositional conditions in the form of diamictites and glacially deformed strata.
2. The abundantly preserved palynoflora assemblage hosted by the cored sediments is clearly representative of the *Spelaeotriletes pretiosus*–*Colatisporites decorus* [PD] miospore Interval Zone, which was originally defined in the adjacent Amazonas Basin and dated as Early Mississippian, specifically late middle to early late Tournaisian, thereby encompassing the Kinderhookian/Osagean transition. The PD Zone can be correlated with the bulk of the PC Zone of Euramerica, even though the post-earliest Osagean (= post-glacial) part of the latter may be unrepresented in Brazil.
3. Recycled palynomorphs (miospores and organic-walled microphytoplankton) of Middle and Late Devonian age, as components of the palynofloras, testify to the incorporation of erosional products of the glacial activity.
4. Brazilian occurrences of palynologically dated, middle/late Tournaisian diamictites (with confirmed records in the Solimões, Amazonas, and Parnaíba basins) provide support for an important glacial

episode about the Kinderhookian/Osagean boundary, as detected in other continents on the basis of sequence-stratigraphic and stable isotope data.

5. This study demonstrates that accurately identified palynofloras can provide a reliable substitute for marine faunas as dating tools in high-palaeolatitude, cool to cold water regions, where thermally sensitive organisms like conodonts and ammonoids are usually scarce or absent.

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Taxonomía y bioestratigrafía de conodontos silúricos en el pliegue acostado del Courel (Lugo, España)

Roselis W. Salazar Ramírez¹, Graciela N. Sarmiento¹ y Juan Carlos Gutiérrez-Marco²

¹ Departamento de Paleontología, Facultad de Ciencias Geológicas (UCM), José Antonio Novais 2, Ciudad Universitaria, 28040 Madrid. rwsalazar@estumail.ucm.es; gsarmien@geo.ucm.es

² Instituto de Geociencias (CSIC-UCM), José Antonio Novais 2, 28040 Madrid. jcgrapto@geo.ucm.es

RESUMEN

En este trabajo se describen, ilustran y analizan desde un punto de vista taxonómico y bioestratigráfico las asociaciones de conodontos registradas en materiales del Silúrico en el Sinclinal del Courel, provincia de Lugo, España. Las formas identificadas corresponden a *Dapsilodus obliquicostatus* (Branson y Mehl, 1933), *Oulodus* cf. *siluricus* (Branson y Mehl, 1933), *Kockelella* cf. *variabilis* (Walliser, 1957), "*Ozarkodina*" *bohemica* (Walliser, 1964), "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964), *Wurmiella excavata excavata* (Branson y Mehl, 1933), *Walliserodus* sp., *Kockelella?* sp. y *Ozarkodina?* sp. A partir de los resultados obtenidos se atribuyen los niveles investigados al intervalo comprendido entre la parte superior de la Zona de "*O*". *sagitta sagitta* y la parte inferior de la Zona de *K. variabilis*, cuyo rango se extiende desde el Wenlock superior (Homeriense) hasta el Ludlow inferior (Gorstiense).

Palabras clave: Conodontos, Silúrico, taxonomía, Courel, Lugo, España.

ABSTRACT

[*Taxonomy and biostratigraphy of Silurian conodonts from the Courel recumbent fold (province of Lugo, Spain)*]. Conodont assemblages registered in materials from the Silurian Courel syncline, province of Lugo (Spain), are described, illustrated and discussed from a taxonomic and biostratigraphic perspective. The identified conodonts are *Dapsilodus obliquicostatus* (Branson and Mehl, 1933), *Oulodus* cf. *siluricus* (Branson and Mehl, 1933), *Kockelella* cf. *variabilis* (Walliser, 1957), "*Ozarkodina*" *bohemica* (Walliser, 1964), "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964), *Wurmiella excavata excavata* (Branson and Mehl, 1933), *Walliserodus* sp., *Kockelella?* sp, and *Ozarkodina?* sp. The results obtained allowed to attribute the studied levels to the interval between the top of the "*O*". *sagitta sagitta* Zone and the bottom of the *K. variabilis* Zone, which range extends from the upper Wenlock (Homerian) to the lower Ludlow (Gorstian).

Keywords: Conodonts, Silurian, taxonomy, Courel, Lugo, Spain.

1. INTRODUCCIÓN Y ANTECEDENTES

El gran pliegue acostado del Courel se sitúa al sur de la provincia de Lugo y es una de las estructuras geológicas más sobresalientes de la Cadena Varisca ibérica. Tras ser

reconocido como Punto de Interés Geológico por el Instituto Geológico y Minero de España (Duque Lucas *et al.*, 1983), figura ya destacado entre los contextos geológicos españoles de relevancia internacional (García Cortés *et al.*, 2001, 2008; Carcavilla Urquí y Palacio Suárez-Val-

grande, 2010), y recientemente fue declarado *Monumento Natural* por la Xunta de Galicia (7 de mayo de 2012), en atención a su indiscutible contribución al patrimonio geológico nacional. El mirador y un museo próximo garantizan su contemplación por los visitantes (Gutiérrez-Marco, 2005; Vidal Romaní, 2011).

Desde el punto de vista geológico, el pliegue se integra en un gran conjunto anticlinal-sinclinal, de planos axiales prácticamente horizontales en el sector central de la Sierra do Courel, que lateralmente se van enderezando tanto en su extremo occidental, como en su prolongación oriental (Sinclinal del Courel-Peñalba, anticlinales del Piornal y Teleno). El estudio estratigráfico y estructural de la Sierra do Courel fue abordado esencialmente por Riemer (1963, 1966), Matte (1968), Iglesias Ponce de León y González Lodeiro (1981), Abril Hurtado y Rodríguez Fernández (1981) y Martínez Catalán (1985). Los modelos geodinámicos más recientes son los de Martínez Catalán *et al.* (1992) y Fernández *et al.* (2011), que datan la estructura en la primera fase de deformación varisca, nacida de un proceso de tipo "buckling", o alternativamente por inversión tectónica, en ambos casos predeterminados por una tectónica sinsedimentaria que, en todo caso, dio paso a un importante acortamiento por cizalla simple y con aplastamiento tectónico de las unidades marinas del Paleozoico.

Pese a la importante deformación sufrida por los materiales cámbrico-devónicos, en la Sierra do Courel se conocen algunos yacimientos paleontológicos con icnofósiles y restos de invertebrados del Ordovícico Inferior y Medio, Silúrico y Devónico Inferior (Hernández Sampelayo, 1915, 1932, 1935; Riemer, 1963, 1966; Matte, 1968; Guillou, 1969; Abril Hurtado y Rodríguez Fernández, 1981; Gutiérrez-Marco *et al.*, 1999), que permanecen sin estudiar en detalle. En el presente trabajo aportamos el primer hallazgo de microfósiles (conodontos) dentro de la sucesión silúrica del flanco inverso del Sinclinal del Courel, que reviste interés para la datación y correlación de las raras facies carbonatadas de este sistema conocidas en el noroeste de la Península Ibérica.

En el área de estudio los hallazgos de conodontos más antiguos corresponden a la Zona de *Amorphognathus ordovicicus* del Ordovícico Superior, y provienen de la Formación Calizas de La Aquiana, concretamente del flanco meridional del Sinclinal de Peñalba (Sarmiento *et*

al., 1999). También se asignan a este intervalo bioestratigráfico los conodontos descritos en ambos flancos del Sinclinario del Sil-Truchas, tanto en la Fm. La Aquiana, como en el nivel carbonatado intercalado en la parte media-superior de la Formación Casaio, así como los conodontos registrados en los cantos heterométricos de caliza incluidos en la Formación Rozadais (Sarmiento, 1993; Sarmiento *et al.*, 1999). Otras citas de conodontos de la Zona de *A. ordovicicus* en el borde septentrional (Dominio del Olló de Sapo) de la Zona Centroibérica, se sitúan en la Unidad del Río Duero del Sinforme de Alcañices (Sarmiento *et al.*, 1997). Dentro de la Zona Cantábrica, los únicos datos sobre conodontos ordovícicos se limitan a la Caliza de Portilla de Luna (Kralodvoriense), igualmente con especies de la misma biozona (Del Moral, 2003).

Gutiérrez-Marco *et al.* (2001b) describen e ilustran una asociación de fósiles del Silúrico superior (bivalvos, cefalópodos, conodontos y radiolarios), procedente de una localidad próxima a La Barosa (León), en el flanco sur del Sinclinal de Peñalba. Se trata de un pequeño lentejón calcáreo intercalado en la base de una unidad de pizarras con cloritoide, donde se identificaron los conodontos *Ozarkodina excavata excavata* (Branson y Mehl), *O. confluens* (Branson y Mehl), *Oulodus siluricus* (Branson y Mehl), *Pseudooneotodus imulato* (Bischoff y Sannemann) y *Kockelella variabilis* Walliser?, una asociación asignada al Ludfordiense (Ludlow medio). Los moluscos y los trilobites del Ludlow encontrados en localidades próximas, muestran afinidades paleobiogeográficas con otras regiones del norte de Gondwana, especialmente con el área de Bohemia (República Checa).

En el mismo sector de La Barosa, Sanz López *et al.* (2000) realizan un estudio bioestratigráfico de la caliza del Devónico Inferior (Formación "Peñalba") del núcleo de este sinclinal. Dividen la unidad en dos miembros, de los cuales el inferior (Miembro Seceda) contiene conodontos del Lochkoviense inferior (*Ozarkodina carlsi* del tramo B1) y medio (*Lanea omoalpha* Murphy y Valenzuela del tramo B2b); en tanto que el miembro Carucedo brinda en su parte superior los primeros conodontos del Praguense inferior (*Icriodus angustoides castilianus* Carls e *I. cf. imulato* Carls).

Fuera de la región, los únicos datos sobre conodontos silúricos del noroeste de España provienen de localidades

aisladas del Rhuddaniense-Telychiense de la Zona Cantábrica (Sarmiento *et al.*, 1994), del Prídoli del norte de la Zona Centroibérica (Sarmiento *et al.*, 1997) y del intervalo Ludlow-Prídoli de la Zona de Galicia-Trás-os-Montes (Piçarra *et al.*, 2006).

2. MARCO GEOLÓGICO DE LA LOCALIDAD ESTUDIADA

La sucesión silúrica de la Sierra do Courel consta esencialmente de una unidad compresiva, de hasta 450 m de potencia, formada por *ampelitas*, pizarras negras con cloritoide y areniscas (Iglesias Ponce de León y González Lodeiro, 1981), descrita por Dozy (1983) con el nombre de Formación Folgoso, por sus amplios afloramientos en torno a Folgoso do Courel. Las “ampelitas” (= pizarras negras graptolíticas) suelen restringirse al tramo basal de la unidad (0-50 m), donde fueron diferenciadas por Riemer (1963) como la parte baja de sus Pizarras “Ordogot” superiores (*Ordogot* es la antigua contracción alemana del sistema *Ordoviciense+Gothlandiense*) y también por Matte (1968: “ampelitas del Llandovery inferior”). El resto de la sucesión la componen pizarras con cloritoide de aspecto masivo, con raras intercalaciones arenosas y delgados y esporádicos lentejones calcáreos. Dichas litologías se extienden hasta la base de las areniscas y calizas del Lochkoviense-Pragiense (Formación “Portela” o “Peñalba”). Las pizarras con cloritoide fueron reconocidas por Riemer (1963, 1966: parte alta de sus Pizarras “Ordogot” superiores) y por Matte (1968: “pizarras negras con cloritoide y *Monograptus*”).

Matte (1968) fue el primer autor en advertir la complejidad de la sucesión silúrica del Sinclinal del Courel, que en el sector oriental del pliegue descansa en aparente concordancia sobre las calizas del Ordovícico Superior (Fm. Aquiana), mientras que en el occidental llega a apoyarse directamente sobre la Cuarcita Armoricana o incluso sobre materiales más antiguos (al sur del balneario de Incio, en el flanco normal del sinclinal). Es por ello que el autor propone una “discordancia del Silúrico superior”, que se manifiesta tanto en el seno de la sucesión de este periodo (separando las ampelitas de las pizarras con cloritoide), como en forma de un contacto transgresivo directo de las pizarras del Wenlock-Ludlow,

sobre diferentes términos de la sucesión ordovícica. La cartografía MAGNA a escala 1:50.000 de la Sierra do Courel (Iglesias Ponce de León y González Lodeiro, 1981; Abril Hurtado y Rodríguez Fernández, 1981) permite visualizar cómo los materiales silúricos contactan regionalmente con distintos tramos de la Cuarcita Armoricana, con las pizarras del Ordovícico Medio (Formación “Luarca” *s.l.*) o con la Caliza de La Aquiana, incorporando localmente en la base el llamado “hard ground ferruginoso”. Este es un horizonte discontinuo que fue explotado a lo largo de todo el Sinclinal del Courel-Peñalba, pero que en realidad es post-tectónico y carece de relación alguna con la sedimentación paleozoica (Gutiérrez-Marco *et al.*, 2001).

Guillou (1969, 1971) analizó en detalle los materiales del tránsito Ordovícico-Silúrico en la Sierra do Courel, e introdujo novedades estratigráficas para la sucesión comprendida entre el techo de la Caliza de La Aquiana y la base de las pizarras negras silúricas. También detectó por vez primera la presencia de “niveles calcáreos con *Monograptus* y *Orthóceras*”, intercalados en las “ampelitas del Wenlock” en torno a las minas de Vilarbacú, al noroeste de Quiroga (Lugo). De acuerdo con el autor francés, en este sector se observan, a techo de la caliza ordovícica, 2-6 m de pizarras grises con anomalía geoquímica de antimonio (“pizarras intermedias”), sucedidas por hasta 6-8 m de “areniscas lenticulares del Valentiense”, sobre las que se apoyan a su vez las ampelitas silúricas. Las “pizarras intermedias” parecen presentar un tránsito gradual en la base (con niveles calcáreos delgados) y en el techo (intercalaciones limolíticas finas) con respecto a las unidades que las enmarcan, y fueron correlacionadas por Guillou (1969) con las facies de pizarras con cantos de caliza del sinclinal del Sil, equiparables en nuestra opinión a los sedimentos glaciomarineros del Hirnantense del Sinclinorio de Truchas (Formación Casaio). Por su parte, el autor sustenta la datación silúrica de las “areniscas del Valentiense” en una correlación tentativa con la Cuarcita del Criadero de Almadén (Zona Centroibérica meridional). Finalmente, las intercalaciones calcáreas las sitúa, dentro de las pizarras silúricas, “algunos metros por encima de las areniscas” y les atribuye “los caracteres pelágicos clásicos del medio ampelítico”, merced al hallazgo de graptolitos, ortocerátidos y algunas placas columnares de equinodermos.

Abril Hurtado y Rodríguez Fernández (1981) encontraron afloramientos similares de caliza con ortocerátidos al noroeste de A Seara. Se trata de dos lentejones calizos, de potencia menor de 1 m, intercalados en la sucesión de pizarras con cloritoide del flanco normal del sinclinal. Entre los fósiles reconocidos citan los ortocerátidos *Michelinoceras icrofós* Barrande y *M. dulce* Barrande, braquiópodos (orthísidos y *Protochonetes* sp.) y "*Coleolus*" sp., "todos ellos del Silúrico Superior" (Abril Hurtado y Rodríguez Fernández, 1981, pág. 23). Por encima de los niveles de caliza, los autores cartografiaban hasta cuatro intercalaciones de cuarcita de color gris azulado, que alcanzan un desarrollo lateral notable al sur de Ferreira de Abaixo y entre A Seara y Gestoso, ya en la provincia de León.

En el presente trabajo describimos el hallazgo de un nuevo afloramiento lenticular de caliza, intercalado en la sucesión silúrica correspondiente al flanco inverso del Sinclinal del Courel, unos 5 km al oeste de las antiguas minas de antimonio de Vilarbacú, donde Guillou (1969, fig. 5) había mencionado otro horizonte calcáreo similar. El punto estudiado por nosotros fue descubierto en 1994 por los geólogos Manuel Lombardero y Xosé Carlos Barros, y se sitúa en el paraje de Salgueiro, unos 800 m al sur de Folgoso do Courel (Lugo), estratigráficamente por encima del afloramiento más oriental de la Caliza de la Aquiana de los cartografiados en la Hoja de Monforte de Lemos por Iglesias Ponce de León y González Lodeiro (1981). Justo al este de la falla que pone en contacto la Cuarcita Armoricana con la Fm. Aquiana, la curva del camino muestra en serie invertida a esta última unidad, sucedida por unos 6 m de pizarras grises (las "pizarras intermedias" de Guillou) y un pequeño tramo cubierto de unos 3 m, en el que se advierten algunos bloques sueltos de arenisca que podrían corresponder a las "areniscas del Valentiense" de Guillou, aquí en posible contacto mecánico con las pizarras negras suprayacentes. La sección silúrica propiamente dicha debuta sobre el tramo cubierto con unos 3 m de pizarras negras, sucedidas por algo menos de 2 m de calizas negras en bancos centimétricos, con intercalaciones de pizarras negras, sobre los que se sitúa un espesor indeterminado de pizarras que progresivamente van aumentando su contenido en cloritoide. El trazado del camino se torna enseguida levemente oblicuo a la estratificación, lo que permite seguir la caliza silúrica durante una treintena de metros hacia el este, aunque tan sólo el afloramiento más occidental, en la curva del ca-

mino, es de donde proceden los conodontos estudiados (latitud= 42° 34' 52,4" N; longitud= 07° 11' 38,5" O).

Del tramo calcáreo se tomaron 13 muestras consecutivas para microfósiles, entre las cuales las muestras COU-2 y COU-3 libraron restos identificables de conodontos. Ambas se localizan hacia la parte media del banco principal de caliza, respectivamente a 0,9 m y 1 m por encima de la base de la primera de las intercalaciones calcáreas (medidas tomadas en flanco inverso).

Desde el punto de vista petrográfico, los niveles calizos estudiados son micritas matriz soportadas con texturas *mudstone-wackestone*, en la que los aloquímicos entran en una proporción aproximada al 5%. Los icrofósiles corresponden esencialmente a conchas de ortocerátidos bastante recrystalizadas, reconocibles tanto en sección como en los planos de estratificación (Fig. 1), a los que se unen fragmentos de graptolitos reducidos a rellenos calcíticos. Toda la roca está muy deformada, observándose procesos de presión-disolución y abundantes venillas de calcita de 1-2 mm de espesor.

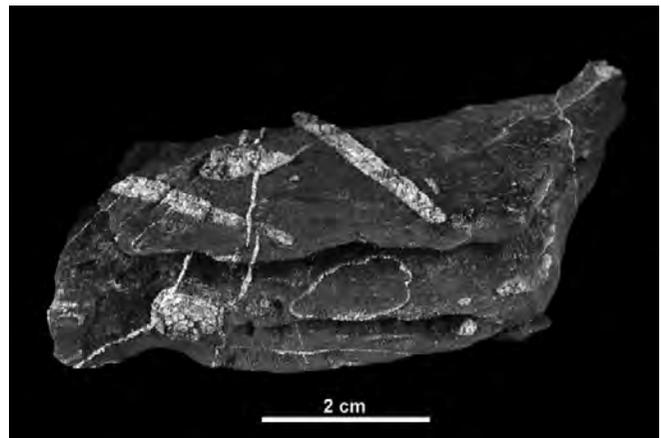


Figura 1. Conchas recrystalizadas de *Michelinoceras* sp. en el techo de uno de los bancos de la caliza de cefalópodos (nivel COU-4) de Folgoso do Courel.

Figure 1. Recrystallized shells of *Michelinoceras* sp., lying on top of one of the bedding planes (level COU-4) of the cephalopod limestone from Folgoso do Courel.

En relación al medio sedimentario, la abundancia de nautiloideos, la asociación de conodontos y el análisis de las microfácies permite inferir un ambiente de plataforma externa. Las pizarras negras que enmarcan a las calizas también están exentas de fósiles bentónicos, aunque algunos planos contienen restos de graptolitos indeterminables debido a la mala conservación.

Los conodontos identificados en las intercalaciones calcáreas estudiadas, establecen un rango de edad comprendido entre el Wenlock superior (Homeriense) y el Ludlow inferior (Gorstiense), tal y como se detalla en el apartado bioestratigráfico.

3. MATERIALES Y MÉTODOS

Durante los años 2001 y 2011 se realizaron sendas campañas al área de estudio con el propósito de recolectar muestras para la búsqueda de conodontos. Se muestrearon seis niveles de calizas que están intercalados en un tramo de pizarras negras de aproximadamente 6 m de espesor. De ellos se obtuvieron trece muestras con un peso total aproximado a los 20 kg; de cada una de ellas se procesaron entre 1 y 2 kg de roca para su investigación micropaleontológica. Una pequeña porción de las muestras se destinó a la confección de láminas delgadas para identificar las microfácies y determinar la posible existencia de otros grupos fósiles que se disuelven durante el tratamiento químico de la roca.

La extracción de los conodontos se realizó siguiendo la metodología clásica sugerida por Stone (1987). La colección micropaleontológica está constituida por un total de 73 ejemplares identificables taxonómicamente y un elevado número de fragmentos cuyos tamaños oscilan entre menos de 100 micras y 500 micras.

4. ASPECTOS TAFONÓMICOS

Como paso previo a la identificación taxonómica de los ejemplares recuperados, se realizó el análisis tafonómico de los elementos aislados, lo que de por sí es indicativo de la desarticulación de los aparatos conodontales en fase bioestratinómica; además de este mecanismo de alteración tafonómica se han reconocido otros procesos como: la *carbonificación* que se manifiesta por las tonalidades oscuras casi negras de los elementos y sus superficies opacas con textura sacarosa, algunos de los ejemplares están parcialmente grafitizados. El índice de alteración del color (CAI) corresponde a valores de 4,5-5 (Epstein *et al.*, 1977), no obstante no se asigna el intervalo térmico correspondiente a los colores observados, porque en las muestras

analizadas existe gran cantidad de materia orgánica que puede enmascarar el color de alteración inducido por la temperatura.

Los *encostramientos* son muy frecuentes y se presentan en forma de pequeños cristales de composición fosfática y también como recubrimientos arcillosos irregulares. Sobre la superficie de un gran número de elementos se observan restos de la matriz rocosa de la que fueron extraídos (Lám. 1, figs. 1-2, 4, 6, 12, 15-16; Lám. 2, figs. 3, 5, 8, 10-11; Lám. 3, figs. 15 y 16). Las cavidades basales de los conodontos se encuentran parcialmente rellenas por material procedente de la matriz rocosa calcárea y en menor medida por precipitados de fosfato de calcio (Lám. 1, figs. 15 y 17; Lám. 2, figs. 2, 6, 11 y 15; Lám. 3, figs. 1, 4, 8, 11, 14 y 23). Por otra parte, la *disolución* es un rasgo común en la mayoría de los ejemplares estudiados y afecta tanto a los márgenes de los elementos como a distintos sectores de su superficie (Lám. 1, figs. 6 y 14; Lám. 2, figs. 3, 8-9, 11 y 15; Lám. 3, figs. 2, 6, 13, 18-20, 22 y 27).

La mayoría de los ejemplares muestran evidencias de *distorsiones mecánicas*: se encuentran fragmentados y presentan fisuras que se disponen paralelas a la cúspide y, en menor medida, perpendiculares a la misma. Un elevado número de ejemplares están notoriamente deformados (Lám. 1, figs. 2, 4, 6, 8-9 y 12; Lám. 2, figs. 4, 7, 9, 11-13 y 15; Lám. 3, figs. 1-2, 5-7, 9-11, 13-16, 18, 20, 23-24 y 27).

En la colección de conodontos estudiada no se han encontrado evidencias tafonómicas de reelaboración y, cabe destacar que no están presentes todos los tipos morfológicos que constituyen las especies multielementales identificadas en este trabajo.

5. ESTUDIO SISTEMÁTICO

En este trabajo se ha adoptado la propuesta de Sweet (1988) tanto para las categorías supragenéricas, como para los términos que designan los distintos tipos morfológicos que integran el aparato conodontal.

La identificación de algunos taxones se ha realizado en nomenclatura abierta siguiendo los criterios expuestos por Matthews (1973) y Bengston (1988).

La colección de conodontos se encuentra depositada en el Departamento de Paleontología de la Facultad de Ciencias Geológicas de la Universidad Complutense de Madrid. Las siglas utilizadas son COU, abreviatura de Courel, seguidas por un dígito que indica el nivel de procedencia, y a continuación el número del ejemplar en la colección micropaleontológica.

Orden BELODELLIDA Sweet, 1988

Familia DAPSILODONTIDAE Sweet, 1988

Se incluyen en esta familia los géneros *Dapsilodus* Cooper, 1976 y *Besselodus* Aldridge, 1982. Los aparatos de las especies de ambos géneros están constituidos por elementos coniformes no geniculados, profundamente excavados y lateralmente comprimidos que exhiben una ornamentación oblicua, muy característica, que se sitúa adyacente al margen anterior de la base (Sweet, 1988).

Género *Dapsilodus* Cooper, 1976

Especie tipo: *Distacodus obliquicostatus* Branson y Mehl, 1933

Observaciones.- Cooper (1976) estableció el género *Dapsilodus* y basó la reconstrucción de su aparato multielemental, tomando como referencia la propuesta de Serpagli (1970) para el aparato de la especie *Acontiodus obliquicostatus* (Branson y Mehl). En este trabajo se sigue a Sweet (1988, p. 49-51) quien establece que el aparato posee cuatro elementos morfológicos diferentes (Sa, Sb-Sc y M), siendo el elemento M no geniculado. Esta última característica nos permite diferenciar este género de *Besselodus*.

Dapsilodus obliquicostatus (Branson y Mehl, 1933)

(Lám. 1, figs. 1-11)

- 1933 *Distacodus obliquicostatus* n. sp. - Branson y Mehl, Lám. 3, fig. 2.
- 1962 *Acodus inornatus* Ethington - Ethington y Furnish, p. 1259.
- 1970 *Acontiodus obliquicostatus* (Branson y Mehl) - Serpagli, p. 8-10, Lám. 23 figs. 1-10.
- 1970 *Acodus mutatus* (Branson y Mehl) - Serpagli, p.23, figs. 11-14.
- 1976 *Dapsilodus obliquicostatus* (Branson y Mehl) - Cooper, pl. 2, figs. 10-13, 18- 20.

- 1981 *Dapsilodus obliquicostatus* (Branson y Mehl) - Aldridge, Dorning y Siveter, Lám. 2.1, figs. 6-8.
- 1983 *Dapsilodus obliquicostatus* (Branson y Mehl) - Mohamed, Lám.16, figs. 12-15.
- 1984 *Dapsilodus obliquicostatus* (Branson y Mehl) - Aldridge y Jeppson, fig.-text. 3, a-c.
- 1987 *Dapsilodus obliquicostatus* (Branson y Mehl) - Over y Chatterton, Lám. 6, figs. 1-2.
- 1990 *Dapsilodus obliquicostatus* (Branson y Mehl) - Armstrong, p. 70-71, Lám. 7, figs. 7-12.
- 1991 *Dapsilodus obliquicostatus* (Branson y Mehl) - McCracken, p. 78-79, Lám. 4, figs. 11, 13-14, 16-28?, 30-32, 35,40.
- 1993 *Dapsilodus obliquicostatus* (Branson y Mehl) - Rodríguez Cañero, p. 220-221, Lám. 1, figs. 6-8.
- 1994 *Dapsilodus obliquicostatus* (Branson y Mehl) - Sarmiento, Méndez-Bedia, Aramburu, Arbizu y Truyols, Lám. 1, figs. 1-6.
- 1998 *Dapsilodus obliquicostatus* (Branson y Mehl) - Ferretti, Corradini y Serpagli, Lám. 2.2.1, fig. 13.
- non 1999 *Dapsilodus obliquicostatus* (Branson y Mehl) - Lehnert, Bergström, Benedetto y Vaccari, fig. 7, d; h; i.
- 2007 *Dapsilodus obliquicostatus* (Branson y Mehl) - Zhang y Barnes, fig. 9, 1-3.
- 2009 *Dapsilodus obliquicostatus* (Branson y Mehl) - Corriga, Corradini y Ferretti, Lám. 2, fig. 13.
- 2009 *Dapsilodus obliquicostatus* (Branson y Mehl) - Mestre, p. 474, 476, Fig. 3, 12-15.

Material estudiado.- Un elemento M proveniente del nivel COU-3 y nueve elementos Sb-Sc provenientes de los niveles COU- 2 y COU-3.

Descripción.- Se atribuye a la posición M (Lám. 1, fig. 2) un elemento asimétrico, caracterizado por una base alta y ancha, con una cúspide ligeramente reclinada, cuyo margen posterior forma con el margen superior de la base un ángulo cercano a los 90°. El ejemplar estudiado es idéntico al ilustrado por Armstrong (1990, Lám. 7, fig. 12).

Los elementos Sb-Sc son asimétricos y presentan una base muy desarrollada, ancha y alta. La cúspide es proclina y su margen posterior forma con el margen superior de la base un ángulo mayor a 90°. Desde la mitad superior de

la base se desarrolla sobre uno de los flancos una costilla que se prolonga en toda la longitud de la cúspide.

Los elementos Sb-Sc atribuidos a esta especie presentan en el margen anterior de la base estrías muy características que son paralelas entre sí. Todos los especímenes estudiados tienen como rasgo común sus bordes anterior y posterior agudos.

Distribución estratigráfica.- Esta especie se ha identificado en el intervalo Silúrico inferior-Devónico Inferior de Europa, Norteamérica, Norte de África y Australia.

Familia BELODELLIDAE Khodalevich y Tschernic, 1973

Se incluyen en esta familia: *Belodella* Ethington, 1959; *Coelocerodontus* Ethington, 1959; *Dvorakia* Klapper y Barrick, 1983; *Stolodus* Lindström, 1955 y *Walliserodus* Serpagli, 1967. Estos géneros tienen aparatos constituidos por cuatro o cinco tipos morfológicos. Se trata básicamente de elementos coniformes elongados, profundamente excavados, con superficies lisas u ornamentadas por costillas y con bordes o márgenes longitudinales que en algunas especies pueden presentar una denticulación incipiente (Sweet, 1988).

Género *Walliserodus* Serpagli, 1967

Especie tipo: *Paltodus debolti* Rexroad, 1967

***Walliserodus?* sp.**

(Lám. 1, fig. 12)

Material estudiado.- Un único ejemplar Sa obtenido en el nivel COU-2.

Descripción.- El elemento tiene una cúspide bien desarrollada, de erecta a ligeramente recurvada y una base muy larga y estrecha. Ambos flancos del elemento están recorridos por una costilla de posición mediana. La cavidad basal es amplia y se extiende hasta la base de la cúspide.

Observaciones.- El ejemplar estudiado presenta una conservación deficiente que impide su asignación específica.

Orden PRIONIODINIDA Sweet, 1988

Familia PRIONIODINIDAE Bassler, 1925

Pertenecen a esta familia 10 géneros, de ellos sólo se tratará el más antiguo: *Oulodus* Branson y Mehl, 1933. El aparato de su especie tipo presenta elementos extensifor-

mes *digyrate* en las posiciones Pa y Pb y un elemento *dolobrate* en la posición M (Sweet, 1988).

Género *Oulodus* Branson y Mehl, 1933

Especie tipo: *Cordylodus serratus* Stauffer, 1930

***Oulodus* cf. *siluricus* (Branson y Mehl, 1933)**

(Lám. 2, figs. 1-6)

cf. 1957 *Lonchodina greilingi* n. sp.- Walliser, p. 45, Lám. 3, fig. 26.

cf. 1957 *Prionodina* cf. *armata* (Hinde, 1879) - Walliser, p. 45, Lám. 2, figs. 21-22.

cf. 1957 *Trichonodella inconstans* n. sp. - Walliser, Lám. 3, figs. 16-17.

cf. 1995 *Oulodus siluricus* (Branson y Mehl, 1933) - Sanz López, p.450-451, Lám. 5. Figs. 11-14; Lám. 6, figs. 2, 4-10.

cf. 1999 *Oulodus siluricus* (Branson y Mehl, 1933) - Viira, Lám. 1, figs. 5-8.

Material estudiado.- 6 ejemplares procedentes del nivel COU-3.

Descripción.- Se atribuyen a esta especie 6 ejemplares cuya posición dentro del aparato multielemental es difícil de establecer. Estos elementos de tipo *bipennate* y *digyrate* se han asignado, con dudas, a las posiciones Pa, Pb y M. Todos ellos presentan como rasgos comunes: el espacio entre denticulos contiguos tiene forma de u, la sección transversal de los mismos, al igual que la de la cúspide, es redondeada, y exhiben una cavidad basal amplia, en ocasiones de gran desarrollo (Lám. 2, fig. 2), que recorre toda la superficie inferior del elemento.

Observaciones.- Esta especie se ha identificado en nomenclatura abierta porque no se han reconocido todos los elementos del aparato conodontal y por la conservación regular de los ejemplares atribuidos a la misma.

Distribución estratigráfica: se ha identificado esta especie en el Silúrico medio-superior a nivel mundial.

Orden OZARKODINIDA Dzik, 1976

Familia KOCKELELLIDAE Klapper, 1981 (*in* Sweet, 1988)

Se incluyen en esta familia los géneros: *Ancoradella* Walliser, 1964 y *Kockelella* Walliser, 1957, ambos del Silúrico.

Género *Kockelella* Walliser, 1957

Especie tipo: *Kockelella variabilis* Walliser, 1957

Observaciones.- El aparato multielemental fue propuesto por Walliser (1964) quien lo denominó "aparato G" y este mismo autor con posterioridad atribuyó a *Kockelella variabilis* (Walliser, 1972). Walliser (1957, 1964) reconoció como elementos Pa de *Kockelella* solo a las formas que tienen procesos laterales denticulados; años más tarde Barrick y Klapper (1976) incluyeron también como elementos Pa a formas que no habían desarrollado estos procesos laterales con dentículos.

En este trabajo se sigue la propuesta de Sweet (1988, p. 97-98), quien establece que el aparato de *Kockelella* está formado por seis elementos morfológicos diferentes que ocupan las posiciones: Pa, Pb, M, Sa, Sb y Sc.

***Kockelella* cf. *variabilis* (Walliser, 1957)**

(Lám. 1, figs. 13-17)

- cf.1957 *Kockelella variabilis* n. sp. - Walliser, p. 35-36, Lám. 1, fig. 3-10.
- cf.1957 *Ozarkodina ziegleri* n. sp. - Walliser, p. 41, Lám. 1, figs. 26-30.
- cf.1957 *Trichonodella inconstans* n. sp. . - Walliser, p. 50-51, Lám. 3, figs. 10-17.
- cf.1957 *Lonchodina greilingi* n. sp. - Walliser, p. 38-39, Lám. 3, figs. 20-26.
- cf.1957 *Ligonodina silúrica* Branson y Mehl - Walliser, p. 38, Lám. 2, figs. 10.
- cf.1964 *Neoprioniodus multiformis* n. sp. - Walliser, p. 50-51, Lám. 29, figs. 14; 17-25.
- cf.1970 *Kockelella variabilis* Walliser - Serpagli, p. 85, Lám. 22, fig. 13.
- cf.1972 *Kockelella variabilis* Walliser - Walliser, p. 77
- cf.1975 *Kockelella variabilis* Walliser - Aldridge, Lám. 1, fig. 19.
- cf.1975 *Kockelella variabilis* Walliser - Klapper y Murphy, p. 53-54, Lám. 9, fig. 5?, 6-9, 11; Lám. 10, figs. 1-7.
- cf.1976 *Kockelella variabilis* Walliser - Barrick y Klapper, p. 77-78, Lám. 3, figs. 12-17.
- cf.1993 *Kockelella variabilis* Walliser - Rodríguez Cañero, p. 312-313, Lám. 12, figs. 10-14.
- cf.1995 *Kockelella variabilis* Walliser - Sanz-López, p. 479-480, Lám. 4, figs. 3-7; Lám. 6, figs. 1; 3; Lám. 8, figs. 1-3.

cf.1998 *Kockelella variabilis* Walliser - Männik y Malkowski, Lám. 2, figs. 5, 8.

Material estudiado.- 2 elementos Sa, un elemento Sb y un elemento M, procedentes del nivel COU-2.

Descripción.- Los elementos *digyrate* que ocupan la posición Sa presentan simetría bilateral, una cúspide robusta y dos procesos laterales que forman un ángulo que varía entre los 85° y 110°. Los dentículos son grandes y de sección circular. La cavidad basal es prominente debajo de la cúspide y recorre los procesos como un surco estrecho.

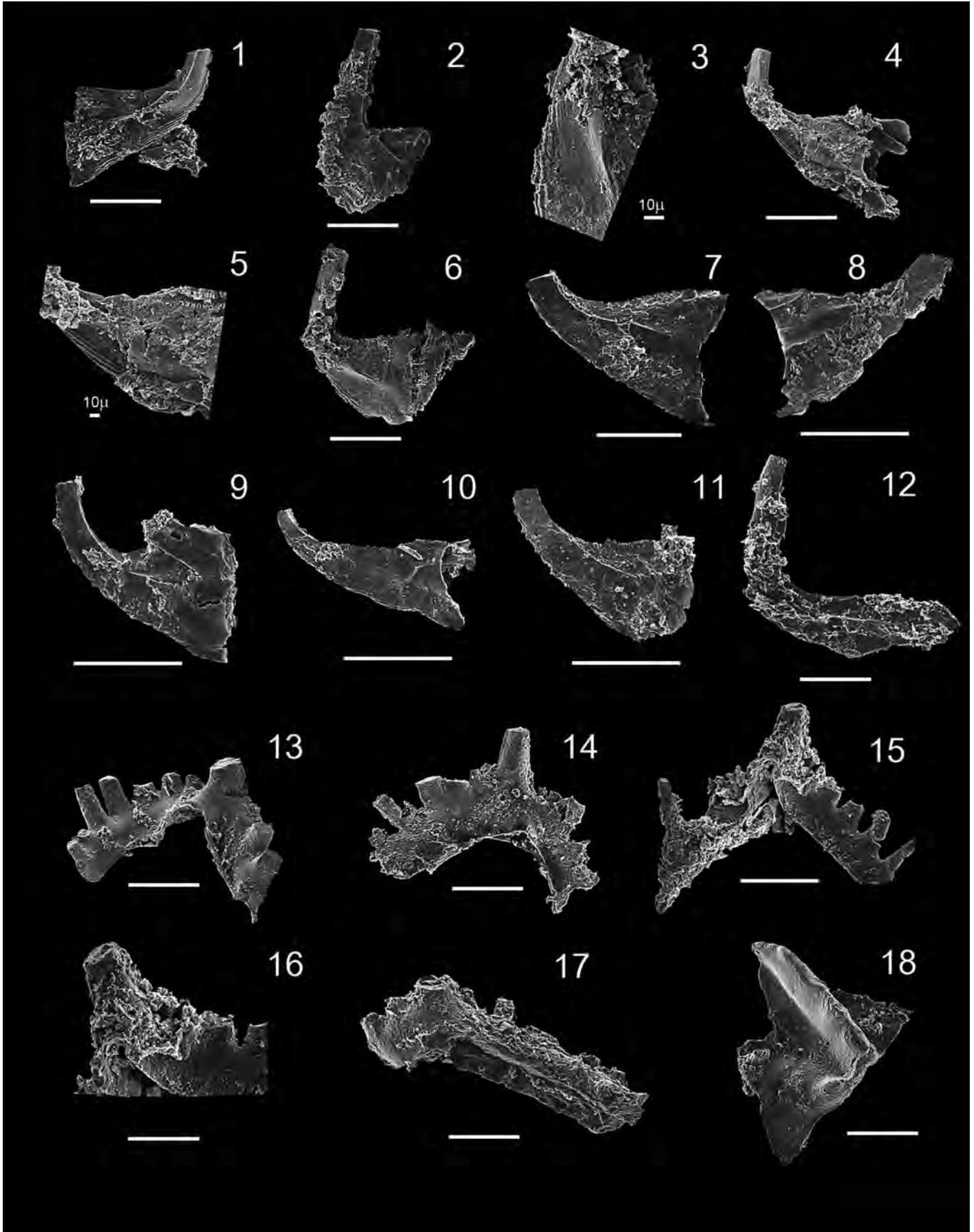
La posición Sb la ocupa un elemento *digyrate* con dentículos robustos, de sección transversal circular. La cavidad basal presenta las mismas características que en el elemento anterior.

El elemento M es también *digyrate* y se caracteriza por presentar un desarrollo desigual de los procesos laterales, y una cavidad basal muy amplia por debajo del proceso de mayor longitud.

Observaciones.- La identificación taxonómica de estos ejemplares se ha realizado en nomenclatura abierta por la ausencia de los elementos que ocupan las posiciones P

Lámina 1. 1-12, *Dapsilodus obliquicostatus* (Branson y Mehl, 1933). 1, Elemento S. Vista lateral, COU- 2- 3001. 2, Elemento M. Vista lateral, COU-2- 3005. 3, Detalle de las estrías cerca del margen anterior en el elemento (2), COU-2-3005. 4, Elemento S. Vista lateral oblicua, COU-3-3002. 5, Detalle de las estrías cerca del margen anterior en el elemento (4), COU-3-3002. 6-11, Elementos S. Vista lateral. 6, COU-3-3003; 7, COU-3-3007; 8, COU-3-3000; 9, COU-3-3009; 10, COU-3-3008; 11, COU-3-3004. 12, *Walliserodus?* sp. Elemento Sa. Vista lateral, COU-2-3010. 13-17, *Kockelella* cf. *variabilis* (Walliser, 1957). 13, Elemento Sa. Vista postero-lateral, COU-2-3021. 14, Elemento Sb. Vista posterior oblicua, COU-2-3023. 15, Elemento Sa. Vista posterior oblicua, COU-2-3025. 16, Encostramiento en la porción basal de la cúspide en el ejemplar (16), COU-2-3025. 17, Elemento M. Vista lateral interna, COU-2-3027. 18, *Kockelella?* sp. Elemento Pa. Vista superior oblicua, COU-2-3029. La escala gráfica representa 200 µm.

Plate 1. 1-11, *Dapsilodus obliquicostatus* (Branson y Mehl, 1933). 1, S element, lateral view, COU-2-3001. 2, M element, lateral view, COU-2-3005. 3, Detail of the striation near the anterior margin in the element (2), COU-2-3005. 4, S element, oblique lateral view, COU-3-3002. 5, Detail of the striation near the anterior margin in the element (4), COU-3-3002. 6-11, S elements lateral view; 6, COU-3-3003; 7, COU-3-3007; 8, COU-3-3000; 9, COU-3-3009; 10, COU-3-3008; 11, COU-3-3004. 12, *Walliserodus?* sp. Sa element, lateral view, COU-2-3010. 13-17, *Kockelella* cf. *variabilis* (Walliser, 1957). 13, Sa element, posterior lateral view, COU-2-3021. 14, Sb element, posterior oblique view, COU-2-3023. 15, Sa element, posterior oblique view, COU-2-3025. 16, Mineral crust lying on the basal part of the cusp, specimen (16), COU-2-3025. 17, M element, inner lateral view, COU-2-3027. 18, *Kockelella?* sp. Pa element, upper oblique view, COU-2-3029. Scale bar = 200 µm.



en el aparato y que son diagnósticos para la determinación de la especie.

Distribución estratigráfica.- Esta especie se ha reconocido en materiales del Ludlow a nivel mundial.

***Kockelella?* sp.**

(Lám. 1, fig. 18)

Material estudiado.- 1 elemento Pa *carminiscaphate* procedente del nivel COU-2.

Observaciones.- Se trata de un elemento fragmentado en el que solo se aprecia parte de la lámina y la porción de la plataforma vinculada a ella.

Familia SPATHOGNATHODONTIDAE Hass, 1959

En esta familia se incluyen 18 géneros de los que se tratará *Ozarkodina* Branson y Mehl, 1933 (Sweet, 1988) y el nuevo género *Wurmiella* Murphy, Valenzuela-Ríos y Carls, 2004. Se trata de grupo de conodontos con morfologías muy variables y con un amplio rango estratigráfico; sus aparatos son simples y están constituidos por seis o siete tipos morfológicos: los elementos pectiniformes *angulate* y *carminate* ocupan las posiciones Pa y Pb, un elemento *dolobrate* o *bipennate* la posición M, y una serie de transición simétrica que incluye elementos *alate*, *digyrate* y *bipennate* las posiciones Sa, Sb y Sc respectivamente (Sweet, 1988).

Género *Ozarkodina* Branson y Mehl, 1933

Especie tipo: *Ozarkodina typica* Branson y Mehl, 1933

Observaciones.- Lindström (1970) designó como *Ozarkodina* al aparato que Jeppson (1969) denominó *Hindeode-lla* y que había sido reconstruido con seis tipos diferentes de elementos. Por otra parte Nicoll (1985) a partir del análisis de diferentes "clusters", propuso que el aparato de *Ozarkodina* estaba integrado por siete tipos morfológicos diferentes representados por: 2 elementos M, 1 elemento Sa, 4 elementos Sc, 2 elementos Sb, 2 elementos Sd, 2 elementos Pb y 2 elementos Pa.

Murphy *et al.* (2004) realizaron una revisión de los spathognatodóntidos del intervalo Silúrico superior-Devónico inferior y subdividieron el género *Ozarkodina*, atribuyendo, entre otras cosas, la especie *Ozarkodina excavata*

al género *Wurmiella* Murphy, Valenzuela-Ríos y Carls, 2004. En este estudio se ha seguido la propuesta de estos autores para la especie antes mencionada, no obstante, para las otras especies se ha optado por escribir el nombre del género entre comillas.

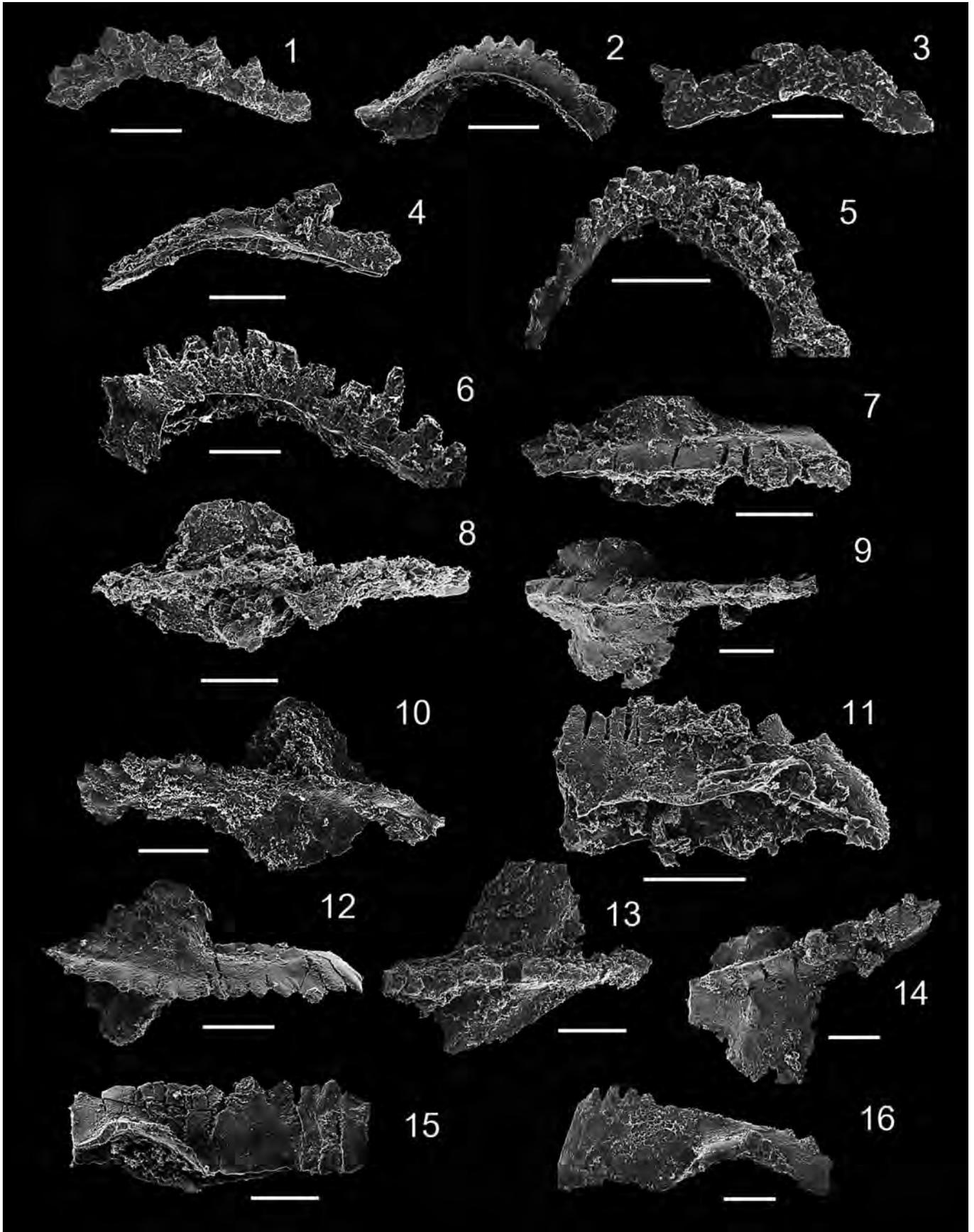
"*Ozarkodina*" *bohemica* (Walliser, 1964)

(Lám. 2, figs. 7-16; Lám. 3, figs. 1-6)

- 1964 *Spathognathodus sagitta bohemicus* n. subsp.- Walliser, Lám. 7, fig. 4; Lám. 18, figs. 23-24.
- 1964 *Ozarkodina edithae* n. sp.- Walliser, 1964, Lám. 26, figs. 15, 16.
- 1975 *Ozarkodina sagitta bohemica* (Walliser) - Aldridge, p. 327, Lám. 47, fig. 21.
- 1976 *Ozarkodina sagitta bohemica* (Walliser) - Barrick y Klapper, p.81, Lám. 4, figs. 1-7, 10-12
- 1981 *Ozarkodina sagitta bohemica* (Walliser) - Aldridge, Dorning y Siveter, Lám. 2.3, figs. 1,3.
- 1987 *Ozarkodina bohemica* (Walliser) - Over y Chatterton, Lám. 8, fig. 12.
- 1993 *Ozarkodina bohemica* (Walliser) - Kříž, Dufka, Jaeger y Sch nlaub, Lám. 1, figs. 1,3,4, 9-14.
- 1998 *Ozarkodina bohemica* (Walliser) - Ferretti, Corradini y Serpagli, Lám. 2.2.2, figs. 10-11.
- 2006 "*Ozarkodina*" *bohemica* (Walliser) - Sanz-López, Perret y Vachard, fig. 5, 1-4.
- 2009 *Ozarkodina bohemica* (Walliser) - Corrigan, Corradini y Ferretti, Lám. 2, fig.12.

Lámina 2. 1-6, *Oulodus* cf. *siluricus* (Branson y Mehl, 1933). 1, Elemento Pb. Vista lateral interna. COU-3-3032. 2, Elemento Pa. Vista lateral externa. COU-3033. 3, Elemento Pb. Vista lateral interna. COU-3-3036. 4, Elemento Pb. Vista lateral oblicua. COU-3-3038. 5, Elemento Pb. Vista lateral interna. COU-2-3031. 6, Elemento M. Vista posterior oblicua. COU-2-3035. 7-16, "*Ozarkodina*" *bohemica* (Walliser, 1964). 7-10, 12-14, Elementos Pa. Vista superior. 7, COU-2-3042; 8, COU-2-3044; 9, COU-2-3046; 10, COU-2-3045; 12, COU-2-3042; 13, COU-2-3043; 14, COU-2-3047. 11, 15-16, Elementos Pa. Vista lateral. 11, COU-2-3049; 15, COU-2-3040; 16, COU-2-3048. La escala gráfica representa 200 µm.

Plate 2. 1-6, *Oulodus* cf. *siluricus* (Branson y Mehl, 1933). 1, Pb element, inner lateral view, COU-3-3032. 2, Pa element, outer lateral view, COU-3-3033. 3, Pb element, inner lateral view, COU-3-3036. 4, Pb element, lateral oblique view, COU-3-3038. 5, Pb element, inner lateral view, COU-2-3031. 6, M element, posterior oblique view, COU-2-3035. 7-16, "*Ozarkodina*" *bohemica* (Walliser, 1964). 7-10, 12-14, Pa elements, upper views; 7, COU-2-3042; 8, COU-2-3044; 9, COU-2-3046; 10, COU-2-3045; 12, COU-2-3042; 13, COU-2-3043; 14, COU-2-3047. 11, 15-16, Pa elements, lateral views; 11, COU-2-3049; 15, COU-2-3040; 16, COU-2-3048. Scale bar = 200 µm.



Material estudiado.- 16 elementos Pa procedentes del nivel COU-2.

Descripción.- El elemento Pa es *carminiscaphate* y presenta una plataforma circular a subcircular bien desarrollada y una lámina alta cuya longitud es similar a la de la plataforma. Los dentículos son grandes, están comprimidos y su sección transversal es ovalada, además presentan sus extremos distales libres, excepto en la porción situada encima de la cavidad basal en donde se observa una tendencia a fusionarse. La cavidad basal es muy amplia y de forma circular a subcircular debajo de la plataforma, y continúa como un surco que se estrecha hacia el extremo distal de la lámina.

Observaciones.- Alguno de los elementos que se atribuyen en este trabajo a *O. bohémica* (Lám. 3, fig. 5 y 6) son similares a los descritos y figurados por Kříž *et al.* (1993, p. 829, Lám.1, figs. 9-14) como *Ozarkodina bohémica* morfotipo 2 y morfotipo 3. Estos autores propusieron tres morfotipos sucesivos en la región de Bohemia que según su opinión representan etapas evolutivas.

Distribución estratigráfica.- Esta especie define la biozona homónima que se extiende desde su primera aparición en el Homeriense (Wenlock medio) hasta el Gorstiense (Ludlow inferior), y ha sido reconocida en Europa, América y China.

“Ozarkodina” sagitta sagitta (Walliser, 1964)

(Lám. 3, figs.15-26)

- 1964 *Spathognathodus sagitta sagitta* n. subsp.- Walliser, Lám. 7, fig. 5, Lám. 18, figs. 7-11.
- 1964 *Neoprioniodus bicurvatooides* n. sp.- Walliser, Lám. 29, fig. 37.
- 1970 *Ozarkodina edithae* (Walliser) - Serpagli, p. 89, Lám. 22, figs. 3,4.
- 1972 *Ozarkodina sagitta* (Walliser) - Walliser, Lám. 77.
- 1975 *Ozarkodina sagitta sagitta* (Walliser) - Aldridge, Lám. 47, figs. 13-20.
- 1976 *Ozarkodina sagitta* (Walliser) - Barrick y Klapper, Lám. 80.
- 1981 *Ozarkodina sagitta sagitta* (Walliser) - Aldridge, Dorning y Siveter, Lám. 2.2, fig.5.
- 1993 *Ozarkodina sagitta sagitta* (Walliser) - Kříž, Dufka,

Jaeger y Sch nlaub, Lám. 1, fig. 6.

- 1993 *Ozarkodina sagitta sagitta* (Walliser) - Rodríguez Cañero, p. 297-299, Lám. 9, figs. 3-9.
- 1995 *Ozarkodina sagitta sagitta* (Walliser) - Simpson y Talent, p. 159, Lám. 10, fig. 12.
- 1998 *Ozarkodina sagitta sagitta* (Walliser) - Ferretti, Corradini y Serpagli, Lám. 2.2.2, fig. 7.
- 2009 *Ozarkodina sagitta sagitta* (Walliser) - Corrigan, Corradini y Ferretti, Lám. 2, fig.14.

Material estudiado.- 7 elementos Pa, 1 elemento Pb, y 2 elementos Sc, que se obtuvieron de los niveles COU-2 y COU-3.

Descripción.- El elemento *carminiscaphate* (Pa) presenta una lámina larga, recta y comprimida lateralmente que sustenta numerosos dentículos que disminuyen de tamaño hacia el extremo distal de la misma; la plataforma es de diseño triangular a subtriangular y la cavidad basal es amplia y ocupa aproximadamente la mitad de la longitud total del elemento.

El elemento *carminate* (Pb) tiene una cúspide prominente, un proceso anterior con 12 dentículos parcialmente fusionados que hacia el extremo del proceso se hacen erectos; la cúspide y los dentículos de este proceso muestran un perfil lateral en línea diagonal; el proceso posterior es más corto y posee 7 dentículos de similar o menor tamaño que los del proceso anterior. La cavidad basal es estrecha, se sitúa por debajo de la cúspide y se prolonga como surcos angostos que recorren ambos procesos sin alcanzar el extremo distal de los mismos.

El elemento Sc es *bipennate*, con procesos anterior y posterior comprimidos lateralmente; este último es muy largo y tiene numerosos dentículos que son reclinados y de sección transversal subredondeada a ovalada. La cavidad basal es pequeña y muestra un mayor desarrollo en coincidencia con el sector basal de la cúspide.

Distribución estratigráfica.- Esta especie define la biozona del mismo nombre que se extiende desde la primera aparición del taxón nominal en la parte más alta del Sheinwoodiense (Wenlock inferior) hasta la primera aparición de *Ozarkodina bohémica* con la que coexiste hasta el Gorstiense inferior (Ludlow inferior) en Europa y América.

Género *Wurmiella* Murphy, Valenzuela-Ríos y Carls, 2004

Especie tipo: *Ozarkodina excavata tuma* Murphy y Matti, 1983.

Observaciones.- Murphy *et al.* (2004, p. 8) en su revisión sobre los Spathognathodontidae, proponen *Wurmiella* y designan como especie tipo a *Ozarkodina tuma* para las formas que previamente se habían incluido en *Ozarkodina excavata*, estableciendo a su vez relaciones filogenéticas entre las especies del género *Ozarkodina* y las que estos autores incluyen en el género *Wurmiella*.

***Wurmiella excavata excavata* (Branson y Mehl, 1933)**
(Lám. 3, figs. 7-14)

- 1933 *Ozarkodina simplex* n. sp. - Branson y Mehl, p. 52, Lám. 3, figs. 46; 47.
- 1933 *Prioniodus excavatus* n. sp.- Branson y Mehl, p. 45, Lám. 3, figs. 7; 8.
- 1933 *Trichognathus excavata* n. sp.- Branson y Mehl, p.51, Lám. 3, figs. 35; 36.
- 1953 *Hindeodella equidentata* n. sp.- Rhodes, p. 303, Lám. 23, figs. 248, 252-254.
- 1953 *Plectospathodus extensus* n. sp.- Rhodes, p. 323, Lám. 23, figs. 286, 236-240.
- 1957 *Spathognathodus inclinatus* (Rhodes) - Walliser, p. 47-48, Lám. 1, figs. 16-20.
- 1957 *Ozarkodina media* n. sp.- Walliser, p. 40-41, Lám. 1, figs. 21-25.
- 1964 *Spathognathodus inclinatus inclinatus* (Rhodes) - Walliser, p. 76-77, Lám. 8, fig. 6; Lám. 19, figs.6-21; Lám. 20, fig. 18.
- 1964 *Ozarkodina media* Walliser - Walliser, p. 58-59, Lám. 8, fig. 5; Lám. 26, figs.19-34.
- 1964 *Neoprioniodus excavatus* (Branson y Mehl) - Walliser, p. 49-50, Lám. 8, fig. 4; Lám. 29, fig. 6.
- 1964 *Trichonodella excavata* (Branson y Mehl) - Walliser, Lám. 31, figs. 26-27.
- 1969 *Hindeodella excavata* (Branson y Mehl) - Jeppsson, fig. text. 1G-L, 3.
- 1970 *Hindeodella equidentata* (Rhodes) - Serpagli, Lám. 21, fig.1-2.
- 1970 *Trichonodella excavata* (Branson y Mehl) - Serpagli, Lám. 21, fig.6.
- 1975 *Ozarkodina excavata excavata* (Branson y Mehl) - Klapper y Murphy, p. 34-37, Lám. 6, figs. 1-20.
- 1976 *Ozarkodina excavata excavata* (Branson y Mehl) - Barrick y Klapper, p. 78, Lám. 4, figs. 13-23,26.
- 1979 *Ozarkodina excavata excavata* (Branson y Mehl) - Lane y Ormiston, Lám. 2, figs. 30,31; Lám. 9, figs. 18-23.
- 1982 *Ozarkodina excavata excavata* (Branson y Mehl) - Nowlan, fig. 3, N,Q,U.
- 1985 *Ozarkodina excavata excavata* (Branson y Mehl) - Savage, p.722, fig. 14, A-L.
- 1985 *Ozarkodina excavata excavata* (Branson y Mehl) - Mastandrea, Lám. 1, figs. 11; 14; 15; 17; 19-21.
- 1986 *Ozarkodina excavata excavata* (Branson y Mehl) - Bischoff, p. 134-137, Lám. 25, figs. 35-40; Lám. 26, figs. 1-40.
- 1987 *Ozarkodina excavata excavata* (Branson y Mehl) - Over y Chatterton, Lám. 7, fig. 18-19.
- 1990 *Ozarkodina excavata excavata* (Branson y Mehl) - Uyeno, p. 91-92, Lám. 3, fig. 11-12; Lám.8, figs. 7, 10, 11; Lám. 14, figs.11, 14-19, 23-25,31; Lám. 15; figs. 15-18.
- 1993 *Ozarkodina excavata excavata* (Branson y Mehl) - Rodríguez Cañero, p. 288-292, Lám. 10, figs. 1 - 11.
- 1995 *Ozarkodina excavata excavata* (Branson y Mehl) - Sanz López, p. 465, Lám. 5, figs. 1-7; Lám. 7, figs. 1-6.
- 1995 *Ozarkodina excavata excavata* (Branson y Mehl) - Simpson y Talent, p. 147-153, Lám. 8, figs. 16-25; Lám. 9, figs. 1-24.
- 1998 *Ozarkodina excavata excavata* (Branson y Mehl) - Ferretti, Corradini y Serpagli, Lám. 2.2.1, fig. 1.
- 1999 *Ozarkodina excavata excavata* (Branson y Mehl) - Cockle, p. 120, Lám. 3, figs. 1-14.
- 2001b *Ozarkodina excavata excavata* (Branson y Mehl) - Gutiérrez-Marco, Sarmiento, Robardet, Rábano y Van k, Lám. 1, figs. 1-8.
- 2004 *Wurmiella excavata excavata* (Murphy, Valenzuela-Ríos y Carls), p. 8-12.
- 2009 *Wurmiella* cf. *excavata* (Murphy, Valenzuela-Ríos y Carls) - Mestre, p. 474, Fig. 3, 4-6.

Material estudiado.- 1 elemento Pa, 2 elementos Pb, 1 elemento M, 1 elemento Sa, 1 elemento Sb y 2 elementos Sc, procedentes de los niveles COU-2 y COU-3.

Descripción.- El elemento Pa, de tipo *carminate*, en vista lateral presenta un margen aboral ligeramente arqueado y

dentículos anchos, comprimidos lateralmente y de tamaño similar. En la lámina es característica la presencia de un ensanchamiento longitudinal que se sitúa en la región basal de los dentículos y que la recorre de un extremo al otro. La cavidad basal es reducida y asimétrica debido a la presencia de una expansión lateral que alcanza mayor desarrollo en uno de los flancos.

Los elementos Pb, *bipennates*, tienen una cúspide prominente y procesos anterior y posterior describiendo un arco basal con un ángulo de más de 100°; el proceso anterior, de mayor desarrollo que el posterior, presenta dentículos anchos, altos y de sección transversal ovalada a subovalada, en tanto que en el proceso posterior los dentículos son más pequeños.

El elemento M, *digyrate*, tiene un proceso posterior largo, ligeramente arqueado y con numerosos dentículos. El proceso anterior es muy breve y se sitúa en un plano distinto al que forman la cúspide y el proceso posterior. La cúspide es prominente y con sección transversal subredondeada. La cavidad basal es profunda por debajo de la cúspide y se extiende como un surco bien desarrollado a lo largo de todo el proceso posterior

El elemento Sa, *bipennate*, describe un arco simétrico, en el que se puede diferenciar una cúspide de sección transversal ovalada y dos procesos laterales de similar longitud. El rasgo más destacado de este elemento es la presencia de una cavidad basal muy prominente, que es continuación del borde posterior de la cúspide.

El elemento Sb, *digyrate*, es muy similar al Sa pero en este caso ambos procesos no son simétricos y la cavidad basal es menos profunda que en el elemento Sa.

Los elementos Sc, *bipennate*, tiene dos procesos: anterior y posterior, de desigual desarrollo y en vista lateral ambos procesos pueden situarse sobre un mismo plano o el anterior estar ligeramente curvado hacia el flanco interno. Los dentículos aumentan de tamaño hacia los extremos distales de ambos procesos.

Distribución estratigráfica.- Esta especie tiene un amplio rango de distribución estratigráfica que se extiende desde el Silúrico inferior hasta el Devónico Inferior.

***Ozarkodina?* sp.**

(Lám. 3, fig. 27)

Material estudiado.- 13 elementos que ocuparían distintas posiciones en el aparato conodontal y que provienen de los niveles COU-2 y COU-3.

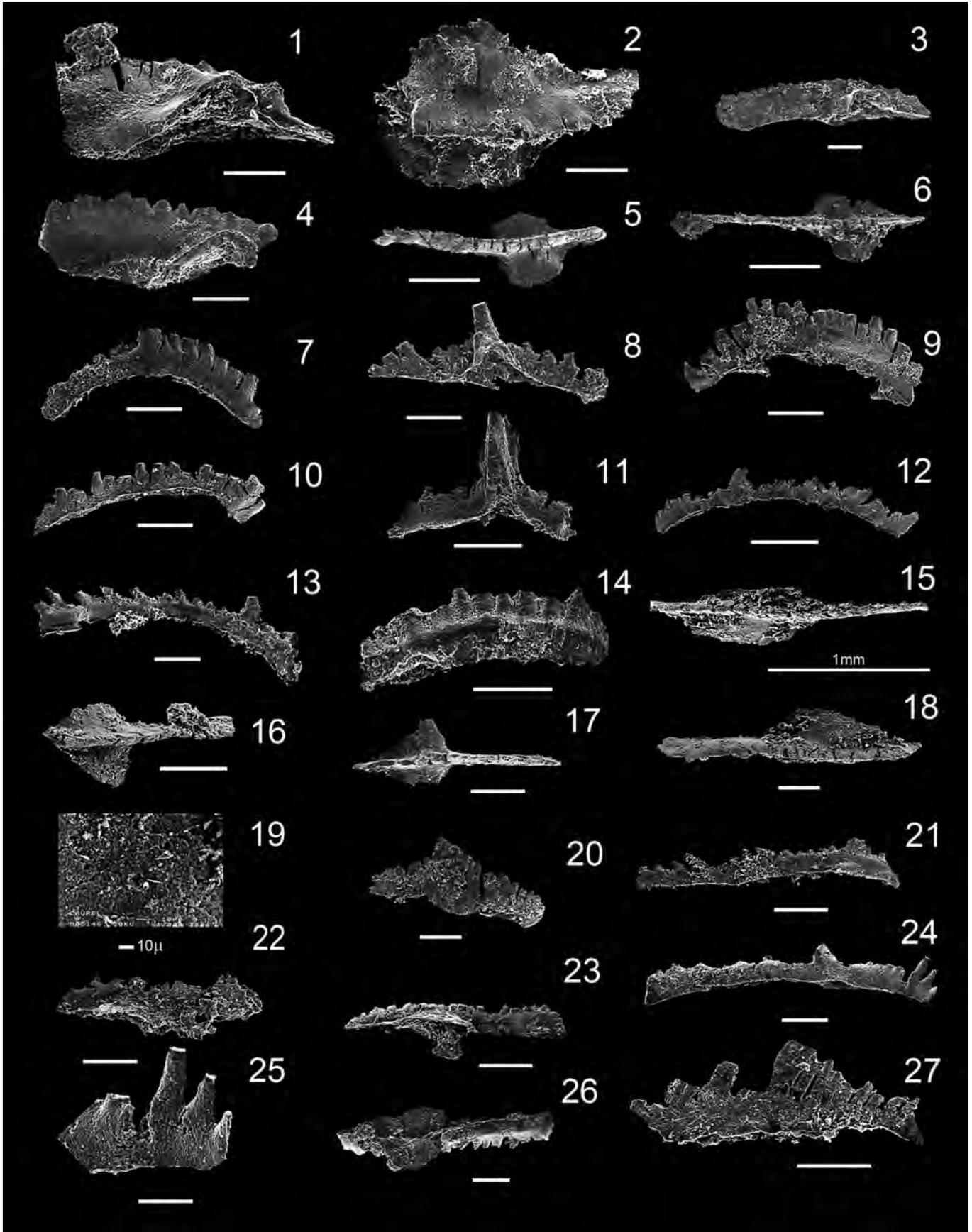
Observaciones.- En la colección estudiada existen numerosos ejemplares que por su deficiente estado de conservación (disolución, encostramiento, deformación, etc.), no pudieron ser atribuidos con certeza a este género.

6. BIOESTRATIGRAFÍA

La primera biozonación en base a conodontos silúricos fue propuesta por Walliser (1964), quien elaboró un es-

Lámina 3. 1-6, "*Ozarkodina*" *bohemica* (Walliser, 1964). 1, 3, 4, Elementos Pa. Vista lateral. 1, COU-2-3058; 3, COU-2-3051; 4, COU-2-3056. 2, Elemento Pa. Vista superior. COU-2-3053. 5-6, Elemento Pa. Vista superior. 5, COU-2-3055; 6, COU-2-3059. 7-14, *Wurmiella excavata excavata* (Branson y Mehl, 1933). 7, Elemento Pb. Vista lateral. COU-3-3062. 8, Elemento Sb. Vista posterior. COU-3-3067. 9, Elemento Pb. Vista lateral. COU-2-3065. 10, Elemento M. Vista lateral. COU-2-3061. 11, Elemento Sa. Vista posterior. COU-2-3069. 12-13, Elementos Sc. Vista lateral. 12, COU-2-3064; 13, COU-2-3068. 14, Elemento Pa. Vista lateral. COU-2-3060. 15-26, "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964). 15-18, Elementos Pa. Vista superior. 15, COU-3-3079; 16, COU-3-3071; 17, COU-3-3074; 18, COU-3-3075. 19, Superficies con evidencia de disolución en el elemento (18). COU-3-3075. 20, Elemento Pb. Vista lateral. COU-2-3077. 21, 24, Elementos Sc. Vista lateral. 21, COU-3-3076; 24, COU-3-3078. 22, 23, Elementos Pa. Vista lateral. 22, COU-2-3072; 23, COU-2-3070. 25, Superficies con evidencia de disolución en el elemento (24). COU-3-3078. 26, Elemento Pa. Vista inferior oblicua. COU-3-3073. 27, *Ozarkodina?* sp. Elemento S?. Vista lateral. COU-3-3086. La escala gráfica representa 200 µm.

Plate 3. Figures 1-6, "*Ozarkodina*" *bohemica* (Walliser, 1964). 1, 3-4, Pa elements, lateral views; 1, COU-2-3058; 3, COU-2-3051; 4, COU-2-356. 2, Pa element; upper view, COU-2-3053. 5-6, Pa elements, upper views; 5, COU-2-3055; 6, COU-2-3059. 7-14, *Wurmiella excavata excavata* (Branson y Mehl, 1933). 7, Pb element; lateral view, COU-3-3062. 8, Sb element; posterior view, COU-3-3067. 9, Pb element; lateral view, COU-2-3065. 10, M element; lateral view, COU-2-3061. 11, Sa element; posterior view, COU-2-3069. 12-13, Sc elements; lateral views, 12, COU-2-3064; 13, COU-2-3068. 14, Pa element; lateral view, COU-2-3060. 15-26, "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964). 15-18, Pa elements, upper views; 15, COU-3-3079; 16, COU-3-3071; 17, COU-3-3074; 18, COU-3-3075. 19, Dissolution on the surface of the element (18), COU-3-3075. 20, Pb element; lateral view, COU-2-3077. 21, 24, Sc elements, lateral views; 21, COU-3-3076; 24, COU-3-3078. 22, 23, Pa elements, lateral views; 22, COU-2-3072; 23, COU-2-3070. 25, Dissolution on the surface of the element (24), COU-3-3078. 26, Pa element; lower oblique view, COU-3-3073. 27, *Ozarkodina?* sp. S? element; lateral view, COU-3-3086. Scale bar = 200 µm.



quema bioestratigráfico constituido por 12 zonas sucesivas a partir de los hallazgos de la sección de Cellon en los Alpes Cárnicos (Austria). Aunque muchas de estas zonas han sido ampliamente reconocidas, su aplicación resulta complicada debido a que en Cellon el Llandovery no está completamente expuesto, lo que ha llevado al desarrollo de distintas biozonaciones locales. En este trabajo se sigue el esquema bioestratigráfico establecido por Corradini y Serpagli (1999) para Cerdeña, ya que en éste concurren las propuestas de diferentes autores para Europa, así como otras de carácter global.

En la sección silúrica del pliegue acostado del Courel, se han identificado las siguientes especies, provenientes de los niveles COU-2 y COU-3: *Dapsilodus obliquicostatus* (Branson y Mehl, 1933), *Oulodus* cf. *siluricus* (Branson y Mehl, 1933), *Kockelella* cf. *variabilis* (Walliser, 1957), "*Ozarkodina*" *bohemica* (Walliser, 1964), "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964) y *Wurmiella excavata excavata* (Branson y Mehl, 1933). La distribución de estos taxones en relación a la biozonación de Corradini y Serpagli (1999) se puede observar en la figura 2.

La especie que domina la asociación estudiada es *Dapsilodus obliquicostatus* (Branson y Mehl, 1933), cuya distribución vertical se extiende desde el Silúrico inferior al Devónico Inferior en Europa, Norteamérica, Norte de África y Australia. Barrick (1977) señala que la extensión de esta especie abarca desde la Zona de *Kockelella ranuliformis* a la Zona de *Kockelella variabilis* (Wenlock-Ludlow); por su parte McCracken (1991) describe este taxón en materiales del Llandovery inferior (Telychiense) al Wenlock superior-Ludlow inferior (Homeriense-Sheinwoodiense). Más tarde Rodríguez Cañero (1993) menciona su hallazgo en el Complejo Maláguide (Cordillera Bética) desde la Zona de *Ozarkodina sagitta* hasta la Zona de *Kockelella crassa* (Wenlock medio-Ludlow inferior); Sarmiento *et al.* (1994) lo describen en las zonas de *Aspelundia petila* y *A. fluegeli* (Llandovery) del norte de España; Ferretti *et al.* (1998) identifican el taxón en la Zona de *K. crassa* (Ludlow inferior) en Cerdeña, y Corradini y Serpagli (1999) lo mencionan desde la Zona de *K. ranuliformis* hasta la Zona de *Ozarkodina reimscheidensis* (Wenlock-Prídoli).

Del género *Ozarkodina* "sensu lato" se han reconocido

dos especies: "*Ozarkodina*" *bohemica* (Walliser, 1964) y "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964). La primera de ellas define la biozona homónima (FAD) en el intervalo Wenlock medio-Ludlow inferior de diversas áreas de Europa, América y China. Walliser (1964) estableció el registro más antiguo de esta especie en el Wenlock superior y el más moderno en el Ludlow inferior, ambos para el área de Bohemia. Barrick y Klapper (1976) plantearon su rango de extensión en la Formación Clarita de Oklahoma, desde la parte más alta de la Zona de *Kockelella amsdeni* hasta un nivel estratigráfico que corresponde a la Zona de *Polygnathoides siluricus* (Ludlow medio) en la parte inferior de la Formación Henryhouse. En Cerdeña la presencia de *O. bohemica* se registra en el intervalo de la Zona de *O. bohemica* a la Zona de *K. crassa* (Corriga *et al.*, 2009).

"*Ozarkodina*" *sagitta sagitta* (Walliser, 1964) define la biozona homónima (FAD) para el intervalo Wenlock inferior-Ludlow inferior y ha sido descrita en Europa y América. Sweet (1988) ha planteado su extensión desde la Zona de *O. sagitta* hasta la Zona de *K. crassa*. Rodríguez Cañero (1993) reconoció este taxón en el Complejo Maláguide y lo atribuyó al Wenlock superior. En Cerdeña su presencia está restringida a la Zona "*Ozarkodina*" *sagitta-sagitta* (Wenlock medio: Corriga *et al.*, 2009), en tanto que en el sureste de Australia Simpson y Talent (1995) describen la especie en el intervalo de las zonas de "*O.*" *sagitta* y *K. crassa*.

Wurmiella excavata excavata (Branson y Mehl, 1933) tiene un amplio rango de distribución estratigráfica, que se extiende desde el Silúrico inferior hasta el Devónico Inferior. Walliser (1964) adscribe la especie a la Zona de *K. patula* (Wenlock inferior); Barrick y Klapper (1976) extienden sus hallazgos hasta la Zona de *Kockelella ranuliformis*, ambas del Wenlock inferior. Rodríguez-Cañero (1993) registra su existencia en las secciones del Complejo Maláguide que corresponden al Devónico Inferior.

Según Chlupá *et al.* (1980) *Oulodus siluricus* (Branson y Mehl) se reconoce desde la Zona de *Polygnathoides siluricus* hasta la Zona de *Ozarkodina eosteinhornensis* (Ludlow medio-Prídoli superior); sin embargo, y de acuerdo con la síntesis elaborada por Corradini y Serpagli (1999), esta especie se registra desde la Zona de *O. sagitta* hasta la Zona de *Oulodus elegans detortus* (Wenlock medio -

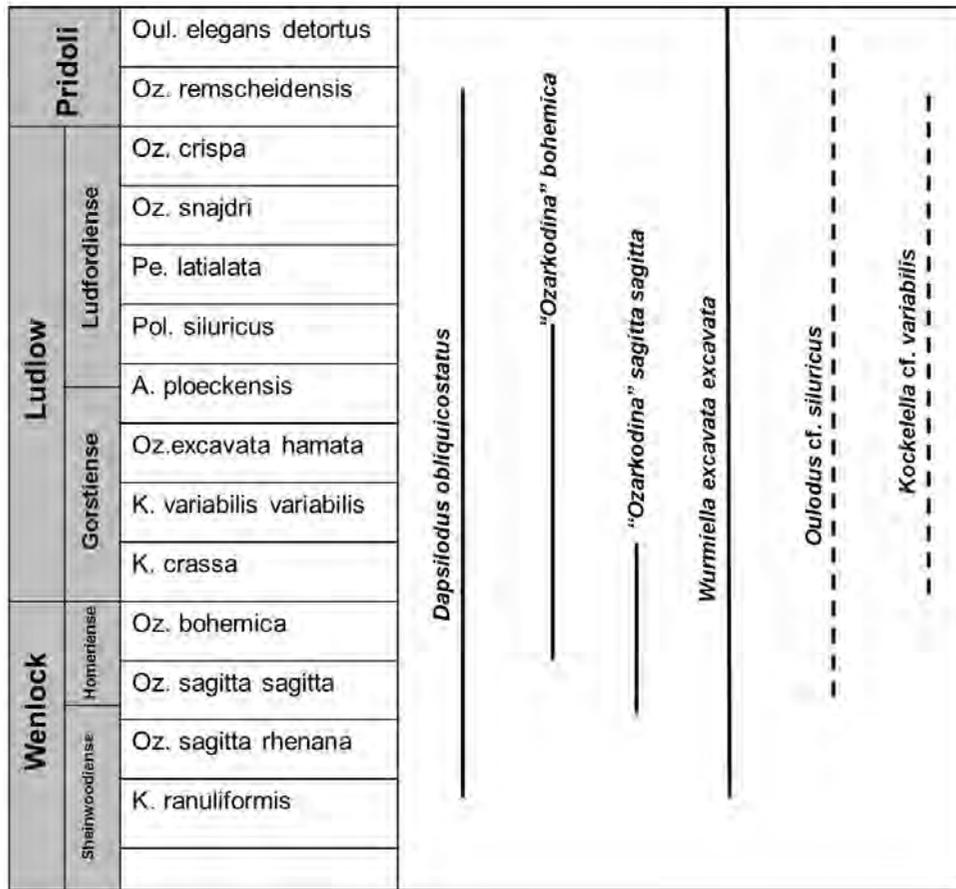


Figura 2. Distribución bioestratigráfica de algunas especies de conodontos identificadas en este trabajo, según datos de Corradini y Serpagli (1999). En línea discontinua se indican las formas en nomenclatura abierta.

Figure 2. Biostratigraphic distribution of some of the conodont species identified in this paper, according to the scheme of Corradini and Serpagli (1999). Taxa in open nomenclature are indicated with a dashed line.

Prídoli superior). Por otra parte, la especie *Kockelella variabilis* (Walliser), que define la biozona de intervalo homónima, se ha reconocido en materiales del Ludlow a nivel mundial. Walliser (1964) la describe en el intervalo correspondiente a la Zona de *K. crassa*-Zona de *P. siluricus*. Sweet (1988) propone su extensión desde la Zona de *K. crassa* hasta la Zona de *Pedavis latialata* (Ludlow inferior-superior). Kleffner (1990) sitúa el hallazgo más moderno de la especie en la parte basal de la Zona de *O. eosteinhornensis*. Por su parte, Corradini y Serpagli (1999) mencionan el registro de esta especie en el intervalo que corresponde a la parte media de la Zona *K. crassa*, hasta la parte superior de la Zona de *Ancoradella ploeckensis* (Ludlow inferior-parte baja del Ludlow superior).

En la asociación estudiada la coexistencia de "Ozarkodina" bohémica y "Ozarkodina" sagitta sagitta, junto con los elementos atribuidos a *Kockelella cf. Variabilis*, permite proponer para los niveles COU-2 y COU-3 una edad comprendida entre la parte alta de la Zona de "O". s. sagitta y la parte basal de la Zona de *K. variabilis*, que corresponde al intervalo Wenlock superior (Homeriense) - Ludlow inferior (Gorstiense).

Al comparar esta asociación de conodontos con la descrita por Gutiérrez-Marco *et al.* (2001b) en la localidad de La Barosa, podemos deducir que aunque ambas comparten algunas especies (*W. excavata* y *K. variabilis*?) esta última es más moderna por la presencia de "Ozarkodina" confluens (Ludlow), y podría representar un ambiente de

sedimentación más distal (Gutiérrez-Marco *et al.*, 2001, p. 250).

7. OBSERVACIONES PALEOECOLÓGICAS

En la asociación de conodontos del área del Courel predominan los elementos coniformes de la especie *Dapsilodus obliquicostatus* que, de acuerdo con Aldridge *et al.* (1981, p. 23) es una forma característica de zonas de plataforma externa. En el mismo sentido, se han identificado numerosos elementos de *Wurmiella excavata* que podrían adscribirse a la Asociación de *Wurmiella excavata* de Aldridge y Jeppsson (1999), típica de ambientes de plataforma marina abierta para el intervalo Wenlock inferior a Ludlow superior.

8. CONCLUSIONES

Se describe, por vez primera para el Silúrico del Sinclinal acostado del Courel, una asociación de conodontos integrada por los taxones *Dapsilodus obliquicostatus* (Branson y Mehl, 1933), *Oulodus cf. siluricus* (Branson y Mehl, 1933), *Kockelella cf. variabilis* (Walliser, 1957), "*Ozarkodina*" *bohemica* (Walliser, 1964), "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964), *Wurmiella excavata excavata* (Branson y Mehl, 1933), *Walliserodus sp.*, *Kockelella?* *sp.* y *Ozarkodina?* *sp.*

La presencia de las especies "*Ozarkodina*" *bohemica* (Walliser, 1964), "*Ozarkodina*" *sagitta sagitta* (Walliser, 1964) y *Kockelella cf. variabilis* permite atribuir los niveles investigados al intervalo comprendido entre la parte superior de la Zona de "*O.*" *sagitta sagitta* y la parte inferior de la Zona de *K. variabilis*, cuyo rango se extiende desde el Wenlock superior (Homeriense) hasta el Ludlow inferior (Gorstiense).

Desde el punto de vista estratigráfico, resulta llamativa la presencia de niveles calcáreos intercalados en las pizarras negras con cloritoide de la Formación Folgoso a lo largo del Sinclinal del Courel-Peñalba, en una distancia de 8 km lineales dentro de la Sierra do Courel (sur de Folgoso, norte de Vilarbacú, norte de A Seara), más la localidad de La Barosa ya en la provincia de León, unos 40 km

al oeste-suroeste del área anterior. Sin embargo los datos preliminares, derivados de la bioestratigrafía de conodontos, indican que el lentejón calizo de La Barosa es algo más moderno (Ludlow medio) que el registrado en la Sierra do Courel (Wenlock-Ludlow inferior).

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A new benthic foraminifer from the Recent coastal lagoon of Torreblanca (Castellón, Western Mediterranean)

Jordi Guillem and Juan Usera

Departament de Geologia, Universitat de València Estudi General (UVEG), Av. Dr. Moliner, 50, 46100 Burjassot, Valencia (Spain)
Jorge.Guillem@uv.es, juan.usera@uv.es

RESUMEN

Se describe una nueva especie de foraminífero bentónico *Paraphysalidia paralica* gen. et sp. nov., localizada en la laguna costera de Torreblanca (Este de España) e incluida en un nuevo género *Paraphysalidia*, asignado aquí provisionalmente a la familia Rotaliellidae. Las principales características de la nueva especie son su pequeño tamaño, su apertura interiomarginal bordeada por un pequeño labio y sus tres cámaras subglobulares que incluyen un prolóculo de gran tamaño dispuesto frente a las otras dos cámaras y separado de ellas por un estrechamiento. Además de en Torreblanca, *P. paralica* ha sido localizada en varios marjales y lagunas parállicas de la costa oriental de la Península Ibérica y también en ambientes de cuevas cársticas costeras en Yucatán (Mexico) y en las Bermudas, siempre en sedimentos cuaternarios o actuales. Adicionalmente, se ha podido detectar también en el yacimiento lacustre de mamíferos fósiles de Venta del Moro, lo que extiende la base de su rango estratigráfico al Mioceno Superior. Debido a su pequeño tamaño, *P. paralica* ha sido posiblemente pasada por alto o confundida con formas juveniles o gamónicas en otros estudios. Su distribución geográfica es por ello probablemente más amplia que la referida aquí, con lo que esta especie podría ser localizada en futuros trabajos en otras localidades de ambiente parálico o lacustre.

Palabras clave: Foraminíferos, Taxonomía, Ambientes parállicos, Península Ibérica, Cuaternario, Actual.

ABSTRACT

Paraphysalidia paralica gen. et sp. nov., a new tiny species of benthic foraminifera found in a coastal lagoon in Torreblanca (eastern Spain) is described here. It has been included in a new genus, *Paraphysalidia*, tentatively assigned to the Family Rotaliellidae. The main characteristics of the new species are its small size, its interiomarginal aperture bordered by a small lip and its three subglobular chambers, including a large proloculus arranged in opposition to the remaining chambers and separated from them by a constriction. In addition to Torreblanca, *P. paralica* has been found in several other paralic marshes and lagoons on the eastern Spanish coast and coastal karstic caves environments in Yucatan (Mexico) and Bermuda, all of them Recent or Quaternary in age. It has also been located in the (lacustrine, non-marine) mammalian site of Venta del Moro, which extends the base of its stratigraphic range to the Upper Miocene. Due to its small size, *P. paralica* has possibly been overlooked or misinterpreted as a juvenile/gamontic form in other studies. Its geographic distribution is thus probably wider and, as a result, it might be found in future surveys of other paralic or lacustrine environments.

Keywords: Foraminifera, Taxonomy, Paralic environments, Iberian Peninsula, Quaternary, Recent.

1. INTRODUCTION

Foraminifers from non-marine or marginal marine environments have now been reported worldwide: compiled data can be found, among others, in Murray (1973, 1991, 2006), Boltovskoy and Wright (1976), Patterson *et al.* (1990), Sen Gupta (1999a) or Wennrich *et al.* (2007). In the context of the Iberian Peninsula, Anadón (1989) described fossil foraminiferal faunas in several Cenozoic inland lacustrine basins. Moreover, fossil (Quaternary) and living foraminifera from numerous paralic environments have been found in the Spanish Mediterranean coasts (data for the Valencia region have been collected in Usera and Mateu, 1995; Usera *et al.*, 2000, 2002, 2003 and Usera, 2003).

One of those paralic environments, the Torreblanca marsh, lies between the localities of Torreblanca and Cabanes in the Valencian Community (eastern Spain), close to the Mediterranean coast (Fig. 1). It constitutes the final infilling phase of one of the once very abundant coastal lagoons on the Valencia coasts, which developed during the Holocene, after the Flandrian transgression, around 6000 BP (Rosselló, 1979; Segura *et al.*, 1995, 1997; Sanjaume *et al.*, 1996). The recent extraction of peat for its agricultural use as a fertilizer has artificially reversed the infilling process, thus partially recovering the previous coastal lagoon conditions (Auernheimer *et al.*, 1990).

Foraminifera in the Torreblanca marsh were first noticed by Colom (1959), who reported the occurrence of *Trichohyalus aguayoi* (Bermúdez) in peat deposits. Collado and Robles (1983) found fossil tests of *T. aguayoi* and *Ammonia beccarii tepida* (Cushman) again in Holocene deposits, whereas Usera *et al.* (1990) described an assemblage of up to 12 species of foraminifera currently living in the different pools created by the peat extraction activities in the Torreblanca marsh, with *T. aguayoi* and *A. beccarii tepida* as the dominant taxa. Additional studies concerning the Torreblanca foraminifera include those of Usera *et al.* (1996) and Guillem and Usera (1999, 2004). A further taphonomic, taxonomic and ecological study of the foraminiferal fauna from nine sampling stations in the marsh (Guillem, 2008) yielded up to 18 species, mainly brackish water taxa like *A. beccarii tepida*, *T. aguayoi*, *Trochammina inflata* (Montagu), *Jadammina macrescens* (Brady) or *Miliammina fusca* (Brady), together with some



Figure 1. Map of the Iberian Peninsula and its eastern coast to show the type locality (Torreblanca) as well as the other locations where *Paraphysalidia paralica* has been found so far.

others, less typical or not previously described in those environments, such as *Disconorbis bulbosus* (Parker).

Among them was a small, unidentified, but relatively abundant species, already remarked in the same locality by Guillem and Usera (1999), which had been left in open nomenclature as *aff. Physalidia* sp. The purpose of this note is to formally describe and name this new taxon. Original descriptions together with figures and type descriptions of the Ellis and Messina Catalogue of Foraminifera (Ellis and Messina, 1940 et seq.) have been followed, as well as the generic and suprageneric classification of Loeblich and Tappan (1987), including later modifications by Loeblich and Tappan (1992) and Sen Gupta (1999b).

2. SYSTEMATIC DESCRIPTIONS

Order ROTALIIDA Lankester, 1885

Superfamily DISCORBACEA Ehrenberg, 1838

Family ROTALIELLIDAE Loeblich and Tappan, 1964

Genus *Paraphysalidia* gen. nov.

Type species: *Paraphysalidia paralica* sp. nov.

Description.— Test free, small and composed by few globular or subglobular chambers. Outline ovate to sub-cylindrical or reniform, with rounded ends and a median constriction separating an ovate to globular inflated proloculus from the rest of chambers. Wall calcareous, hyaline, thin and lamellar in structure. Surface smooth and finely perforate. Aperture, a long interiomarginal umbilical slit, which can be covered by a more or less pronounced lip, at the base of the final chamber.

Etimology.— From the similarities in general outline and chamber arrangement with the genus *Physalidia* Heron-Allen and Earland.

Stratigraphic range.— Upper Miocene to Recent.

Remarks.— *Paraphysalidia* is similar in general outline to *Physalidia* Heron-Allen and Earland, with which it has in common a thin calcareous hyaline wall, a low number (2–4) of subglobular chambers arranged in opposition, an ovate or reniform outline and an interiomarginal aperture at the base of the last chamber. Other characteristics however are markedly different. *Paraphysalidia* lacks any distinctive imperforate area near the aperture on the opposite chamber, a diagnostic character of the family Bagginidae Cushman, and is much smaller in size. Due to some characteristics shared with the gamontic forms of *Rotaliella elatiana* Pawlowski and Lee, such as its few inflated chambers, its umbilical aperture and its large oval proloculus, *Paraphysalidia* has been here tentatively included in the family Rotaliellidae Loeblich and Tappan, although it lacks other typical traits like bordering umbilical projections.

***Paraphysalidia paralica* sp. nov.**

(Plate 1, figures 1–5; plate 2, figures 1–9)

1999 *Physalidia?* sp., Guillem and Usera, p. 246, plate I, figures 3–5 (Recent, Torreblanca marsh, Eastern Spain).

2002 *Physalidia?* sp., Usera *et al.*, p. 145, figure 3.6 (Recent, Torreblanca marsh, Eastern Spain).

2008 Aff. *Physalidia* sp., Guillem, p. 120–124, plate X, figures 1–6; plate XI, figures 1–6; plate XII, figures 1–6 (Recent, Torreblanca marsh, Eastern Spain).

2008 *Ammonia tepida* var. juvenile, Van Hengstum *et al.*, p. 314, plate 1, figure 13 (Recent, Yucatan Peninsula, Mexico).

2011 *Physalidia simplex*, Van Hengstum and Scott, p. 226, figures 12.6, 12.7 (Recent, Green Bay Cave, Bermuda).

2011 *Physalidia simplex*, Van Hengstum *et al.*, p. 40, figures 4.15, 4.16 (Holocene, Green Bay Cave, Bermuda).

Diagnosis.— A small species of *Paraphysalidia*, always <150 µm in length with a thin wall; oval test with a median constriction and only three chambers including a large oval to globular proloculus; interiomarginal umbilical aperture consisting of a long slit covered by a projection or lip at the base of the last chamber.

Description.— Test free, minute (greatest diameter never exceeding 150 µm), invariably formed by three chambers. Wall calcareous, hyaline, thin and with a laminar aspect in section. Surface smooth and with small sized perforations. Outline oval, with rounded ends and a slight median constriction. Spherical to slightly oval inflated proloculus representing more than a half of the shell length and arranged opposite to the remaining two chambers, which constitute a clearly distinct group, always smaller, less conspicuously perforate than the proloculus and separated from it by the median constriction. Second chamber minute and somewhat pear-shaped. Third chamber intermediate in size and frequently characterized by a somewhat arched apertural zone. Aperture long, interiomarginal, umbilical, slit-like and located at the base of the third chamber, opening towards the proloculus and covered by a short prolongation of the wall edge.

Etimology.— The name *paralica* refers to the environment (mostly marginal marine localities) where this species has been found so far.

Type specimens.— MGUV-24565 (holotype), MGUV-24563–MGUV-24564, MGUV-24566–MGUV-24573, MGUV-24575–MGUV-24577 (figured paratypes) and

MGUV-24574 (unfigured paratype). Deposited in the “Museu de Geologia de la Universitat de València” (MGUV).

Material.— More than 22 000 specimens have been sampled in the Torreblanca marsh.

Type locality.— Torreblanca coastal lagoon, eastern Spain, western Mediterranean coast (coordinates: 40°11'N, 0°12'E). Holotype MGUV-24565 and paratypes MGUV-24563, MGUV-24564, MGUV-24567–MGUV-24573 and MGUV-24575 from sample station TB-8 (coordinates: 40°11'34.3"N, 0°12'55.8"E). Paratypes MGUV-24566 and MGUV-24574 from sample station TB-7 (coordinates: 40°11'33.5"N, 0°12'45.2"E) and paratypes MGUV-24576 and MGUV-24577 from sample station TB-9 (coordinates: 40°11'58.1"N, 0°12'32.3"E).

3. DISCUSSION

3.1. Juvenile forms

The low number of chambers and the large proloculus might suggest that *Paraphysalidia paralica* only represents juvenile megalospheric stages of one of the remaining foraminiferal species in Torreblanca. Prior to further considerations, this possibility must be ruled out. The gamontic megalospheric form of the marine polymorphic species *R. elatiana* Pawlowski and Lee is also characterized by a roughly oval outline, a large proloculus and very few chambers (Pawlowski and Lee, 1992), but they differ from *P. paralica* in many other characteristics such as the deeply open umbilicus with radiating striae and numerous denticles or the circular / arched aperture in *R. elatiana*. The genus *Physalidia* Heron-Allen and Earland, also characterized by few chambers and an oval / reniform outline, has been considered by Hofker (1980) to be just a mix of embryonic megalospheric forms of different Homotrematidae species like *Miniacina miniacea* (Pallas) or *Sporadotrema mesenterica* (Carter), thus not deserving any generic status.

Complete or partial life cycles are known only for few foraminiferal species (Lee *et al.*, 1991; Goldstein, 1999). Three of them occur in the Torreblanca lagoon: *Spirillina vivipara* Ehrenberg, *A. beccarii tepida* (Cushman) and

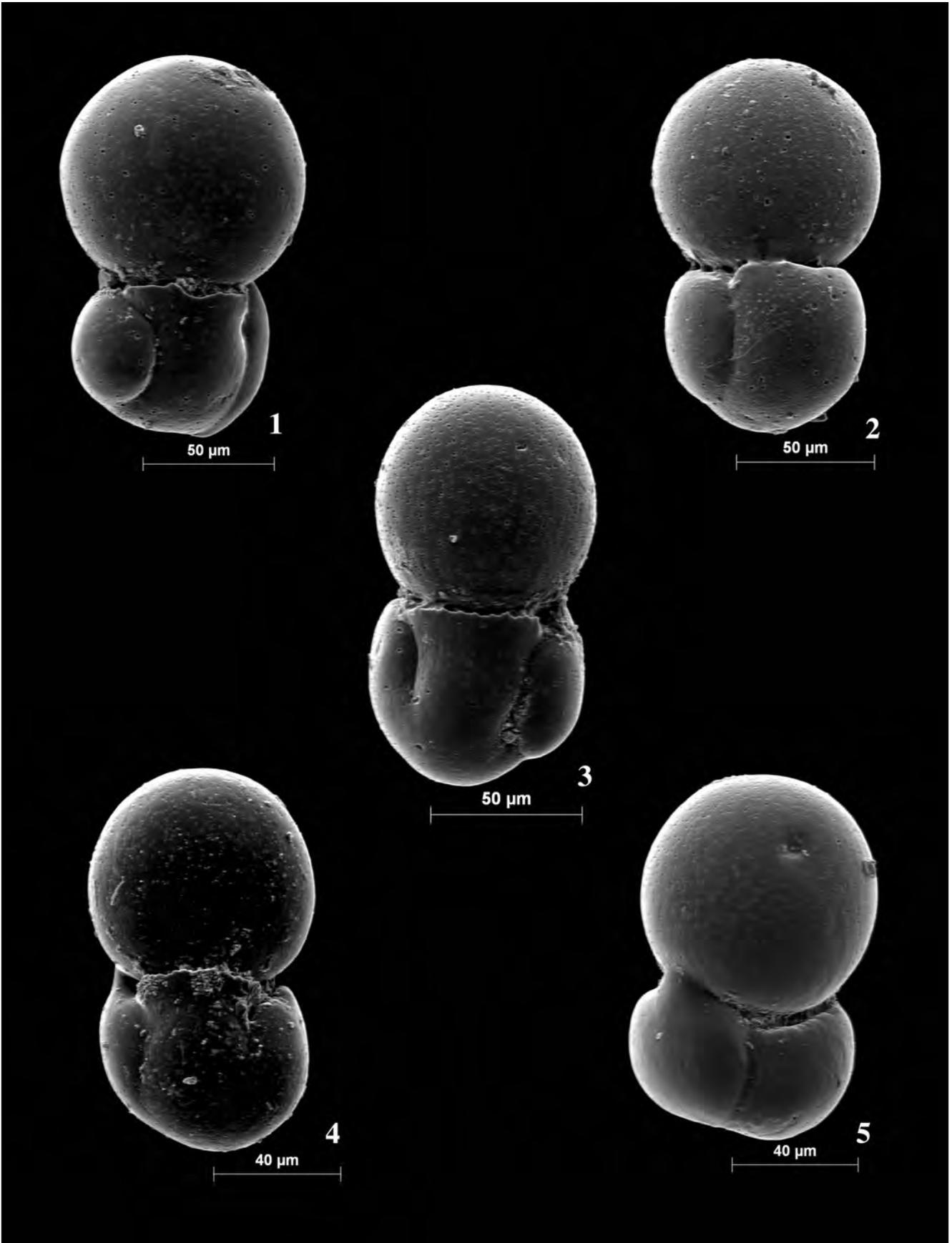
Rubratella intermedia Grell (Guillem, 2008). Our specimens are clearly distinct from any morphologic stage of their cycles and cannot be related to any embryonic megalospheric form of other species detected in the coastal lagoon.

Could those specimens represent propagules or juvenile forms of marine foraminiferal species introduced (e.g. transported by birds or storms) into the Torreblanca lagoon? As found by Alve and Goldstein (2002, 2003, 2010), juvenile forms (consisting only in the proloculus or first growth stages) can be transported as propagules and survive up to two years in a cryptic state until conditions improve and are suitable for growth or even reproduction. Although this might constitute a major way of dispersal for some foraminiferal species, in our opinion, the possibility that *P. paralica* specimens merely represent a mix of first growth stages or juvenile forms introduced into an unfavorable environment can be ruled out basing on the following reasons.

Firstly, there is a quite narrow range of morphological variation between the specimens, so they can undoubtedly be attributed to a single species, i.e. they do not represent a “pool” of juveniles belonging to different taxa. Although they are tiny individuals, many of them are nevertheless large enough to be preserved in the 63-125 µm size fraction. To show this the dimensions of the holotype and the (unbroken) paratypes are given in Table 1. Three measures have been taken in each case: the specimen whole length (≈ maximum diameter), the proloculus width (≈ minimum diameter) and the width at the constriction separating the proloculus from the rest of chambers. As shown in Table 1, although none of the selected specimens exceeds 150 µm, all of them are over 100 µm in length. The proloculus diameters range between 70 and 94 µm. Thus, they are far from being minute cryptic propagules preserved in the <63 µm or even the <32 µm size fractions, as described by Alve and Goldstein (2003, 2010) in the Oslofjord and the Skagerrak Basin (North Sea).

Additionally, as stated above, our specimens are clearly different from juvenile forms of the remaining species oc-

Plate 1. *Paraphysalidia paralica* Guillem and Usera, gen. et sp. nov., SEM photographs. Scale bars are shown next to each specimen. 1, Paratype (MGUV-24563). 2, Paratype (MGUV-24564). 3, Holotype (MGUV-24565). 4, Paratype (MGUV-24566). 5, Paratype (MGUV-24567).



<i>Paraphysalidia paralica</i>	Figure	Width: constriction (µm)	Width: proloculus (µm)	Length (µm)
MGUV-24563	Fig. 2.1	59	94	143
MGUV-24564	Fig. 2.2	54	83	135
MGUV-24565*	Fig. 2.3	56	74	132
MGUV-24566	Fig. 2.4	51	70	120
MGUV-24567	Fig. 2.5	55	76	114
MGUV-24568	Fig. 3.1	52	75	130
MGUV-24569	Fig. 3.2	58	83	145
MGUV-24570	Fig. 3.3	50	73	124
MGUV-24574	Not figured	57	83	139
MGUV-24575	Fig. 3.7	59	85	140
MGUV-24576	Fig. 3.8	51	77	120
MGUV-24577	Fig. 3.9	50	72	138

Table 1. Dimensions of some studied specimens. * Holotype.

curing in the Torreblanca lagoon. Particularly, they can be easily distinguished from 3-chambered stages of *A. beccarii tepida*, with which they could be mistaken.

On the other hand, these forms cannot be related to any foraminifera from marine regions near the Torreblanca lagoon that could have acted as hypothetical source areas for propagules transported to the lagoon. Foraminifera recovered from marine bottoms (18-80 m depth) surrounding the Columbretes islands (off the Spanish coast, less than 50 km away from the Torreblanca lagoon) have been studied and figured by Alberola (1997). Our forms cannot in any way be ascribed to any of those foraminiferal species, neither adult nor juvenile stages.

Finally, all specimens recovered (excepting broken tests, like those in plate 2, figures 4, 5 and 6) had exactly three chambers. If they were embryonic or juvenile forms, a mixing of various growth stages should have been easily found. Even if growth had immediately ceased after introduced into a hostile environment (i.e. the Torreblanca lagoon), at least some 4- or 5-chambered stages should have been detected, as it is the case for other taxa like *A. beccarii tepida*. Nevertheless, despite the great amount of individuals examined (more than 22 000 specimens; Guillem, 2008), not even one of them exceeded three chambers.

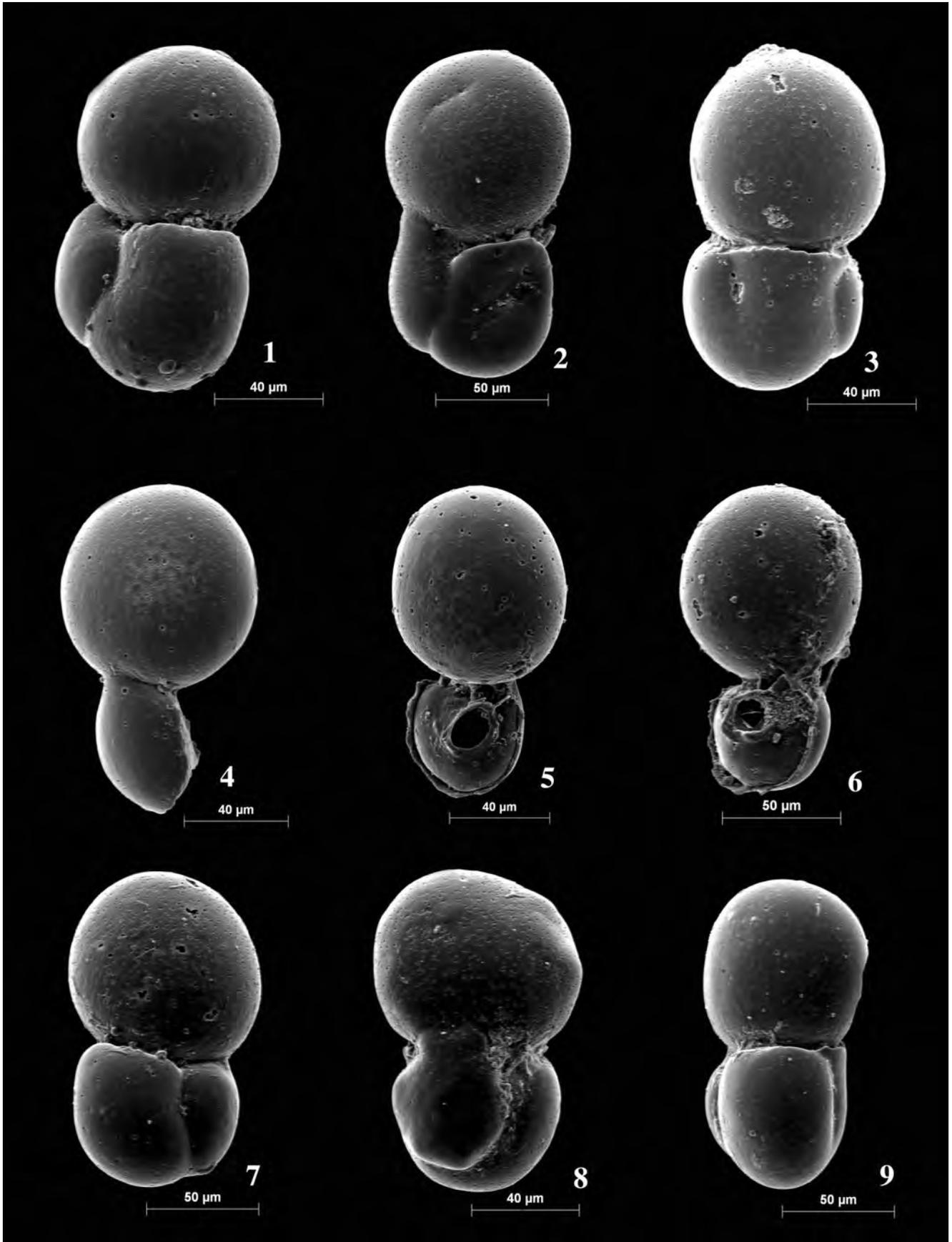
An eventual confirmation of the occurrence of *P. paralica* in the lacustrine beds of the Upper Miocene mammalian site of Venta del Moro, eastern Spain (see below), as an

autochthonous mono-specific assemblage, would provide additional support for its distinct taxonomic status.

3.2. *Physalidia* species

Heron-Allen and Earland (1928) described the genus *Physalidia*, and two species: *Physalidia simplex* Heron-Allen and Earland, later designated as the type species by Galloway (1933), and *Physalidia bicamerata* Heron-Allen and Earland. The description of both species was based on just one specimen each, respectively recovered from marine shallow waters around Rarotonga (Cook Islands) and Papeete (French Polynesia) in the Southern Pacific. *Paraphysalidia paralica* shows an apparent similarity with *P. simplex*: both species are characterized by a test with only three sub-globular chambers arranged in a similar way and a slit-like aperture at the base of the final chamber covered by a slight lip. Our species is equally similar to *P. bicamerata* in its general outline and its cylindrical shape with rounded ends and with a clearly visible median constriction dividing the whole test in two parts arranged in opposition. Nevertheless, *P. paralica* differs from both *P. simplex* and *P. bicamerata* in lacking any imperforate area below the aperture, which characterizes the genus *Physalidia* and even the family Bagginidae according to Loebllich and Tappan (1987). The aperture in *P. paralica* is an umbilical interiomarginal slit and no tubes or thickened areas, like those characterizing the aperture in *P. simplex* and *P. bicamerata*, have been detected. Although the chamber order of succession in *P. simplex* was uncertain for Heron-Allen and Earland (1928) they clearly remarked a middle chamber “larger than the end ones”, which contrasts with the minute size of the corresponding second chamber in *P. paralica*. On the other hand, the test of *P. bicamerata* is apparently constituted by only two chambers instead of three. Seiglie (1965) even considered *P. bicamerata* as different enough from *P. simplex* to deserve a different generic status.

Plate 2. *Paraphysalidia paralica* Guillem and Usera, gen. et sp. nov., SEM photographs. Scale bars are shown next to each specimen. 1, Paratype (MGUV-24568). 2, Paratype (MGUV-24569). 3, Paratype (MGUV-24570). 4, Paratype (MGUV-24571): broken specimen. 5, Paratype (MGUV-24572): broken specimen showing foramen between chambers nº 2 and nº 3 (not present). 6, Paratype (MGUV-24573): broken specimen showing foramen between chambers nº 2 and nº 3 (not present). 7, Paratype (MGUV-24575). 8, Paratype (MGUV-24576). 9, Paratype (MGUV-24577).



Our specimens are identical to those figured by Van Hengstum *et al.* (2008) as *Ammonia tepida* var. juvenile and by Van Hengstum and Scott (2011) and Van Hengstum *et al.* (2011) as *P. simplex*. Nevertheless, they differ from those figured by Loeblich and Tappan (1987, pl. 592, figs. 7 and 8), which clearly show the large middle chamber and the imperforate area below the aperture described by Heron-Allen and Earland (1928) for *P. simplex*.

Physalidia earlandi Bermúdez, from the island of Cuba, shows some traits in common with *P. paralica* as well, such as a test constituted by a few sub-globular chambers with depressed constrictions or sutures between them and an interiomarginal slit-like aperture at the base of the last chamber with a small lip. No imperforate area near the aperture is apparent in the figured holotype of *P. earlandi*, and Bermúdez (1935) does not mention it in the original description (cf. Ellis and Messina, 1940 *et seq.*). However, this species differs from *P. paralica* in having four chambers, instead of three, and a proloculus that is clearly smaller than the subsequent chambers in *P. earlandi*, whereas it is always the largest chamber in *P. paralica*.

Additionally, although size differences are not in principle a taxonomically valid character, since they could be due to ecological intraspecific variation, all the *Physalidia* species are considerably larger than any of the specimens attributed to *P. paralica*. The types of *P. simplex* and *P. bicamerata* are respectively 300 µm and 520 µm in length (Heron-Allen and Earland, 1928) and the type of *P. earlandi* is even larger: 1130 µm (Bermúdez, 1935). In contrast, none of the measured specimens of *P. paralica* exceeds 150 µm in length (Table 1), which is less than the width of any *Physalidia* types and is even smaller than the middle chamber in *P. simplex*.

So far, *Physalidia* species have only been reported from tropical marine settings in the South Pacific and the Caribbean and never in European or Mediterranean localities or anywhere in marginal marine environments, excepting those specimens designated as "*P. simplex*" by Van Hengstum *et al.* (2010, 2011) and Van Hengstum and Scott (2011) in cenotes and caves from Yucatan (Mexico) and Bermuda, which in our opinion are co-specific with *P. paralica*.

3.3. Distribution and stratigraphic range

Paraphysalidia paralica occurs in both dead and living (Rose Bengal stained) assemblages from most sample stations studied in the Torreblanca marsh and generally averages 5–25% of the samples, which are generally dominated by *T. aguayoi* and *A. beccarii tepida*. Nevertheless, its relative abundance can be highly variable totaling more than 60% of the assemblage in some samples. The species seems therefore well adapted to the environmental conditions prevailing in the marsh, characterized by shallow, relatively warm (average: 10°C in January to 28°C in August), mesohaline (salinity range: 6.6–13.4‰), slightly basic (pH mostly between 7.3 and 8.3) and generally well oxygenated waters (Guillem, 2008).

In the Iberian Peninsula, and besides Torreblanca, specimens identified as *Physalidia* sp. and identical to *P. paralica* were found by Blázquez (2005) in Holocene sediments from a drill core obtained in the Elx coastal lagoon (Fig. 1, Alicante, SE Spain, coordinates: 38°12'6.8"N, 0°36'36.3"E). As it is the case for Torreblanca, the foraminiferal assemblages were dominated by *A. beccarii tepida* and *T. aguayoi*, which suggests very similar environmental conditions (Blázquez, 2005; Blázquez and Usera, 2010). Nevertheless, due to its extremely low abundance, the occurrence of *P. paralica* at this locality still requires further confirmation.

The species, named as aff. *Physalidia* sp., was also found by Usera *et al.* (2006) in core samples from the Holocene lagoon of Peníscola (Fig. 1, coordinates: 40°22'N, 0°24'E), located nearby the Torreblanca marsh. It is rare and never exceeds 3% of the total assemblage, which is clearly dominated by *T. aguayoi* and *A. beccarii tepida*, as in Torreblanca and Elx.

Furthermore, specimens undoubtedly identifiable as *P. paralica* have been recently found within lacustrine sediments from the locality of Venta del Moro (Fig. 1, Valencia, Eastern Spain, coordinates: 39°28'33.2"N, 1°20'36.2"E), dated as uppermost Miocene (Opdyke *et al.*, 1990, 1997). This constitutes one of the most interesting sites of the Spanish Neogene fossil record and contains an abundant and diverse mammalian fauna together with remains from many other groups: plants (both pollen and

macroflora), invertebrates (mostly gastropods, bivalves and ostracods) and non-mammalian vertebrates (Morales, 1984; Montoya *et al.*, 2006, 2009, 2011 and references therein), but no foraminifera had been detected so far. Abundant *P. paralica* tests have been found in a 70 cm long microcore sampled within the fossiliferous beds at the Venta del Moro site together with several thecamoebian species, such as *Centropyxis aculeata* (Ehrenberg) and *Diffugia globulus* (Ehrenberg), typical of brackish to freshwater environments. No additional foraminiferal species occurrence has been confirmed yet. According to this data the known stratigraphic range of the species would be Upper Miocene – Recent.

Outside the Iberian Peninsula, *P. paralica*, first identified as juvenile forms of *A. tepida* and then as *P. simplex*, has also been located both in Recent and Holocene sediments of several anchialine caves and cenotes from Yucatan, Mexico (Van Hengstum *et al.*, 2008, 2009, 2010; Gabriel *et al.*, 2009) and Bermuda (Van Hengstum and Scott, 2011; Van Hengstum *et al.*, 2011). Those are particular cases of paralic environments and *P. paralica* has been found either in brackish waters –oligohaline (0,5-5 ‰) to polyhaline (18-30 ‰) – from the meteoric water lenses that float on more saline, marine waters intruding from the coast, or at least in anchialine (i.e. terrestrially influenced) cave biofacies.

Although many marsh foraminifera have narrow altitudinal distributions in tidal marshes and constitute excellent proxies for sea level changes worldwide (e.g. Scott and Medioli, 1980; Edwards and Horton, 2000; Scott *et al.*, 2001; Leorri *et al.*, 2008, 2010; Rossi *et al.*, 2011 among others), no such data can be given for *P. paralica* in the Torreblanca lagoon. Tidal range along the Spanish Mediterranean coasts is always less than 2 m (Zazo and Goy, 1994) and does not exceed 20 cm in the Gulf of Valencia (López García, 1995). Due to this microtidal regime, tides do not have any significant influence on the foraminiferal distributions in coastal lagoons like Torreblanca.

As it happens with many microforaminifera, due to its small size this species has possibly been misinterpreted as a juvenile form or simply overlooked, (Pawlowsky, 1991; Pawlowski and Lee, 1991; Pawlowski *et al.*, 1992, 1993; Pawlowski and Zaninetti, 1993). As stated above,

Van Hengstum *et al.* (2008, 2009) and Gabriel *et al.* (2009) first misidentified this species as juvenile forms of *A. tepida* (cf. Van Hengstum and Scott, 2011) in Mexican anchialine caves. It is therefore suggested that *P. paralica* has probably a much wider distribution and should be found elsewhere in further studies concerning paralic, or even lacustrine (non-marine) environments.

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Palynology and paleoenvironments of Middle Jurassic to Cenomanian successions, Alamein-IX well, north Western Desert, Egypt

Sameh S. Tahoun¹, Mohamed I.A. Ibrahim² and Suzan Kholeif³

¹Department of Geology, Faculty of Science, Cairo University, 12613, Giza, Egypt. stahoun@yahoo.com

²Department of Environmental Sciences, Faculty of Science, University of Alexandria, Moharam Bey 21511, Alexandria, Egypt. mibrah@gmail.com

³National Institute of Oceanography and Fisheries, Alexandria, Egypt.

RESUMEN

Se presentan los resultados palinológicos del estudio de los sedimentos de edad Calloviense tardío-Cenomaniense medio del sondeo Alamein-IX, perforado en la parte norte del Desierto Occidental de Egipto. La sección estudiada abarcó cinco formaciones geológicas que, de base a techo, son las denominadas Masajid, Alam El Bueib, Alamein, Kharita y Bahariya. Como resultado, se han identificado cincuenta y una especies de miosporas y treinta y tres de dinoquistes. Los taxones bioestratigráficamente importantes se han utilizado para diferenciar de forma informal cinco biozonas palinoestratigráficas. Éstas se establecieron principalmente a partir de miosporas en el intervalo correspondiente al Cretácico Inferior, y mediante dinoquistes en el Jurásico Medio- Jurásico Superior. Las palinozonas pertenecen respectivamente a los intervalos Calloviense-Oxfordiense, Neocomiense-Barremiense, Aptiense, Albiense medio-Albiense superior y Cenomaniense temprano-Cenomaniense Medio. Se ha utilizado la palinología para identificar el evento Cimmeriense, que se observa como un hiato entre el Jurásico y el Cretácico, y está caracterizado por la ausencia de depósitos del Oxfordiense tardío, Kimmeridgiense, Portlandiense, Berriasiense y Valanginiense. Los paleoambientes de las sucesiones sedimentarias fluctúan desde condiciones marinas someras (Formación Masajid) e incluso marinas marginales (cerca de la costa y costeras) que se encuentran en la base del sondeo, hasta marinas profundas (Formación Alam El Bueib) y otra vez someras (Formación Alamein). A techo de la sucesión se infieren condiciones marinas marginales (Formaciones Kharita y Bahariya inferior).

Palabras clave: Palinología, paleoambientes, Calloviense-Cenomaniense, Alamein, Desierto Occidental, Egipto.

ABSTRACT

This paper presents the palynological results of the late Callovian – middle Cenomanian sediments in the Alamein-IX well drilled in the northern part of the Western Desert of Egypt. The studied section yielded five formations from base to top, namely Masajid, Alam El Bueib, Alamein, Kharita and Bahariya. Fifty one miospore and thirty three dinoflagellate cyst species were identified. Biostratigraphically important taxa were used to differentiate the sequence into five informal palynostratigraphic biozones, which were established mainly by miospores in the lower Cretaceous interval and mostly by dinoflagellate cysts in the middle-upper Jurassic interval. The palynozones belong to the Callovian-Oxfordian, Neocomian-Barremian, Aptian, middle-upper Albian and early-middle Cenomanian ages, respectively. Palynology was used to identify the Cimmerian event as a hiatus between the Jurassic and Cretaceous, and is characterized by the absence of late Oxfordian, Kimmeridgian, Portlandian, Berriasian and Valanginian deposits. The paleoenvironments of the sedimentary successions fluctuated from shallow marine (Masajid Formation) through marginal marine (nearshore to coastal)

to deeper conditions (Alam El Bueib Formation) and shallow marine (Alamein Formation), and finally marginal marine (Kharita and lower Bahariya formations).

Keywords: Palynology, paleoenvironment, Callovian-Cenomanian, Alamein, Western Desert, Egypt.

1. INTRODUCTION

Since the search for exploitable hydrocarbon reserves on the unstable shelf in the early 1940's, much attention has been paid to studies of the Jurassic-Cretaceous palynomorphs and their use in age assignment, paleoenvironmental interpretation, determination of the source-rocks and the organic thermal maturation history. Palynological studies of Jurassic and Lower Cretaceous deposits in Egypt commenced in the mid 1960's (e.g., Helal, 1965, 1966). Since then there has been a rise in activity concerning the Late Jurassic and Early Cretaceous palynology of Egypt. Relevant publications include Abdelmalik *et al.* (1981), Aboul Ela and Aly (1988), Aboul Ela *et al.* (1989), Aboul Ela and Mahrous (1992), Bassiouni *et al.* (1981), El Beialy (1994a), Ibrahim and El Beialy (1995), Ibrahim and Schrank (1996), Ibrahim *et al.* (2001, 2002), Kedves (1986), Mahmoud (1991), Omran *et al.* (1990), Penny (1991, 1992), Saad and Ghazaly (1976), Schrank (1984), Soliman (1975, 1977), Sultan (1985, 1986, 1987), and Urban *et al.* (1976). Outside Egypt, some important related Early Cretaceous palynological studies comprise Heimhofer *et al.* (2007), Heimhofer and Hochuli (2010), Villanueva-Amadoz *et al.* (2011).

More recent studies (Schrank and Mahmoud, 2000; Ibrahim *et al.*, 2009) based on better resolutions of microscopy, laboratory techniques and palynofloral maturation show that interpretations by some of the pioneering studies are flawed in their palynozonation and paleoenvironmental interpretations. Thus, more comprehensive and higher resolution can be achieved by restudying material published in the mid 1960's and 1970's using newer techniques and available data. The main objective of this paper is to interpret the palynostratigraphy and paleoenvironmental conditions of the Middle Jurassic to Lower Cretaceous successions penetrated by Alamein-IX borehole, north Western Desert, Egypt (Figure 1).

2. STRATIGRAPHIC SETTING

The northern part of the Western Desert is almost featureless plain in spite of its intricate buried geological history. The sedimentary cover is very thick as recorded in many deep boreholes ranging in age from Cambrian to Recent (Abu El Naga, 1984; Barakat *et al.*, 1987; Hantar, 1990). This area was often covered by shallow seas during most of its geological history (Said, 1961). The Jurassic/Cretaceous boundary is unconformable throughout Egypt, and a major break resulting from the "Cimmerian" event separates Jurassic and Cretaceous sequences. This regional event led to tilting, uplift, folding and erosion of Jurassic and older deposits. It occurred during the mid-Tithonian to mid-Berriasian time interval (Keeley *et al.*, 1990; Ibrahim *et al.*, 2002). The Cimmerian event coincides with a worldwide period of eustatic low-stand (Haq *et al.*, 1987). The Alamein-IX borehole is located in the north Western Desert, Egypt (Lat. 30° 36' 37" N and Long. 28° 43' 55" E; Figure 1). The Middle Jurassic-Cenomanian successions encountered in the Alamein-IX borehole are subdivided into five lithostratigraphic units; from base to top, they are the Masajid, Alam El Bueib, Alamein, Kharita, and Bahariya formations (Figures 2 and 3).

2.1. Masajid Formation

The type section of the Masajid Formation is Wadi El-Masajid, Gebel El Maghara (Sinai). This formation is dominated by carbonates with lesser volumes of clastic material, principally shale. In the North Western Desert, the basal part of this formation is mainly composed of shale intercalating with thin layers of sandstone and siltstone. The shale is occasionally dark grey, green, calcareous, silty or sandy. Here the Masajid Formation unconformably underlies the Alam El Bueib Formation (Early Cretaceous). In Alamein-IX well, it is 512 m thick representing the basal part which encountered in the studied borehole (Figures 2 and 3).

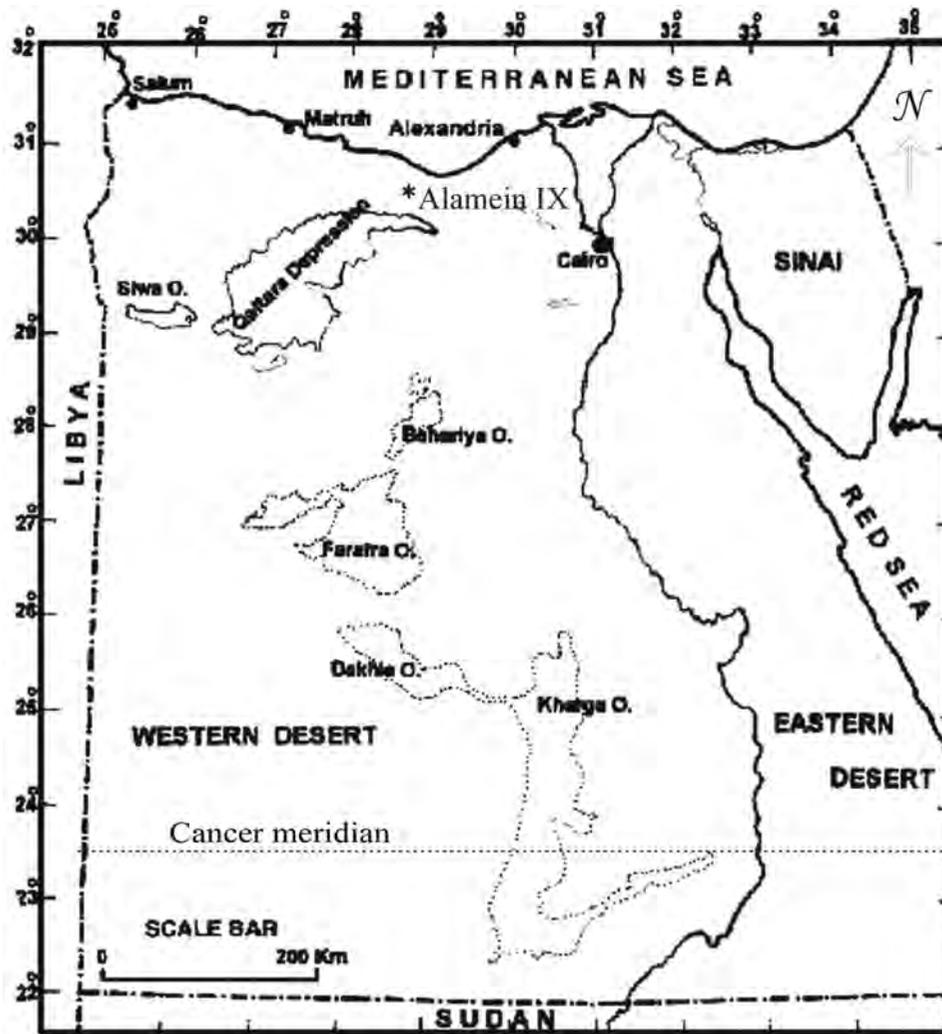


Figure 1. Location map of the Alamein-IX well in the north Western Desert, Egypt.

2.2. Alam El Bueib Formation

The Alam El Bueib Formation is a sandstone unit with frequent shale and occasional limestone beds, both of which, especially the limestone, increase in thickness and abundance to the northwest. The overall thickness of the formation is 1290 m. The type section of this unit is in the Alam El Bueib-1 well, at a drill depth interval between 3927 m and 4297 m. Several operating companies in the area have given this unit different names such as Matruh Group, Aptian clastic, Alamein shale, Dawabis, Shaltut, Umbaraka, and Mamura. The age of the unit is considered to be either Barremian to Aptian (Hantar, 1990) or Berriasian to Aptian (Ibrahim and Schrank, 1996). The environment of deposition has been interpreted to be shallow marine with more continental influence toward the south

(Hantar, 1990). The Alam El Bueib shale is probably the source rock for the gas in the Abu Gharadig field (Khaled, 1999).

2.3. Alamein Formation

Norton (1967) proposed this unit as a member of Burg El Arab Formation. This is a widespread unit known all over North Africa and Arabia and consists of light brown hard microcrystalline and porous dolomite. A few thin shale interbeds are present. The type section of this unit is the interval between 2489 to 2573 m in the Alamein-I well. The Alamein Dolomite grades laterally into shale along a narrow belt that runs to the north and parallel along 30° N between longitudes 31° to 27° and in the Matruh area. Its

thickness is in the range of 20 to 80 m over most of the area except in the north where maximum thicknesses are reported in Kanayis-I (97 m). The age of the Alamein Dolomite is Aptian to Albian in the type area. The unit appears to have been deposited in a shallow marine, low to moderate energy environment (Hantar, 1990). The Alamein Dolomite overlies the Alam El Bueib Formation and it is overlain by Dahab and/ or Kharita formations. The Alamein Dolomite is the hydrocarbon reservoir in the Alamein, Razzak and Yidma fields. In Alamein-IX well, it is 70 m thick.

2.4. Kharita Formation

The Kharita Formation is made up of fine to coarse-grained sandstones with subordinate grey, sub-blocky, medium firm, silty and non-calcareous shale and carbonate beds. The type section is in the Kharita-1 well, at a drill depth interval of 2501 m to 2890 m. The maximum thickness of this unit was reported from the Mersa Matruh-I well (1100 m). The unit was deposited in a high-energy shallow marine shelf. In the extreme north, the unit appears to have been deposited in deeper waters, while in the south it was under the influence of continental conditions (Hantar, 1990). It has a thickness of 244 m in the bore hole studied.

2.5. Bahariya Formation

The Bahariya Formation is composed essentially of variegated shale alternating with sandstone, siltstone and limestone. The shale is varicolored and thinly laminated greyish green to green, and partly calcareous and silty. The siltstone and sandstone are greyish white to yellowish white, glauconitic and pyritic. The Bahariya Formation conformably underlies the Abu Roash Formation. Several names such as Razzak Sand, Meleiha Shale and Medeiwar Member of the Abu Subeiha Formation, have been given previously to the Bahariya Formation by operating companies and other researchers (El Gezeery and O'Connor, 1975). The age of this formation was dated as early-middle Cenomanian (Hantar, 1990; Schrank and Ibrahim, 1995; Ibrahim 2002a, Ibrahim *et al.*, 2009). In the

Alamein-IX well, the formation is encountered from 1585 m to 2175 m and is 590 m thick (Figures 2 and 3).

3. MATERIAL AND METHODS

This study is based on the analysis of 38 ditch samples from the Alamein-IX well. The samples were obtained from Formations, 6 from the basal Masajid, 16 from Alam El Bueib, only one sample from Alamein, 3 from Kharita and 12 samples from Bahariya. The lithostratigraphic succession of the encountered formations in the present well and samples depth are shown in figures 2 and 3.

All samples were processed using standard palynological techniques. The samples were treated with diluted HCl (10%), HF (40%) and concentrated HCl to remove carbonates, silicates and fluorides respectively. The residues were then divided into two portions. The first was used to prepare a kerogen slide; the second was filtered using wet sieving with 10 µm polyester sieves. In order to avoid possible damage or colour change to the palynomorphs, treatment with oxidative agents and hydroxides was avoided. Epoxy resin was used as a mounting medium. The first 300 palynomorphs were counted for statistical purpose. All slides were microscopically examined and diagnostic palynomorph taxa were photographed (Plates 1 and 2). All slides and figured specimens are housed in the palynological collection of the second author (M.I. Ibrahim) in the Department of Environmental Sciences, Alexandria University, Egypt.

4. PALYNOSTRATIGRAPHY

Fifty one miospore and thirty three dinoflagellate cyst taxa have been identified in this study. Based on their abundances, quantitative distribution and last occurrences in the Alamein-IX well, five informal and local palynozones have been proposed for the Callovian to middle Cenomanian interval in the well. The distributions of the miospore and dinoflagellate cyst species are arranged in stratigraphical order according to their last occurrences in Text-Figure 3 and 4, respectively. These palynozones are depicted from top to base as follows:

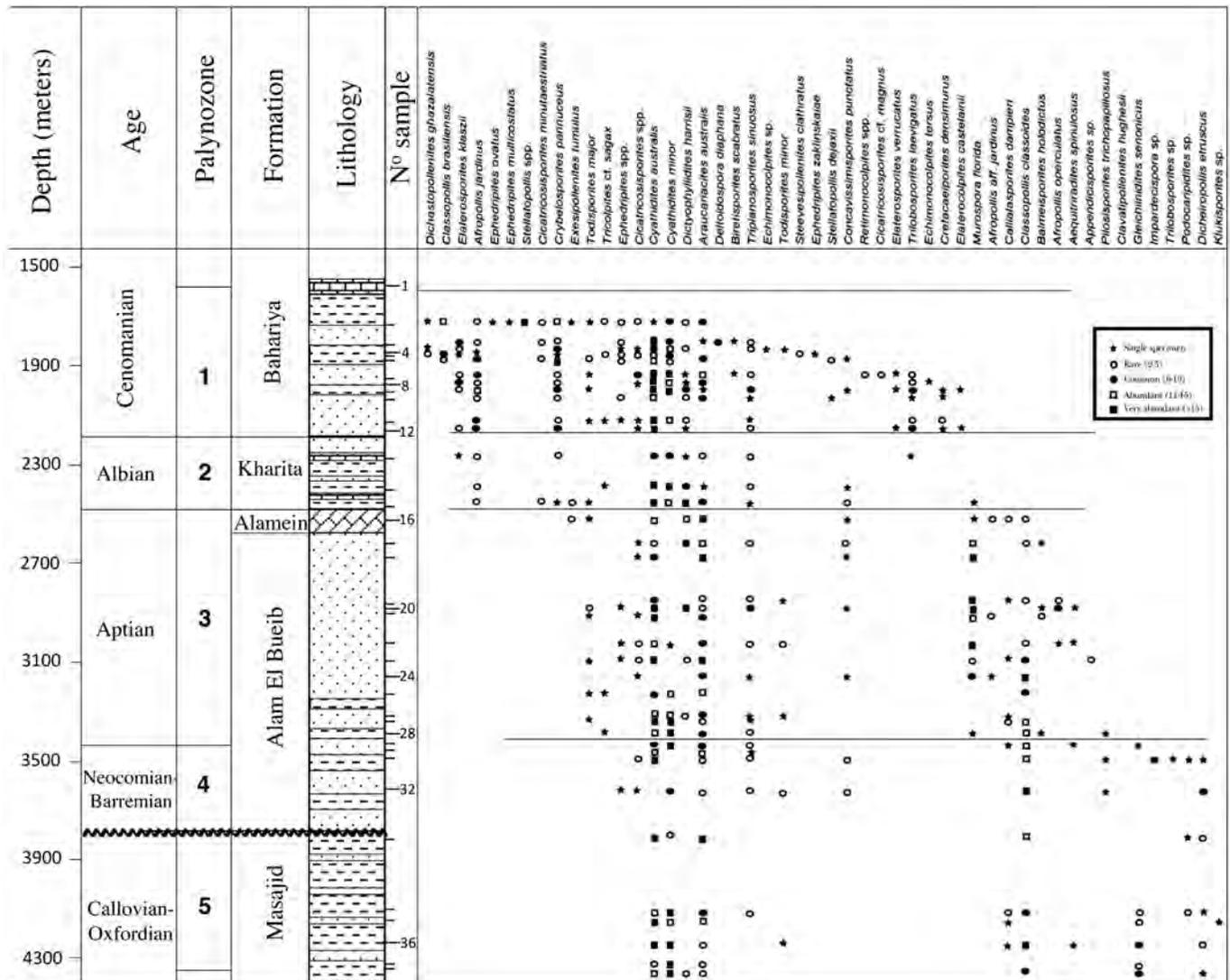


Figure 2. Distribution range chart of the recovered miospore species from the Alamein-IX well in order of highest occurrences.

4.1. Zone 1

Definition: The upper boundary is defined by the top occurrences and co-existence of the zonal index species *Dichastopollenites ghazalatensis*, *Classopollis brasiliensis* and *Elaterosporites klaszii*.

Depth and thickness: This zone covers the interval 1585-2164 m (579 m thick), Bahariya Formation.

Palynological characteristics: Some important miospore taxa are also common in this zone, and they include *Tricolpites cf. sagax*, *Elaterosporites verrucatus*, *Trilobosporites laevigatus*, *Cretaceiporites densimurus* and *Elaterocolpites castelainii*. The main dinocysts occurring

in this zone are *Trichodinium castaneum*, *Cyclonephelium vannophorum*, *Florentinia mantellii*, *Dinopterygium cladoides* and *Xiphophoridium alatum*. It is worth noting that the acritarch *Leiosphaeridia* sp. and the freshwater alga *Pediastrum* sp. are also present in this zone.

Age: Early-middle Cenomanian. *Classopollis brasiliensis* was recorded from the middle-late Cenomanian in different regions in North Africa (Libya and Egypt) by Thusu and Van der Eem (1985), Aboul Ela and Mahrous (1992), Schrank and Ibrahim (1995), Ibrahim (1996); West Africa: Jardiné and Magloire (1965), Lawal and Moullade (1986), and Brazil: Hengreen (1975). Also, *Dichastopollenites ghazalatensis* was identified in the middle-upper Ceno-

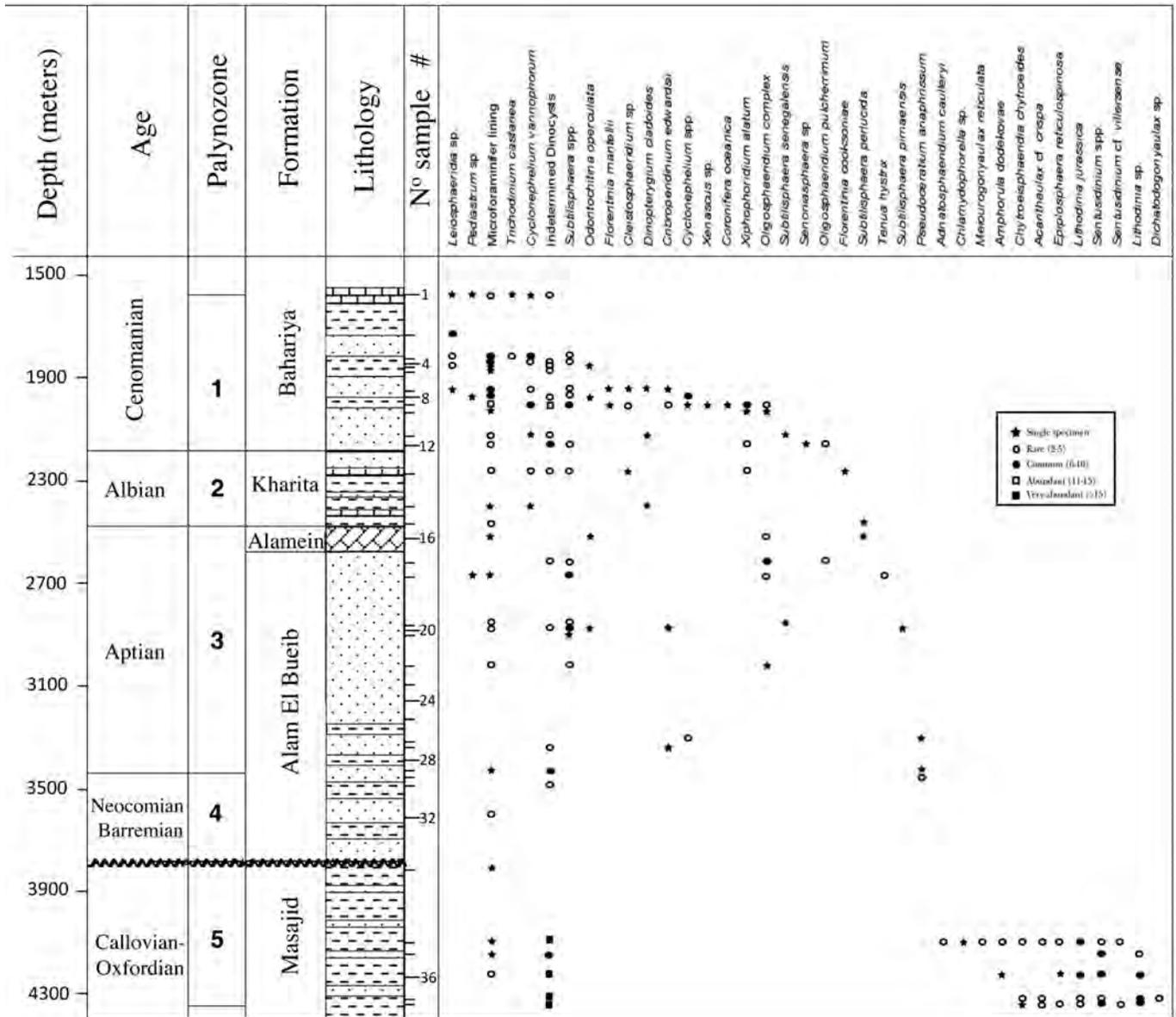


Figure 3. Distribution range chart of the recovered acritarch, fresh algae, microforaminiferal lining and dinoflagellate cysts species from the Alamein-IX well in order of highest occurrences.

manian deposits of Egypt (Schrank and Ibrahim, 1995; Ibrahim, 1996).

Early-middle Cenomanian age for the palynozone is deduced from top ranges of elaterates, *Trilobosporites laevigatus*, *Cretacaeiporites densimurus* and the common occurrence of *Afropollis jardinus* and *Crybelosporites pannuceus* elsewhere in Africa and Brazil (Jardiné, 1967; Hengreen, 1974, 1975; Thusu and Van der Eem, 1985; Lawal and Moullade, 1986; Hengreen and Dueñas-Jimenez, 1990; Aboul Ela and Mahrous, 1992; Schrank

and Ibrahim, 1995; Ibrahim, 1996; Ibrahim *et al.*, 2009). Although the top ranges of the dinoflagellate species *Dinopterygium cladoides* and *Florentinia mantellii* were respectively Santonian in Tethyan realm (Foucher, 1979) but according to the previous studies in north Western Desert of Egypt, they have been recorded in early-middle Cenomanian assemblages (El Beialy 1993, 1994; Ibrahim, 2002a), the top ranges of these two species locally in Egypt were mid-Cenomanian. So, their last appearance data herein confirm this age range.

Correlation: This zone is remarkably comparable to the zones previously erected in Western Desert (Table 1). These zones include the lower part of Zone V and Zone IV of El Shamma and Arafa (1992), Zones II and III of Aboul Ela and Mahrous (1992), upper part of Zone III of Abdel-Kireem *et al.* (1993), Zones V and IV of Schrank and Ibrahim (1995), Zone 3 of Ibrahim (1996), Zone PS-III of Mahmoud and Moawad (2002), Assemblage Zone "B" of Ibrahim (2002a), and Zone I of Ibrahim *et al.* (2009).

4.2. Zone 2

Definition: This zone is defined by the common and continuous co-occurrences of *Afropollis jardinus*, *Crybelosporites pannuceus* and *Araucariacites australis*.

Depth and thickness: This zone spans the interval 2164-2408 m (244 m thick), Kharita Formations.

Palynological characteristics: Other significant miospore taxa occurring in this zone are *Elaterosporites klaszii*, *Concavissimisporites punctatus* and *Cicatricosisporites minutaestriatus*. The major dinoflagellate cysts present are *Subtilisphaera* spp., *Dinopterygium cladoides*, *Xiphophoridium alatum*, *Oligosphaeridium complex* and *Odontochitina operculata*. The freshwater alga *Pediastrum* occur sporadically in this zone.

Age: middle-late Albian. *Crybelosporites pannuceus* is a well-known spore from the Albian of Egypt (Sultan, 1978, 1987; Bassiouni *et al.*, 1992; Abdel-Kireem *et al.*, 1993; El-Beialy, 1994a; Ibrahim, 2002a). It has been reported from the Albian and Albian-Cenomanian of Maryland (Brenner, 1963), Tunisia (Reyre, 1966), Peru (Brenner, 1968), Brazil (Herngreen, 1973), northeast Libya (Batten and Uwins, 1985), Colombia (Herngreen and Dueñas-Jimenez, 1990) and Sudan (Schrank, 1990).

Elaterosporites klaszii is another important middle Albian-middle Cenomanian index taxon. It has been recorded from the Albian of Egypt (Sultan, 1978, 1987; Sultan and Aly, 1986; Omran *et al.*, 1990; Schrank 1991, 1992; Ibrahim, 2002a), the Albian-Cenomanian of West Africa and South America in northern Gondwana, (Stover, 1963, Jardiné and Magloire, 1965; Jardiné, 1967), south Switzerland and Italy (Hochuli, 1981), northeast Libya and Sudan (Batten and Uwins, 1985; Thusu and Van der Eem, 1985; Uwins and Batten, 1988; Schrank, 1990, 1991, 1994). In their review of the distribution of elaterates, Herngreen and Dueñas-Jimenez (1990) observed that *Elaterosporites klaszii* was not older than middle Albian.

The first appearance of *Afropollis jardinus* is used to demarcate Albian deposits from the older deposits (Doyle *et al.*, 1982; Penny, 1989, 1991). Based on the present data,

El Shamma & Arafa (1988)	Schrank (1992)	El Shamma & Arafa (1992)	Aboul Ela & Mahrous (1992)	Abdel-Kireem <i>et al.</i> (1993)	El Shamma & Baioumi (1993)	Ibrahim & El Beialy (1995)	Schrank & Ibrahim (1995)	Ibrahim (1996)	Ibrahim & Schrank (1996)	Mahmoud <i>et al.</i> (1999)	Ibrahim <i>et al.</i> (2001)	Mahmoud & Moawad (2002)	Ibrahim (2002a)	Ibrahim <i>et al.</i> (2009)	Present work
		lower part of Zone V and Zone IV	Zone II and Zone III	upper part of Zone III			Zone V and Zone IV	Zone III				Zone PS-III	Assemblage Zone "B"	Zone I	Zone I
		lower part of Zone IV	Zone IV	Zone III			Zone III	the upper part of Zone II					lower part of Assemblage Zone "A"		Zone II
Zone I					<i>Afropollis operculatus</i> Zone		Zone I	Zone I		Zone III	Zone IV				Zone III
	2 b and 2a					Zone PS I and Zone PS II			Zone III and lower part of Zone VI	middle and upper parts of Zone II					Zone IV

Table 1. Comparison showing the correlation of the present zones with those previously erected in Western Desert of Egypt.

Zone 2 most likely represents a middle-late Albian age. A hiatus exists between Zone 2 and the underlying Zone 3 due to the absence of lower Albian Dahab Formation that have a good turnout in Alamein area.

Correlation: Based on the sporomorph assemblage, this palynozone can be correlated with other zones in Western Desert, including the lower part of Zone IV of El Shamma and Arafa (1992), Zone IV of Aboul Ela and Mahrous (1992), Zone III of Abdel-Kireem *et al.* (1993), Zone III of Schrank and Ibrahim (1995), the upper part of Zone 2 of Ibrahim (1996) and the lower part of Assemblage Zone "A" of Ibrahim (2002a) (see Table 1).

4.3. Zone 3

Definition: Zone 3 is defined by the last appearance datums and restricted occurrences of the zonal marker species *Murospora florida*, *Afropollis* aff. *jardinus*, and *Afropollis operculatus*.

Depth and thickness: This zone is located in the interval between 2408 m and 3424 m (1016 m thick), Alamein Formation and upper part of Alam El Bueib Formation.

Palynological characteristics: *Concavissimisporites punctatus* and *Aequitriradites spinulosus* are the most important miospore elements. *Subtilisphaera* spp., *Oligosphaeridium* spp., *Pseudoceratium anaphrissum*, *Cribroperidinium edwardsii* and *Odontochitina operculata* are among the major dinoflagellate cysts in this zone.

Age: Aptian. The last appearance datum of *Murospora florida* was recorded near the late Aptian boundary in north Egypt (Omran *et al.*, 1990; El Beialy *et al.*, 1990; El Shamma and Baioumi, 1993; Mahmoud and Moawad, 2000; Ibrahim *et al.*, 2001). The zonal marker *Afropollis operculatus* was identified in the early Aptian of Gabon and Senegal (Doyle *et al.*, 1982; Doyle, 1992), Morocco (Gubeli *et al.*, 1984) and Egypt (Schrank, 1983, 1987; Penny, 1986, 1991, 1992; Ibrahim *et al.*, 1995; Schrank and Ibrahim, 1995; Ibrahim, 1996; Ibrahim *et al.*, 2001). *Afropollis* aff. *jardinus* was defined in the Aptian of Gabon, Senegal, England and Egypt (Doyle *et al.*, 1982; Doyle, 1992; Penny, 1989). The combined ranges of key taxa like *Afropollis* aff. *jardinus* and *Afropollis opercula-*

tus in the African-South American Province (e.g. Herngreen and Chlonova, 1981; Doyle *et al.*, 1982; Thusu and Van der Eem, 1985; Schrank, 1987; Penny, 1989; Salard-Cheboldaeff, 1990) suggest an Aptian age for this zone.

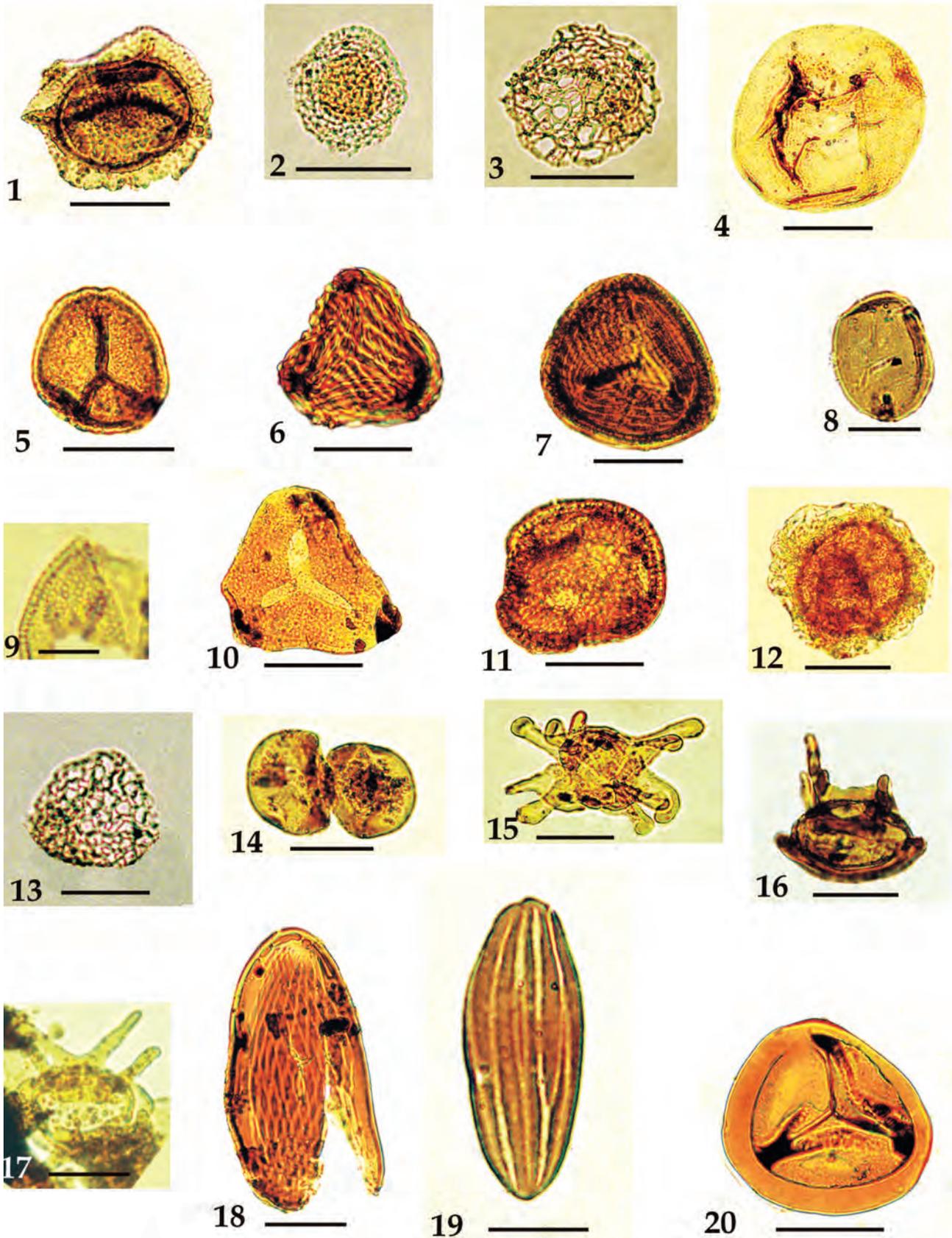
Correlation: This zone is remarkably comparable to the zones established in Western Desert as follows: Zone 1 (El Shamma and Arafa, 1988); Zone III (Mahmoud *et al.*, 1999); *Afropollis operculatus* Zone (El Shamma and Baioumi, 1993); Zone I (Schrank and Ibrahim, 1995); Zone 1 (Ibrahim, 1996); and Zone V (Ibrahim *et al.*, 2001) (see Table 1).

4.4. Zone 4

Definition: Zone 4 is defined by the co-occurrences of the zonal species *Pilososporites trichopapillosus*, *Dicheiropollis etruscus* and *Aequitriradites spinulosus*.

Depth and thickness: This zone spans the interval 3424–3822 m (398 m thick), basal part of Alam El Bueib Formation.

Plate 1. Each transmitted light photograph is sequentially attended by corresponding depth, slide designation beside light microscopic citation coordinates. Scale bar = 30 µm. 1, *Aequitriradites spinulosus* (Cookson and Dettmann) Cookson and Dettmann, 1961; depth 3424 m, slide 1, corr. 112.3/5.6. 2, *Afropollis jardinus* (Brenner) Doyle *et al.*, 1982; depth 1854 m, slide 2, corr. 94.8/11.8. 3, *Afropollis operculatus* Doyle *et al.*, 1982; depth 2899 m, slide 1, corr. 88.4/16.1. 4, *Araucariacites australis* Cookson, 1947 ex Couper, 1953; depth 1731 m, slide 1, corr. 128.6/10.9. 5, *Biretisporites scabratus* Kar and Jain, 1981; depth 1951 m, slide 2, corr. 99.3/4.2. 6, *Cicatricosisporites* cf. *magnus* Döring, 1965; depth 1951 m, slide 1, corr. 131.3/17. 7, *Cicatricosisporites* sp.; depth 1854 m, slide 1, corr. 108.2/14.7. 8, *Classopollis brasiliensis* Herngreen, 1975; depth 1731 m, slide 1, corr. 98.7/3.3. 9, *Clavatipollenites hughesii* Couper, 1958 emend. Kemp, 1968; depth 3424 m, slide 2, corr. 123.4/11.5. 10, *Concavissimisporites punctatus* (Delcourt and Sprumont) Brenner, 1963; depth 2533 m, slide 1, corr. 110.5/8.2. 11, *Cretacaeiporites densimurus* Schrank and Ibrahim, 1995; depth 2012 m, slide 1, corr. 92/8.4. 12, *Crybelosporites pannuceus* (Brenner) Srivastava, 1977; depth 1875 m, slide 1, corr. 108.7/18.1. 13, *Dichastopollenites ghazalatensis* Ibrahim, 1996; depth 1731 m, slide 1, corr. 99.5/7.5. 14, *Dicheiropollis etruscus* Trevisan, 1971; depth 3618 m, slide 2, corr. 102.6/4.2. 15, *Elaterocolpites castelainii* Jardiné and Magloire, 1965; depth 2164 m, slide 1, corr. 96.7/12.6. 16, *Elaterosporites klaszii* (Jardiné and Magloire) Jardiné, 1967; depth 1832 m, slide 1, corr. 114.1/8.3. 17, *Elaterosporites verrucatus* (Jardiné and Magloire) Jardiné, 1967; depth 2164 m, slide 1, corr. 93.4/11.8. 18, *Ephedripites* sp.; depth 2042 m, slide 2, corr. 87.2/16.5. 19, *Ephedripites zaklinskaiae* Azéma and Boltenhagen, 1974; depth 2899 m, slide 1, corr. 104.9/13.7. 20, *Murospora florida* (Balme) Pocock, 1961; depth 2533 m, slide 1, corr. 103/7.1.



Palynological characteristics: The important miospore taxa occurring in this zone include *Clavatipollenites hughesii*, *Impardecispora* sp., and *Trilobosporites* sp. Dinoflagellate cysts are very rare and represented chiefly by *Pseudoceratium anaphrissum*.

Age: Neocomian to Barremian. Abdel Malik *et al.* (1981), El Bealy (1994a) and Ibrahim and Schrank (1996) recorded *Pilosisporites trichopapillosus* in the Barremian sediments of the north Western Desert. The same species had been illustrated in late Neocomian sediments of Western Sahara (Reyre, 1973), Libya (Thusu and Van der Eem, 1985). In Cameroon, Angola, Gabon and Congo, Salard-Cheboldaef (1990) identified this taxon as early Barremian. *Dicheiropollis etruscus* is a good marker species for the late Neocomian (late Hauterivian-early Barremian) in Algeria (Jardiné *et al.*, 1974), Gabon (Doyle *et al.*, 1977), northeast Libya (Thusu and Van der Eem, 1985), north Western Desert of Egypt (Penny, 1986; Ibrahim *et al.*, 1995; Ibrahim and Schrank, 1996), central Sudan (Kaska, 1989; Awad, 1994) and intertropical Africa (Salard-Cheboldaef, 1990).

Although *Dicheiropollis etruscus* was recorded also in succeeding Zone 5, it has not wider range than the currently admitted by different authors in South Europe, Brazil, Africa, South East Asia (Doyle *et al.*, 1977; Salard-Cheboldaef, 1990; Racey and Goodall, 2009). Actually, it has similar range but its presence in Jurassic samples owing to extensive caving process during the drilling of the borehole.

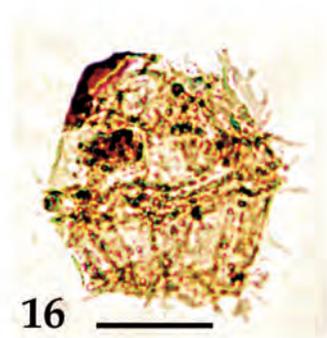
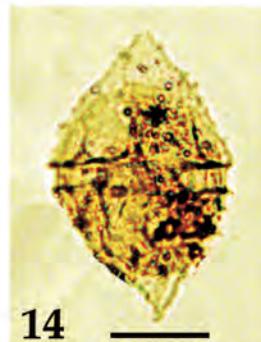
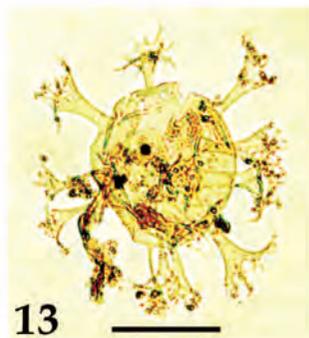
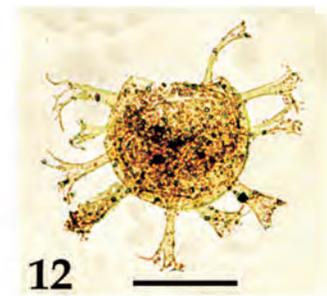
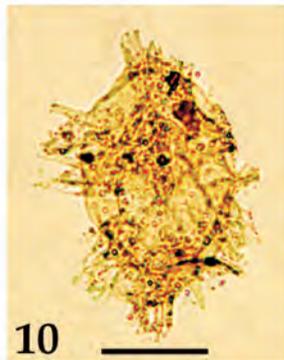
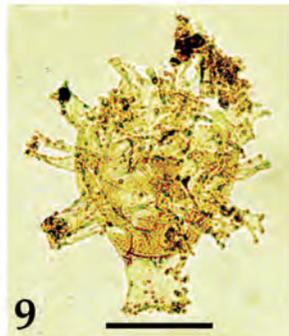
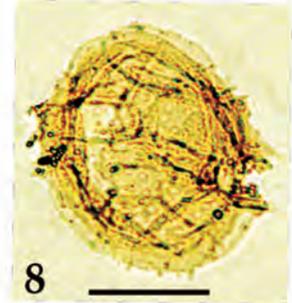
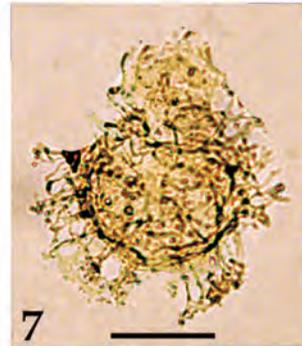
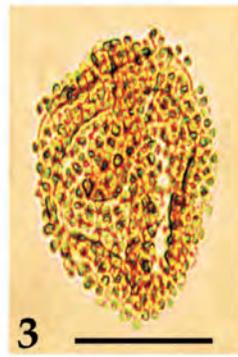
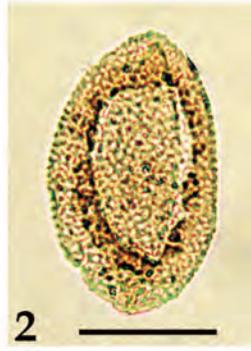
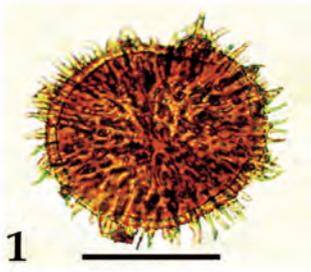
The age assigned to this zone is also supported by the occurrence of the marker *Aequitriradites spinulosus* that characterizes the Barremian in North Africa. Thusu and Van der Eem (1985) and Schrank and Mahmoud (2002) detected its top range in the Barremian deposits of Libya and Egypt, respectively. Also in Egypt, Abdel Malik *et al.* (1981), Schrank (1987), Schrank and Mahmoud (1998), Mahmoud *et al.* (1999) and Ibrahim (2002b) identified it close to the Barremian boundary. The presence of *Clavatipollenites hughesii* was one of the criteria used by Schrank (1992) to define the late Barremian age for sub zone 2b of Egypt and northern Sudan.

For the same reason of caving from Neocomian deposits during drilling, *Aequitriradites spinulosus* was recorded also in the Jurassic samples.

Pseudoceratium anaphrissum is considered an important marker species, which has its top occurrence in the early Aptian/late Barremian in many regions. In Egypt, it has been identified in the Western Desert (El Bealy, 1994b; Ibrahim and Schrank, 1996), Sinai (Ibrahim and El Bealy, 1995), and Eastern Desert (Ibrahim *et al.*, 2002). *Pseudoceratium anaphrissum* is also present in northeast Libya (Uwins and Batten, 1988), northwest Europe and north England (Davey, 1974, 1979; Duxbury 1977, 1980; Harding, 1990; Costa and Davey, 1992). Its presence alongside *Impardecispora* sp. and *Trilobosporites* sp. support the Neocomian-Barremian age proposed for the zone.

Correlation: Zone 4 could be correlatable with several zones previously established in Western Desert, such as subzone 2b and 2a (Schrank, 1992), Zone PS1 and Zone PS2 (Ibrahim and El Bealy, 1995), Zone III and lower part of Zone VI (Ibrahim and Schrank, 1996), and middle and upper parts of Zone II (Mahmoud *et al.*, 1999) (see Table 1).

Plate 2. Each transmitted light photograph is sequentially attended by corresponding depth, slide designation beside light microscopic citation coordinates. Scale bar = 30 µm. 1, *Pilosisporites trichopapillosus* (Thiergart) Delcourt and Sprumont, 1955; depth 3478 m, slide 1, corr. 89.2/16.4. 2, *Retimonocolpites* sp.; depth 1951 m, slide 2, corr. 88.2/17.6. 3, *Stellatopollis dejaxii* Ibrahim, 2002; depth 1875 m, slide 1, corr. 102.5/15.6. 4, *Tricolpites* cf. *sagax* Norris, 1967; depth 1854 m, slide 1, corr. 93.8/11.9. 5, *Trilobosporites laevigatus* El Bealy, 1994; depth 2164 m, slide 1, corr. 132.2/17.4. 6, *Acanthaulax* cf. *crispa* (W. Wetzell) Woollam and Riding, 1983; depth 4334 m, slide 2, corr. 95.1/19. 7, *Adnatosphaeridium caulleryi* (Deflandre) Williams and Downie, 1969; depth 4106 m, slide 1, corr. 113.3/6.8. 8, *Dinopterygium cladoides* Deflandre, 1935; depth 2377 m, slide 1, corr. 105.6/3.3. 9, *Florentinia cooksoniae* (C. Singh) Duxbury, 1980; depth 2292 m, slide 1, corr. 87.8/16.8. 10, *Florentinia mantellii* (Davey and Williams) Davey and Verdier, 1973; depth 1951 m, slide 2, corr. 108.3/19.2. 11, *Lithodinia* sp.; depth 4221 m, slide 1, corr. 92.3/15.5. 12, *Oligosphaeridium complex* (White) Davey and Williams, 1966; depth 2012 m, slide 1, corr. 102.9/4.5. 13, *Oligosphaeridium pulcherrimum* (Deflandre and Cookson) Davey and Williams; depth 2533 m, slide 2, corr. 107.6/8.1. 14, *Subtilisphaera perlucida* (Alberti) Jain and Millepie, 1973; depth 2408 m, slide 1, corr. 96.4/15.2. 15, *Subtilisphaera senegalensis* Jain and Millepie, 1973; depth 2134 m, slide 2, corr. 102.9/20.1. 16, *Xiphophoridium alatum* (Cookson and Eisenack) Sarjeant, 1966; depth 2012 m, slide 2, corr. 107/12.6.



4.5. Zone 5

Definition: Zone 5 is defined by the last occurrences of the index species *Adnatosphaeridium caulleryi*, *Epiplosphaera reticulospinosa*, *Acanthaulax* cf. *crispa* and *Sentusidinium* spp.

Depth and thickness: This zone comprises the interval 3822–4334 m (512 m thick), Masajid Formation.

Palynological characteristics: *Amphorula dodekova*, *Lithodinia jurassica*, *Meiourogonyaux reticulata* and *Sentusidinium* cf. *villersense* are the main dinoflagellate cyst elements in this zone. Among the miospores, *Calialasporites dampieri*, *Classopollis classoides*, *Araucariacites australis*, *Cicatricosisporites* spp. and *Klukisporites* sp. are the most notable taxa.

The top range of *Adnatosphaeridium caulleryi* is Kimmeridgian for Williams and Bujak in Bolli et al., 1985 but Portlandian for Riding and Thomas in Powell, 1992. *Adnatosphaeridium caulleryi* is considered a zonal index for the Callovian in North America and Europe (Williams, 1977; Erkmen and Sarjeant, 1980; Riding, 1982; Riding, 1987; Dimter and Smelror, 1990; Poulsen, 1998). Also, it is present in the early Oxfordian in Switzerland (Berger, 1986); Poland (Courtinat, 1999); Spain (Smelror, 1987) and Morocco (Hssaïda and Morzadec-Kerfourn, 1993). In Egypt, the top range of this taxon is recorded in the Callovian-early Oxfordian (Naim et al., 1990; El Beialy and Ibrahim, 1997).

Epiplosphaera reticulospinosa holotype has been described from Early Kimmeridgian by Klement (1960). The top range of *Epiplosphaera reticulospinosa* has been reported in the early Oxfordian in different regions, such as the northern hemisphere (Stancliffe and Sarjeant, 1988); Palestine (Conway, 1990); Europe (Smelror, 1993) and France (Huault, 1999). In Germany, Prauss (1989) noted its last occurrence in the Callovian deposits.

Lithodinia jurassica is also one of the important taxa with a last appearance datum in the early Oxfordian of Gondwanaland (Stancliffe and Sarjeant, 1988), Libya (Thusu et al., 1988); Europe (Smelror, 1993) and Egypt (Mahmoud and Moawad, 2000). Its last appearance event has been reported in the Callovian sediments in Poland and Ger-

many (Riley and Sarjeant, 1972); Scotland, France and England (Riley and Fenton, 1982), and Egypt (Mahmoud, 1991; Ibrahim et al., 2002).

Amphorula dodekova holotype has been described from Early Kimmeridgian sediment by Zotto et al. (1987). The range chart of *A. dodekova* is Late Oxfordian-Early Kimmeridgian sensu Brenner to Late Kimmeridgian according Zotto et al., 1987, see Monteil, 1990.

Sentusidinium villersense is present in the Callovian in England (Smelror, 1993) and Egypt (Ibrahim et al., 2002). *Acanthaulax crispa* is considered as an important marker for the late Bathonian-Callovian in England and France (Smelror, 1993; Ainsworth et al., 1998; Huault, 1999) and Egypt (Ibrahim et al., 2002). *Sentusidinium* spp. characterized the upper Bathonian-Callovian deposits in Europe (Smelror, 1993), Libya (Thusu and Vigran, 1985), and Egypt (Naim et al., 1990; Mahmoud, 1991).

It is worthy to notice the fact that Masajid Formation underlies directly the unconformity surface between Jurassic and Cretaceous. So, the proposed age in this case depending upon recorded top ranges of the recovered taxa may be slightly younger than the actual age. Herein, the uppermost part of the recorded ranges may be missed in erosion of upper Jurassic succession during the “Cimmerian” event, which coincides with the major break separating the Jurassic and Cretaceous sequences.

Also, the Kimmeridgian age cannot be excluded for the top of the Masajid Formation (see Keeley and Wallis, 1991) and the authors cannot dismiss this possibility.

Age: Callovian-?Kimmeridgian age is proposed based upon the combined top ranges recorded herein for *Adnatosphaeridium caulleryi*, *Epiplosphaera reticulospinosa*, *Amphorula dodekova*, *Lithodinia jurassica*, and *Sentusidinium* cf. *villersense*.

5. PALEOENVIRONMENTAL INFERENCES

The qualitative and quantitative variations in composition of palynomorph assemblages may provide valuable information for interpreting depositional environments (Wall

et al., 1977; Lister and Batten, 1988; Harker et al., 1990; Tyson, 1993, 1995; Batten, 1996; Ibrahim, 2002a). The relative percentage abundance of the recovered spores, pollen grains and dinocysts in the examined Alamein-IX samples are illustrated in Text-Fig. 4.

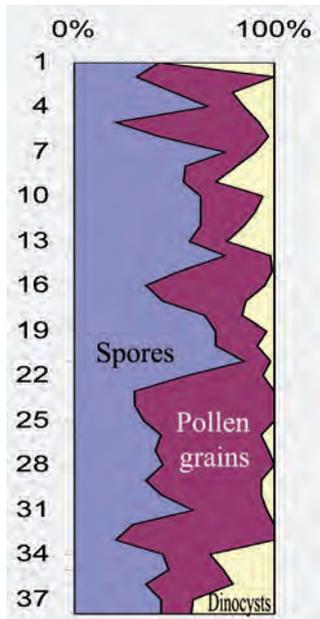


Figure 4. Relative percentage abundance of spores, pollen grains and dinocysts in the examined Alamein-IX samples.

A ternary plot of the main palynomorph elements (Text-Fig. 5) is used for paleoenvironmental interpretations in this study.

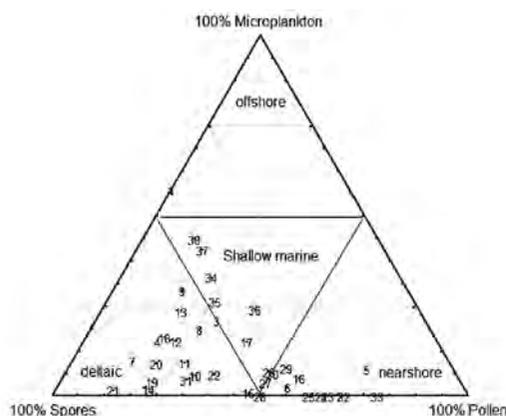


Figure 5. Ternary plot of spores, pollen and microplankton percentages demonstrates the possible depositional environments of the encountered formations in the Alamein-IX well. The numbered plotted points represent the examined samples. Ternary plot modified after Federova (1977) and Düringer and Doubringer (1985).

5.1. Masajid Formation (Callovian-early Oxfordian)

The palynomorph assemblage is distinguished by relatively moderate percentages (32–43.5%) (Text-Fig. 4) of marine microplankton associated with moderate to relatively higher percentages (39–48%) of pteridophytic spores in samples 33 to 38. These evidences indicate shallow marine environment of deposition for the Masajid Formation in the area of study. A warm humid to semi-arid climate is deduced from the considerable amount of the spores including *Cyathidites*, *Deltoidospora* and *Dictyophyllidites* (Schrank, 1987; Thusu et al. 1988; Schrank and Mahmoud, 1998).

5.2. Alam El Bueib Formation (Neocomian - Aptian)

The marine microplankton in Alam El Bueib exhibit relatively low percentages (1.5 – 16%) and even may disappear and are not recorded in samples 23-25, 28 and 32. The upper part of this formation tends to record higher percentages for marine microplankton and pteridophytic spores compared to the lower part. Peridinioid cavate cysts (*Subtilisphaera*) mostly dominate the marine microplankton assemblage. The pollen grains are abundant (39 – 70%) (Text-Fig. 4) and represented mainly by the gymnospermous genera *Classopollis*, *Spheripollenites*, *Araucariacites* and *Inaperturopollenites*. These trends suggest that the probable depositional environments vary from marginal marine (nearshore) in the lower part where the fern spores and peridinioid cysts are the main elements (Ibrahim, 2002a) through deltaic/marginal to shallow marine in the upper part where the pollen and gonyaulacoid take the supremacy (Ibrahim, 2002a); deltaic/fluvial interval occurs in between which may represent a pronounced regressive phase. Warm arid or semi-arid climate were in existence during the deposition of the lower portion of the formation (high gymnospermous pollen including, *Classopollis*, *Araucariacites* and *Inaperturopollenites*) before changing slightly to more humid conditions in the upper portion (high fern spores including, *Cicatricosisporites*, *Aequitriradites*, *Murospora*, *Cyathidites*, *Deltoidospora* and *Dictyophyllidites*).

5.3. Alamein Formation (late Aptian)

Only one sample was available for study in the Alamein Formation; thus its interpretation may not be representative for the entire unit. The sample recorded low percentages for marine microplankton (~ 5 %) and predominance of pollen over spores. Marginal marine (nearshore) environment is inferred with warm semi-arid climate.

5.4. Kharita Formation (middle –late Albian)

The assemblage is characterized by a low to moderate percentage (0.7–23.5%) of marine palynomorphs with increasing trend in the upper part, and an abundance of the spores, in the studied samples, prevailing over the pollen also in the upper part. These criteria indicate that sedimentation took place in a deltaic (lower part, samples: 14, 15) to marginal marine environment (upper part, sample: 13). The high percentage of pteridophytic spores in this formation indicates that a warm humid or semi-arid climate prevailed at that time.

5.5. Bahariya Formation (early-middle Cenomanian)

In most samples representing this formation, pteridophytic spores exceed pollen grains and the abundance and good preservation of both groups suggest that deposition took place relatively close to land. The composition of the spore-pollen assemblage (*Elaterosporites*, *Elaterocolpites*, *Crybelosporites*, *Afropollis*, *Trilobosporites*, *Cicatricosisporites*, *Todisporites*, and *Dictyophyllidites*) suggests that deposition of the Bahariya Formation occurred during a warm humid to semi-arid climate (Sultan, 1986; Schrank and Mahmoud, 1998; Schrank, 2001).

The present lower part of the Bahariya Formation exhibits moderate to high abundances for marine microplankton (2.5–57%) and great diversity (17 species) compared to the overall recorded diversities for this group in the entire studied section. The abundance and diversity of the marine microplankton decrease gradually from the lower to upper part of this formation. Peridinioid cavate cysts (*Subtilisphaera*) dominate the assemblage, while gonyaulacoid

cysts (*Cyclonephelium*, *Cleistosphaeridium*, *Xiphophoridium*, *Florentinia*, *Coronifera*, *Oligosphaeridium*) are subordinate. Many authors (e.g., Williams and Sarjeant, 1967; Wall *et al.*, 1977; Lister and Batten, 1988; Ibrahim, 2002a) have noted the relationship between increasing numbers of dinoflagellate cyst species and seaward direction; a decreasing trend occurs in inshore environments. Powell *et al.* (1992) deduced a Palaeogene peridinioid-dominated assemblage as being indicative of nearshore lagoonal, estuarine, or brackish water environments. Therefore, the domination of peridinioid cysts in this assemblage implies shallow marine – marginal marine (nearshore) depositional environments. It is worth noting that the noticeable counts of foraminiferal test linings recorded in this formation alongside the high diversity of the marine microplankton support deposition in shallow shelf (inner neritic) environments (Lister and Batten, 1988; Tyson, 1993, 1995; Batten, 1996).

6. CONCLUSIONS

A moderate to rich palynomorphs assemblage has been recovered through the palynological investigation carried out on the Middle Jurassic - Upper Cretaceous succession of the Alamein-IX well in north Western Desert of Egypt. Five palynozones in a descending stratigraphical order (Zone 1 to Zone 5) are proposed to cover the studied interval based on stratigraphically important taxa. Depending on the recovered dinocyst assemblage, the Jurassic Zone 5 is dated as late Callovian/Oxfordian, while the Cretaceous Zones 1-4 is dated primarily upon the sporomorphs. The dinocysts in the latter zones are secondary but they confirm the proposed age. The detected Jurassic/Cretaceous boundary in the studied borehole is unconformable.

Paleoenvironmentally, the Masajid Formation was deposited in a shallow marine environment. The lower part of the Alam El Bueib Formation took place in marginal marine (nearshore), the middle part occurred in deltaic/fluviial setting while the upper part was deposited under a deltaic/marginal to shallow marine condition. With respect to the Alamein Formation, unfortunately only one sample was available for the study; thus its interpretation may not be representative for the entire unit. A mar-

ginal marine (nearshore) environment is inferred from that sample. For the Kharita Formation, The observed palyno criteria indicate that sedimentation took place in a deltaic suite in the lower part and a marginal marine environment in upper part. Based on the diversity of the dinocysts in the upper part of the Bahariya Formation, the recorded assemblage implies a shallow marine environment. The noticeable counts of the foraminiferal test linings recorded in the upper part of this formation support well the proposed shallow shelf (inner neritic) depositional environment. A warm arid or semi-arid climate was dominant during the deposition of most of the studied sequence.

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APPENDIX: TAXONOMIC SPECIES LIST

The taxonomy and nomenclature of the recovered miospores are arranged alphabetically according to their botanical affinities. In addition, the nomenclature of the recorded dinoflagellate cysts, acritarchs, fresh water algae and other miscellaneous groups is listed in alphabetical order.

Pteridophytic spores

- Aequitriradites spinulosus* (Cookson and Dettmann, 1958) Cookson and Dettmann, 1961
Appendicisporites sp.
Balmeisporites holodictus Cookson and Dettmann, 1958
Biretisporites scabratus Kar and Jain, 1981
Cicatricosisporites cf. *magnus* Döring, 1965
Cicatricosisporites minutaestriatus (Bolkhovitina, 1961) Pocock, 1964
Cicatricosisporites spp.
Concavissimisporites punctatus (Delcourt and Sprumont, 1955) Brenner, 1963
Crybelosporites pannuceus (Brenner, 1963) Srivastava, 1977
Cyathidites australis Couper, 1953
Cyathidites minor Couper, 1953
Deltoidospora diaphana Wilson and Webster, 1964
Dictyophyllidites harrisii Couper, 1958
Gleichiinidites senonicus Ross, 1949
Impardecispora sp.
Klukisporites sp.
Murospora florida (Balme, 1957) Pocock, 1961
Pilososporites trichopapillosus (Thiergart, 1949) Delcourt and Sprumont, 1955
Todisporites major Couper, 1958
Todisporites minor Couper, 1958
Trilobosporites laevigatus El Beialy, 1994
Trilobosporites sp.
Triplanosporites sinuosus (Pflug, 1952) Thomson and Pflug, 1953

Gymnosperm pollen

- Araucariacites australis* Cookson, 1947 ex Couper, 1953
Callialasporites dampieri (Balme, 1957) Dev, 1961
Classopollis brasiliensis Herngreen, 1975
Classopollis classoides Pflug, 1953 emend. Pocock and Jansonius, 1961
Dicheiropollis etruscus Trevisan, 1971
Ephedripites multicostatus Brenner, 1963
Ephedripites ovatus (Pierce, 1961) Norton in Norton and Hall, 1969
Ephedripites zaklinskaiae Azéma and Boltenhagen, 1974
Ephedripites spp.
Exesipollenites tumulus Balme, 1957
Podocarpidites sp.
Steevesipollenites clathratus Stover, 1964

Angiosperm pollen

- Afropollis jardinus* (Brenner, 1968) Doyle, Jardiné and Doerenkamp, 1982
Afropollis aff. *jardinus* (Brenner, 1968) Doyle, Jardiné and Doerenkamp, 1982
Afropollis operculatus Doyle, Jardiné and Doerenkamp, 1982
Clavatipollenites hughesii Couper, 1958 emend. Kemp, 1968
Cretacaeiporites densimurus Schrank and Ibrahim, 1995
Dichastopollenites ghazalatensis Ibrahim, 1996
Echimonocolpites tersus Ward, 1986
Echimonocolpites sp.
Retimonocolpites sp.
Stellatopollis dejaxii Ibrahim, 2002
Stellatopollis spp.
Tricolpites cf. *sagax* Norris, 1967

Incertae sedis

- Elaterocolpites castelainii* Jardiné and Magloire, 1965
Elaterosporites klaszii (Jardiné and Magloire, 1965) Jardiné, 1967
Elaterosporites verrucatus (Jardiné and Magloire, 1965) Jardiné, 1967

Dinoflagellate cysts

- Acanthaulax* cf. *crispa* (W. Wetzel, 1967) Woollam and Riding, 1983
Adnatosphaeridium caulleryi (Deflandre, 1939) Williams and Downie, 1969
Amphorula dodekovae Zotto, Drugg and Habib, 1987
Chlamydophorella sp.
Chytroisphaeridia chytrooides (Sarjeant, 1962) Downie and Sarjeant, 1965
Cleistosphaeridium sp.
Coronifera oceanica Cookson and Eisenack, 1958 emend. May, 1980
Cribopteridinium? *edwardsii* (Cookson and Eisenack, 1958) Davey, 1969
Cyclonephelium vannophorum Davey, 1969
Cyclonephelium spp.
Dichadogonyaulax sp.
Dinopterygium cladooides Deflandre, 1935
Epiplosphaera reticulospinosa Klement, 1960
Florentinia mantellii (Davey and Williams, 1966) Davey and Verdier, 1973
Florentinia cooksoniae (C. Singh, 1971) Duxbury, 1980
Lithodinia jurassica Eisenack, 1935 emend. Gocht, 1975
Lithodinia sp.
Meiourogonyaualax reticulata Dodekova, 1975
Odontochitina operculata (O. Wetzel, 1933) Deflandre and Cookson, 1955
Oligosphaeridium complex (White, 1842) Davey and Williams, 1966
Oligosphaeridium pulcherrimum (Deflandre and Cookson, 1955) Davey and Williams, 1966
Pseudoceratium anaphrissum (Sarjeant, 1966) Bint, 1986
Senonisphaera sp.
Sentusidinium cf. *villersense* (Sarjeant, 1968) Sarjeant and Stover, 1978
Sentusidinium spp.
Subtilisphaera perlucida (Alberti, 1959) Jain and Millepied, 1973

Subtilisphaera? pirnaensis (Alberti, 1959) Jain and Millepied, 1973

Subtilisphaera senegalensis Jain and Millepied, 1973

Subtilisphaera spp.

Tenua hystrix Eisenack, 1958 emend. Sarjeant, 1985

Trichodinium castanea Deflandre 1935 ex Clarke and Verdier, 1967

Xenascus sp.

Xiphophoridium alatum (Cookson and Eisenack, 1962) Sarjeant, 1966

Acritarchs

Leiosphaeridia sp.

Freshwater algae

Pediastrum sp.

Miscellaneous

Microforaminiferal test linings

Evolutionary classification of the coiled Upper Cretaceous (Turonian-Lower Campanian) planktic foraminifera with simple-ridged test wall

M. Dan Georgescu and Caitlin Carrigy

Department of Geosciences, University of Calgary, 2500 University Drive NW, Calgary, Alberta T2N 1N4 Canada.
dgeorge@ucalgary.ca

RESUMEN

Se realiza una revisión taxonómica de los foraminíferos planctónicos del Cretácico tardío (Turonense-Campanense temprano) del género *Hillsella* Georgescu, 2008a. Se trata de un género tipológico que se modernizó hasta desarrollar un linaje ramificado, una unidad que puede ser utilizada en una clasificación evolutiva. Esta unidad consta de tres especies, *H. hillsi* Georgescu, 2008a, *H. elongans* – nueva especie, y *H. planispiralis* – nueva especie, caracterizada por una concha con pared simplemente crestada, que probablemente evolucionó a partir de *Hedbergella planispira* (Tappan, 1940). La especie más antigua de *Hillsella*, *H. hillsi*, presentó conchas trocoespirales y cámaras globulares a lo largo de su ontogenia. La primera radiación adaptativa a partir de *H. hillsi* se tradujo en la evolución durante el Turonense más tardío de *H. elongans*. Esta última especie presentó cámaras alargadas con una proyección bulbosa en el estado adulto. La segunda radiación adaptativa tuvo lugar durante el Campanense temprano. Durante este período, como resultado de la evolución de *H. planispiralis*, aparecieron especies con conchas enrolladas planoespiralmente. *Hillsella* se extinguió por debajo del límite entre el Campanense inferior y el medio. La singular homogeneidad en la ultraestructura de su concha, características de la porosidad y alta variabilidad en el grosor de su arquitectura, cuestionan la validez de la clasificación tipológica en uso.

Palabras clave: Foraminíferos planctónicos, Cretácico tardío, taxonomía, clasificación evolutiva.

ABSTRACT

Taxonomic revision of the Late Cretaceous (Turonian-early Campanian) planktic foraminiferal genus *Hillsella* Georgescu, 2008a results in its upgrading from typological genus to branched lineage, a unit that can be used in evolutionary classification. It consists of three species, *H. hillsi* Georgescu, 2008a, *H. elongans* - new species, and *H. planispiralis* - new species, which are characterized by simple-ridged test wall and probably evolved from *Hedbergella planispira* (Tappan, 1940). The earliest species of *Hillsella*, *H. hillsi*, exhibits trochospiral tests and globular chambers throughout the ontogeny. The first adaptive radiation from *H. hillsi* resulted in the evolution of *H. elongans* of the latest Turonian, which presents elongate chambers with one distal bulbous projection in the adult stage. The second adaptive radiation resulted in the evolution of *H. planispiralis* in the early Campanian, a globular chambered species with planispirally coiled test. *Hillsella* became extinct below the lower Campanian/middle Campanian boundary. The remarkable homogeneity in the test ultrastructure, porosity characteristics, and high variability in the gross test architecture challenges the validity of typological classification framework in use.

Keywords: Planktic foraminifera, Late Cretaceous, taxonomy, evolutionary classification.

1. INTRODUCTION

Test wall ultrastructure, ornamentation, and pore features (e.g., diameter and density) are paramount morphological features used in the classification of Cretaceous planktic foraminifera. Their importance was demonstrated in the case of all major planktic foraminiferal groups: serials (superfamily Heterohelicacea Cushman, 1927), planispirals (superfamily Planomalinaea Bolli *et al.*, 1957) and trochospirals (superfamilies Rotaliporacea Sigal, 1958 and Globotruncanacea Brotzen, 1942). The taxonomic levels at which such high detailed morphological features are used is apparently a more complex problem and has received various solutions (Huber, 1994). The test ornamentation was first used at the species level, when the first Cretaceous species based on this feature, namely *Globigerina washitensis* and *G. rugosa* respectively (Carsey, 1926; Plummer, 1927) were described. These two studies opened a new path in the classification of Cretaceous planktic foraminifera, which, at that time, was based only on features pertaining of gross test architecture [e.g., chamber arrangement and shape, number and position of the aperture(s), etc].

The use of test ornamentation features in Upper Cretaceous trochospiral planktic foraminiferal taxonomy at the supraspecific levels began with the description of new genera and subgenera based on meridional and/or coarse ornamentation lacking apparent regular distribution (Brönnimann, 1952; Brönnimann and Brown, 1956). Subsequent studies demonstrated the taxonomical importance of test ornamentation in foraminiferal taxonomy and this resulted in describing new units at the genus level and above (Michael, 1972; Petters *et al.*, 1983; Georgescu and Huber, 2006) (Fig. 1). A breakthrough in developing a higher resolution taxonomic framework for the Cretaceous planktics happened when the taxonomic significance of the test ornamentation and porosity was demonstrated (Banner and Desai, 1988). This opened new pathways in understanding the Cretaceous planktic foraminiferal taxonomy and was further refined in a number of high resolution taxonomical studies (Huber, 1994; BouDagher-Fadel *et al.*, 1997; Moullade *et al.*, 2002; Verga and Premoli Silva, 2005; Georgescu 2008a, 2008b, 2009a, 2009b, 2009c, 2010a, 2010b, 2011a, 2011b; Georgescu *et al.*, 2008; Premoli Silva *et al.*, 2009; Georgescu *et al.*, 2011).

Increasing the observation resolution by extensively using the scanning electron microscope (SEM) and environmental scanning electron microscope (ESEM) represented a major advance in the study of Cretaceous planktics. One advantage of SEM/ESEM analysis is that it allows direct observations on test ultrastructure, ornamentation, and porosity. Such features are otherwise often inaccessible when using the classical optical stereomicroscope. By applying this method it became possible to recognize evolutionary relationships between taxonomic units in terms of test wall ultrastructure, ornamentation, and porosity and the development of a non-Linnaean evolutionary classification based on natural units (Georgescu 2009b, 2009c, 2010a, 2011b; Georgescu *et al.*, 2011).

Hillsella was initially described as a monospecific genus from upper Turonian-Coniacian sediments of the Deep Sea Drilling Project (DSDP) Site 150 (Venezuelan Basin, Caribbean region) (Georgescu 2008a); *H. hillsi* Georgescu, 2008a was designated the genus type species. It was originally described as having a cancellate test wall and ornamentation consisting of scattered pustules, with the statement that the test terminology is provisional and the cancellate test wall in *Hillsella* should not be considered in the sense it is used for the planktic foraminifera above the Cretaceous/Paleocene boundary (Georgescu 2008a, p. 57). Georgescu (2011a, b) used the term “simple-ridged” for this kind of test wall. New occurrences of *Hillsella* from two DSDP sites in the Caribbean region (Site 95, Yucatan outer shelf) and equatorial Central Pacific Ocean (Site 463, Mid-Pacific Mountains), including two new species, *H. elongans* and *H. planispiralis*, help in defining its evolutionary history. The new dataset shows that *Hillsella* is a branched lineage characterized by tests with a simple-ridged test wall, a unique feature among the Cretaceous coiled globular-chambered planktic foraminifera. The two distinct evolutionary trends within the *Hillsella* branched lineage led to the development of radially elongate chambers with one distal bulbous projections in the latest Turonian and planispirally coiled tests in the early Campanian.

Figure 1. Synthetic presentation of the coarsely ornamented Cretaceous planktic foraminiferal genera; note that the ornamentation features are among the combinations of diagnostic features. Illustrations show detail chamber surfaces of well-preserved hypotypes of the type species; scale bars = 10 μ m. *Rugoglobigerina* (*Plummerella*) Brönnimann, 1952 was subsequently renamed *Plummerita* (Brönnimann and Brown, 1956, p. 555).

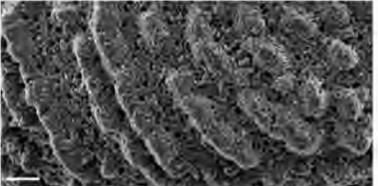
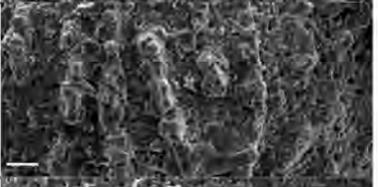
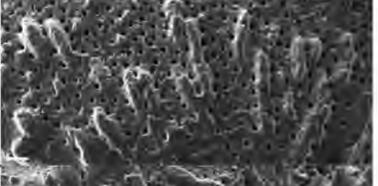
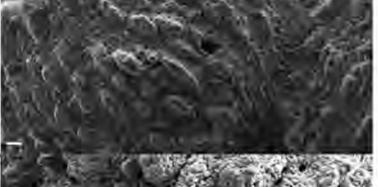
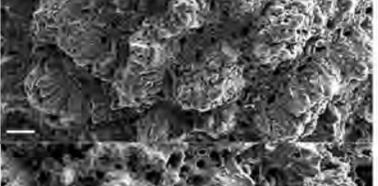
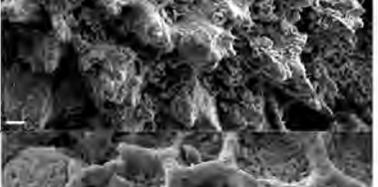
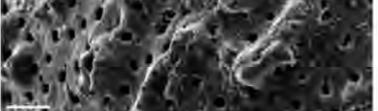
Genus/subgenus (original designation)	Author	Occurrence	Diagnostic features	Ornamentation
<i>Rugoglobigerina</i> (<i>Rugoglobigerina</i>)	Brönnimann (1952)	middle Campanian to Maastrichtian	1) chambers globular; 2) meridional ornamentation consisting of pustules, rugosities and costellae; 3) umbilical system with tegilla.	
<i>Rugoglobigerina</i> (<i>Plumerella</i>)	Brönnimann (1952)	Maastrichtian	1) chambers radially elongate, with tubulospines; 2) meridional ornamentation consisting of pustules, rugosities and costellae; 3) umbilical system with tegilla.	
<i>Trinitella</i>	Brönnimann (1952)	Maastrichtian	1) last formed chambers laterally compressed; 2) meridional ornamentation consisting of pustules, rugosities and costellae; 3) umbilical system with tegilla.	
<i>Rugotruncana</i>	Brönnimann and Brown (1956), emended Georgescu and Huber (2007)	upper Campanian to Maastrichtian	1) two peripheral keels; 2) asymmetrical ornamentation consisting of pustules, rugosities and costellae; 3) umbilical system with tegilla.	
<i>Kuglerina</i>	Brönnimann and Brown (1956)	Maastrichtian	1) chambers globular; 2) high trochospire; 3) papillose ornamentation; 4) periphery broadly rounded and simple.	
<i>Bucherina</i>	Brönnimann and Brown (1956)	Maastrichtian	1) plano-convex shape; 2) very low trochospire; 3) papillose ornamentation; 4) one peripheral keel.	
<i>Favusella</i>	Michael (1973)	Barremian to middle Cenomanian	1) chambers globular; 2) reticulate ornamentation.	
<i>Costellagerina</i>	Petters et al. (1983)	Santonian to lower Campanian	1) meridional ornamentation consisting of pustules, rugosities and costellae; 2) umbilical system with portici.	
<i>Paracostellagerina</i>	Georgescu and Huber (2006)	upper Albian to lower Cenomanian	1) periphery simple; 2) asymmetrical ornamentation consisting of pustules, rugosities and costellae; 3) main aperture bordered by imperforate lip.	

Figure 1.

2. MATERIAL

The material was collected from the Upper Cretaceous sediments of three DSDP sites: Site 150 (Venezuelan Basin, Caribbean region), Site 95 (Yucatan outer shelf, offshore Mexico), and Site 463 (Mid-Pacific Mountains, equatorial Central Pacific) (Fig. 2). The three sections were re-sampled with 5 or 10 cubic centimeter samples and additional samples from the Ocean Micropaleontology Collection (National Museum of Natural History, Washington, D.C.) supplemented the studied material. All the samples are labeled according to the DSDP/Ocean Drilling Program (ODP)/Integrated Ocean Drilling Program (IODP) standards as follows: leg number-site number-core number-core section, interval sampled in centimeters. The type material of the newly described species is deposited in the Willi Karl Braun Micropaleontological Collection of the University of Calgary.

Upper Cretaceous deep oceanic carbonates were sampled from Cores 9 and 10 in DSDP Site 150; this stratigraphic interval was originally assigned to the Turonian–Santonian (Premoli Silva and Bolli, 1973). Revised biostratigraphy shows two distinct sedimentary episodes of late Turonian(?) Coniacian and late Santonian-early Campanian age, respectively (Table 1). Two biozones are recognized in the upper Turonian(?) Coniacian stratigraphic interval,

namely *Marginotruncana schneegansi* Biozone and *Dicarinella concavata* Biozone; the boundary between them is defined by the first occurrence of *D. concavata* (Brotzen, 1936). Only the lower part of the latter biozone occurs at this site, representing the uppermost Turonian and possibly lower part of the Coniacian stage. The uppermost part of the section is assigned to the undivided upper Santonian-lower Campanian interval based on the occurrences of two serial planktic foraminiferal species: *Hendersonites carinata* (Cushman, 1938) and *Ventilabrella eggeri* Cushman, 1928. Planktic foraminiferal assemblages are very rich and diverse in most of the samples collected at this site; fragmentary tests or with traces of dissolution are frequent throughout the section. Two *Hillsella* species occur in the upper Turonian(?) Coniacian sediments: *H. hillsi* and *H. elongans* n.sp. (Table 1).

New *Hillsella* occurrences are reported from the upper Turonian–lower Santonian distal shelf–upper bathyal carbonates of DSDP Site 95. The deep oceanic carbonates recorded in Cores 13 through 17 were originally assigned a Santonian age (McNeely, 1973; Nederbragt, 1991). A succession of four planktonic foraminiferal biozones of the upper Turonian–lower Campanian stratigraphic interval was identified (Georgescu and Huber, 2009; Georgescu, 2010a) (Table 2). *Hillsella hillsi* is the only species of the genus that occurs in the lower portion of

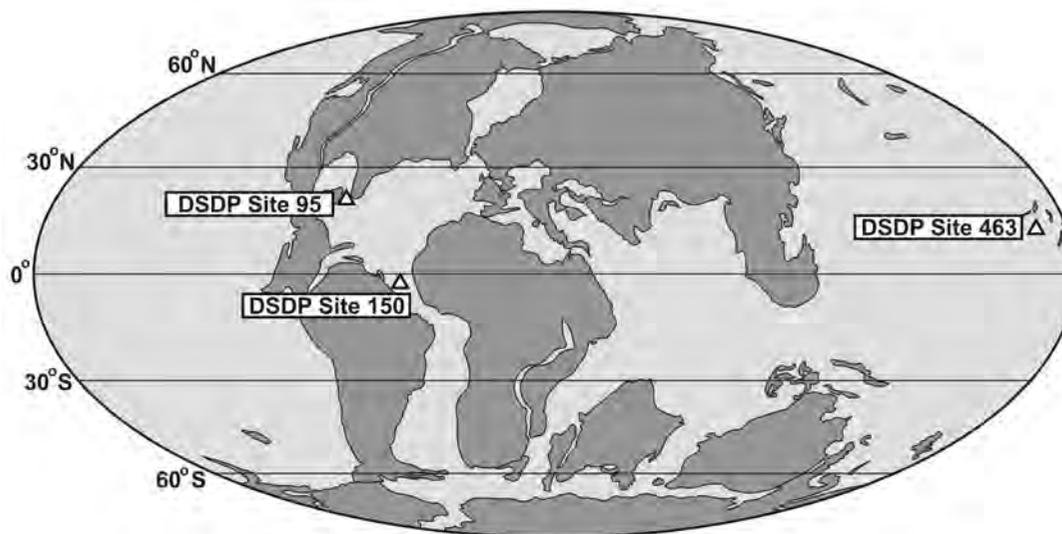


Figure 2. Geographical position of the DSDP sites that yielded occurrences of *Hillsella*. Paleogeographic reconstruction map at 100 m.y. (Hay *et al.*, 1999-simplified).

Samples	Stages	Planktic foraminiferal biozonation	Selected planktic foraminiferal species														
			<i>Hendersonites carinata</i>	<i>Ventilabrella eggeri</i>	<i>Globigerinelloides</i> sp.	<i>Schackoina cenomana</i>	<i>Hedbergella delrioensis</i>	<i>Hillsella hillsi</i>	<i>H. elongans</i> n. sp.	<i>Whiteinella ballica</i>	<i>Dicarinella</i> cf. <i>concovata</i>	<i>D. concavata</i>	<i>D. primitiva</i>	<i>Marginotruncana renzi</i>	<i>M. schneegansi</i>	<i>M. pseudolinneiana</i>	<i>M. coronata</i>
9-1, 46-47 cm	Santon - L. Campan	NO INDEX	R	R	R												
9-2, 94.5-95.5 cm	upper Turonian - (?) Coniacian	<i>D. concavata</i> (lower part)		C	R	R	F			F	R	A	F	A	R	R	
10-1, 25-26 cm				R		R	A	R	R	C	R	A	F	F			
10-1, 58-73 cm*				R	R	F	A	R		?	R?	A	A	C	R		
10-1, 75-76 cm				C		R	F	F				F	C	R			
10-2, 24-25 cm			R	R	R	C	F				F	C	R				
10-2, 48-62 cm*			R	R	F	R	F	R	C		C	R	F	R	R		
10-2, 75-76 cm			R		R	R	R			R			F	C	R		
10-core catcher				R		R			R				R	R			

Table 1. Stratigraphic distribution and frequencies of selected planktic foraminiferal species in the upper Turonian-(?) Coniacian sediments at DSDP Site 150 (Venezuelan Basin, Caribbean region). The samples from the Ocean Micropaleontology Collection (NMNH) are marked with an asterisk. Frequency key: R-rare (1-5 specimens/sample), C-common (6-10 specimens/sample), F-frequent (11-25 specimens/sample), A-abundant (26 or more specimens/sample).

Samples	Stages	Planktic foraminiferal biozonation	Selected planktic foraminiferal species													
			<i>Globigerinelloides asperus</i>	<i>Hillsella hillsi</i>	<i>Whiteinella ballica</i>	<i>Costellagerina pilula</i>	<i>Dicarinella primitiva</i>	<i>D. concavata</i>	<i>D. asymetrica</i>	<i>Marginotruncana sigali</i>	<i>M. pseudolinneiana</i>	<i>M. coronata</i>	<i>Contusotruncana fornicata</i>	<i>Globotruncana arca</i>	<i>Globotruncanites elevata</i>	<i>G. stuartiformis</i>
13-1, 101.5 -103 cm	lower Campanian	<i>G. elevata</i>				R							R	C		R
13-2, 101.5 -102.5 cm	Santonian	<i>D. asymetrica</i>							R	R		R	R	C		
13-3, 100-102 cm			F							A	F		A	A	R	R
13-4, 100-101 cm			C			R				F	C		R	A	R	R
14-1, 102-105 cm						R			C	A	R	R	C	R	R	R
15-1, 97-100 cm								R	R	A	A		R			R
15-2, 100.5-101.5 cm						R				A	F	R				R
15-3, 101.5-102.5 cm						F			R	C	A	C				
15-4, 99.5-100.5 cm						F				A	R	A	R			R
16-1, 99.5-101 cm								R	R	F	C	C				
16-2, 100.5-102.5 cm					R			R	F	F	A	F	R			
16-3, 97.5-99 cm	Coniacian	<i>D. concavata</i>	C	R			R	R		F	A	F				
16-4, 100.5-101.5 cm				R					F	A	C	C				
16-5, 100.5-102 cm				R	R		F			A	A	A				
16-6, 99-100 cm										R						
17-1, 99.5-100.5 cm							C	R		F	A	R				
17-2, 101.5-102.5 cm					R			R	R		F	R	R			
17-3, 100-101 cm	upper Turonian	<i>M. schneegansi</i> (upper part)					R		C	R	R					
17-4, 102-103.5 cm				R	R		R		F	F	R					
17-5, 99.5-100.5 cm				R	R					C	A					
17-6, 100-101 cm				C						F	F					

Table 2. Stratigraphic distributions and frequencies of selected planktic foraminiferal species in the Turonian-lower Campanian sediments at DSDP Site 95 (Yucatan outer shelf, offshore Mexico). Frequency key as in Table 1.

the section, namely in the upper part of the *M. schneegansi* Biozone and throughout *D. concavata* Biozone; this species is a minor component of the planktic foraminiferal assemblages, which are dominated by double-keeled species of *Dicarinella*, *Marginotruncana*, *Contusotruncana*, *Globotruncana* and *Globotruncanita*.

This report of *Hillsella* from DSDP Site 463 is also the first occurrence of the genus from outside the Caribbean region. Turonian-Campanian sediments were recognized in the initial report between Cores 24 and 35 (Boersma, 1981). The existence of an unconformity spanning the middle Coniacian-middle Campanian at this site (Ando et al., 2009) cannot be recognized in the succession of samples at our disposal. The biostratigraphic framework for the lower Turonian-middle Campanian (from the *Helvetoglobotruncana helvetica* Biozone throughout

Globotruncana ventricosa Biozone) is revised herein (Table 3). *Hillsella hillsi* and *H. planispiralis* n. sp. are the two species of the branched lineage that occur in the deep oceanic carbonate succession at this site in the Turonian-lower Campanian and lowermost Campanian, respectively. The *Hillsella* species are minor components of the deep water planktic foraminiferal assemblages dominated by plano-convex species of *Helvetoglobotruncana*, *Dicarinella*, *Marginotruncana* and *Globotruncanita*.

The foraminiferal tests are well preserved; the calcitic test wall presents the common low-magnitude recrystallization but still retains the high detail structures, ultrastructure, ornamentation elements and porosity features (Fig. 3). These well-preserved features demonstrate that the material can sustain a high resolution taxonomical study.

Samples	Stages	Planktic foraminiferal zonation	Selected planktic foraminiferal species																					
			<i>Schackoina cenomana</i>	<i>Hillsella hillsi</i>	<i>H. planispiralis</i> n. sp.	<i>Helvetoglobotruncana praehelvetica</i>	<i>H. helvetica</i>	<i>Dicarinella algeriana</i>	<i>D. hagni</i>	<i>D. concavata</i>	<i>D. asymetrica</i>	<i>Marginotruncana marianosi</i>	<i>M. coronata</i>	<i>M. sigali</i>	<i>M. schneegansi</i>	<i>Globotruncana bulloides</i>	<i>G. ventricosa</i>	<i>Globotruncanita elevata</i>	<i>G. stuartiformis</i>	<i>Contusotruncana plummerae</i>	<i>C. fornicata</i>	<i>Archaeglobigerina blowi</i>		
24-1, 50-52 cm	middle Campanian	<i>G. ventricosa</i>																						
24-2, 50-52 cm																								
24-3, 50-52 cm																								
25-1, 51-53 cm	lower Campanian	<i>G. elevata</i> equivalent			F											A	R		C	F	F	C		
25-2, 51-53 cm					F											R			C	F	A	R		
26-1, 50-52 cm					F												F			R	F	A	R	
26-2, 52-54 cm					C												A			C	F	A	R	R
26-3, 52-54 cm					R	R	R										C			C	F	F	C	
26-4, 52-54 cm	upper Santonian	<i>D. asymetrica</i>	R	C						C	F				F			R		F	C	R		
26-5, 53-58 cm	Coniacian-lower Santonian	<i>D. concavata</i>	R	C						F											A	R	R	
26-6, 53-55 cm					R					R					R	C					F	C		
27-1, 50-52 cm					R						F					F	A					A	F	
27-2, 20-22 cm	Turonian	<i>M. schneegansi</i>								F											R	R		
29-1, 50-52 cm					R										A	R	R							
30-1, 50-53 cm					R	R									A	F	A							
30-2, 50-52 cm					R										R	C	R							
31-1, 11-13 cm															R	A	A							
33-1, 50-52 cm					R											C								
33-2, 52-54 cm																								
34-1, 50-52 cm																								
34-2, 53-55 cm																								
34-2, 75-78 cm					R	R																		
34-3, 50-53 cm																								
35-1, 10-12 cm																								

Table 3. Stratigraphic distributions and frequencies of selected planktic foraminiferal species in the Turonian-middle Campanian sediments of DSDP Site 463 (Mid-Pacific Mountains, equatorial Central Pacific Ocean). Frequency key as in Table 1.

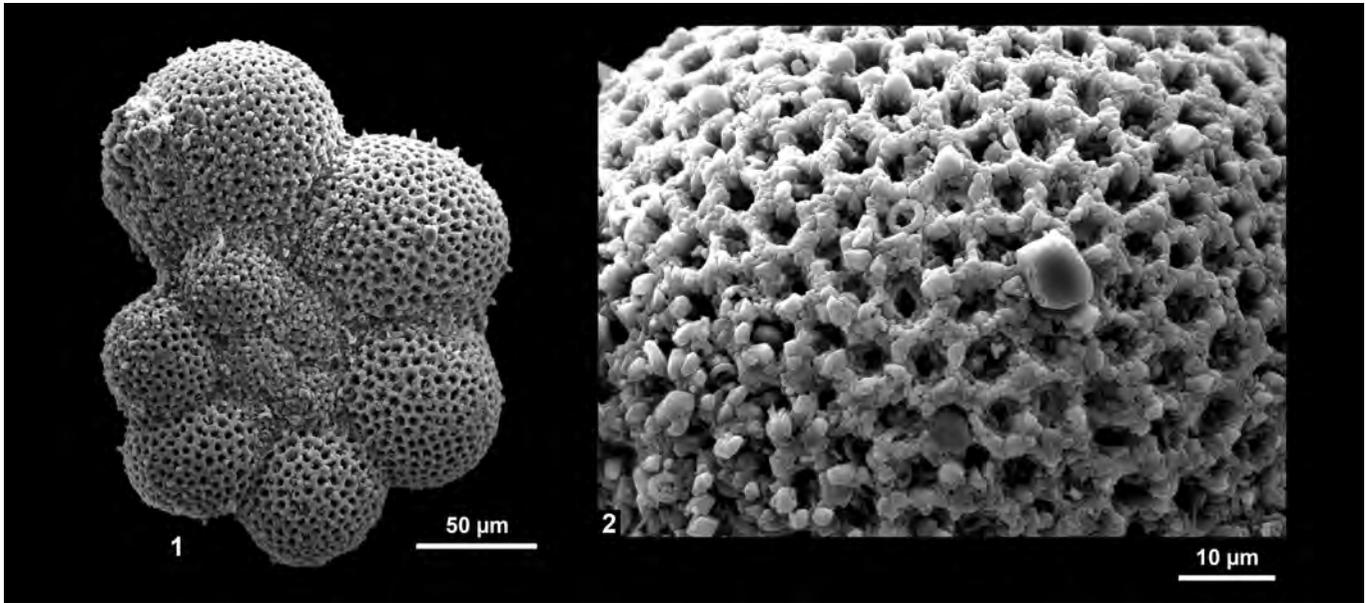


Figure 3. Specimen of *Hillisella hillsi* from the type locality showing the discrete test calcite recrystallization. The regular wall ultrastructure and porosity characteristics indicate that the high detail morphological features can be used in a taxonomy study.

3. SYSTEMATIC DESCRIPTIONS

Evolutionary classification units and acronyms above the species level are after Georgescu (2010a, 2011c). The typological categories of order and genus are maintained temporarily until the development of an evolutionary classification framework at higher levels. The composite paleontological species concept is followed throughout (Georgescu and Huber, 2007, 2009).

Order FORAMINIFERIDA Eichwald, 1830

Branched lineage *Hillisella* Georgescu, 2008a

Type species: *Hillisella hillsi* Georgescu, 2008a.

2008a *Hillisella* Georgescu, p. 58.

Emended diagnosis.— Coiled Cretaceous planktic foraminifera with, simple-ridged test wall.

Emended description.— Test trochospiral to planispiral consisting of globular or perpendicularly to tangentially elongate chambers, occasionally with distal bulbous projections. Sutures are distinct, radial and depressed. Periphery is broadly rounded, occasionally with imperforate peripheral band on the earlier chambers of the last whorl. Main aperture is a low to high arch bordered by an imperforate lip, extraumbilical-peripheral in the trochospiral

species and equatorial-peripheral in the planispiral species. Umbilicus is shallow, medium-sized and often with relict apertures. Chamber surface is smooth or ornamented with small and scattered pustules. Test wall is calcitic hyaline, simple-ridged and perforate.

Remarks.— *Hillisella* was originally described from upper Turonian sediments of the Venezuelan Basin, Caribbean region as a monospecific genus (Georgescu, 2008a). The simple-ridged test wall, which is unique among the hedbergellid planktic foraminifera, had a major role in defining this genus. The two new occurrences from the Yucatan outer shelf (Caribbean region) and Mid-Pacific Mountains (equatorial Central Pacific Ocean) refine our knowledge on its stratigraphic distribution and evolution, primarily by confirming and expanding the taxonomic significance of the simple-ridged test wall. *Hillisella* is herein redefined as a branched lineage that exhibits a high variability with respect to the chamber arrangement and shape: the earlier species (Turonian-Santonian) are trochospiral with globular or elongate chambers whereas the later (early Campanian) and evolved species is planispirally coiled. *Hillisella* differs from any other Cretaceous planktic foraminiferal genera with trochospiral or planispiral coil, with globular or radially elongate chambers and non-keeled periphery by the simple-ridged test wall.

Species included.— *Hillsella hillsi* Georgescu 2008a, *H. elongans* n. sp. and *H. planispiralis* n. sp.

Occurrences.— Upper Turonian-(?) Coniacian (Venezuelan Basin), upper Turonian-Coniacian (Yucatan outer shelf) and Turonian-lowermost Campanian (Mid-Pacific Mountains).

IS: *Hillsella hillsi* Georgescu, 2008a

(Pl. 1, Figs 1-11, Pl. 2, Figs 1-7, Figs 3: 1-2, 6: 10-12)

2008a *Hillsella hillsi* Georgescu, p. 59, pl. 1, figs 1-4, pl. 2, figs 1-3, pl. 3, figs 1-4, text-fig. 1: 4.

2011a *Hillsella hillsi* Georgescu – Georgescu, fig. 2: 1-2, 2: 7-8.

Diagnosis.— *Hillsella* with trochospiral test and globular chambers.

Description.— Test is a very low trochospire, consisting of 13 to 15 globular chambers arranged in 2-2 ½ whorls; there are 5 ½-6 ½ chambers in the last whorl. Chambers increase gradually in size; the last-formed one can be occasionally slightly radially elongate. Sutures are distinct, radial and depressed. Test shape is biconvex, nearly symmetrical in edge view; periphery is broadly rounded and without peripheral structures. The main aperture is a medium high arch, extraumbilical-peripheral in position and bordered by an imperforate lip. Umbilicus is wide and shallow, frequently with relict apertures; umbilical diameter is one fourth to one third of the maximum test diameter. Earlier chambers are ornamented with small and scattered pustules, 2.2-3.3 µm in diameter. Test wall is calcitic, simple-ridged, hyaline and coarsely perforate; pores are circular to subpolygonal and with a diameter of 3.1-4.2 µm, rarely smaller.

Remarks.— *Hillsella hillsi* differs from other Cretaceous hedbergellid taxa with globular chambers throughout by the simple-ridged test wall.

Occurrences.— Upper Turonian-(?) Coniacian (Venezuelan Basin), upper Turonian-Coniacian (Yucatan outer shelf) and Turonian-lowermost Campanian (Mid-Pacific Mountains).

FDS: *Hillsella elongans* n. sp.

(Pl. 2, Figs 8-11, Pl. 3, Figs 1-11)

1973 *Hastigerinoides alexanderi* (Cushman, 1931) – Premoli Silva and Bolli, pl. 4, figs 1, 5.

1973 *Clavihedbergella simplex* (Morrow, 1934) – Premoli Silva and Bolli, pl. 4, figs 2-4.

2002 *Clavihedbergella watersi* (Cushman, 1931) – Wilson et al., fig. 1:2.

Hillsella n. sp. – Georgescu 2011a, fig. 2: 3-4.

Holotype.— Specimen WKB 010098.

Dimensions of the holotype.— Maximum diameter: $D_{max}=0.227$ mm; minimum diameter: $D_{min}=0.136$ mm; $D_{min}/D_{max}=0.599$; thickness: $T=0.061$ mm; $T/D_{max}=0.269$; umbilical diameter: $U_D=0.041$ mm; $U_D/D_{min}=0.304$; $U_D/D_{max}=0.181$.

Paratypes.— Five specimens, WKB 010099-010103.

Dimensions.— $D_{max}=0.183-0.257$ mm; $D_{min}=0.119-0.193$ mm; $D_{min}/D_{max}=0.599-0.816$; $T=0.061-0.084$ mm; $T/D_{max}=0.269-0.367$; $U_D=0.041-0.064$ mm; $U_D/D_{min}=0.304-0.391$; $U_D/D_{max}=0.181-0.284$. Ranges are based on the average measurements of the holotype and paratypes.

Material.— Circa 40 specimens.

Type locality.— DSDP Site 150 (Caribbean region, Venezuelan Basin); geographical coordinates: 14°30.69'N and 69°21.35'W.

Type level.— Upper Turonian (*M. schneegansi* Biozone), Sample 15-150-10-2, 48-62 cm.

Derivation.— Species name is inspired by the elongate shape of the last-formed chambers of the final whorl.

Diagnosis.— *Hillsella* with low trochospiral test and elongate chambers in the adult.

Description.— Test is a very low trochospire, consisting of 10 to 13 chambers arranged in 2-2 ½ whorls; there are four to five chambers in the final whorl. Earlier chambers are globular, the last one to four are perpendicular to the previous whorl to tangentially elongate and with a distal bulbous projection in the adult specimens. Chambers increase in size at variable rate, which is low in the earlier portion and high in the last whorl. Sutures are distinct, ra-

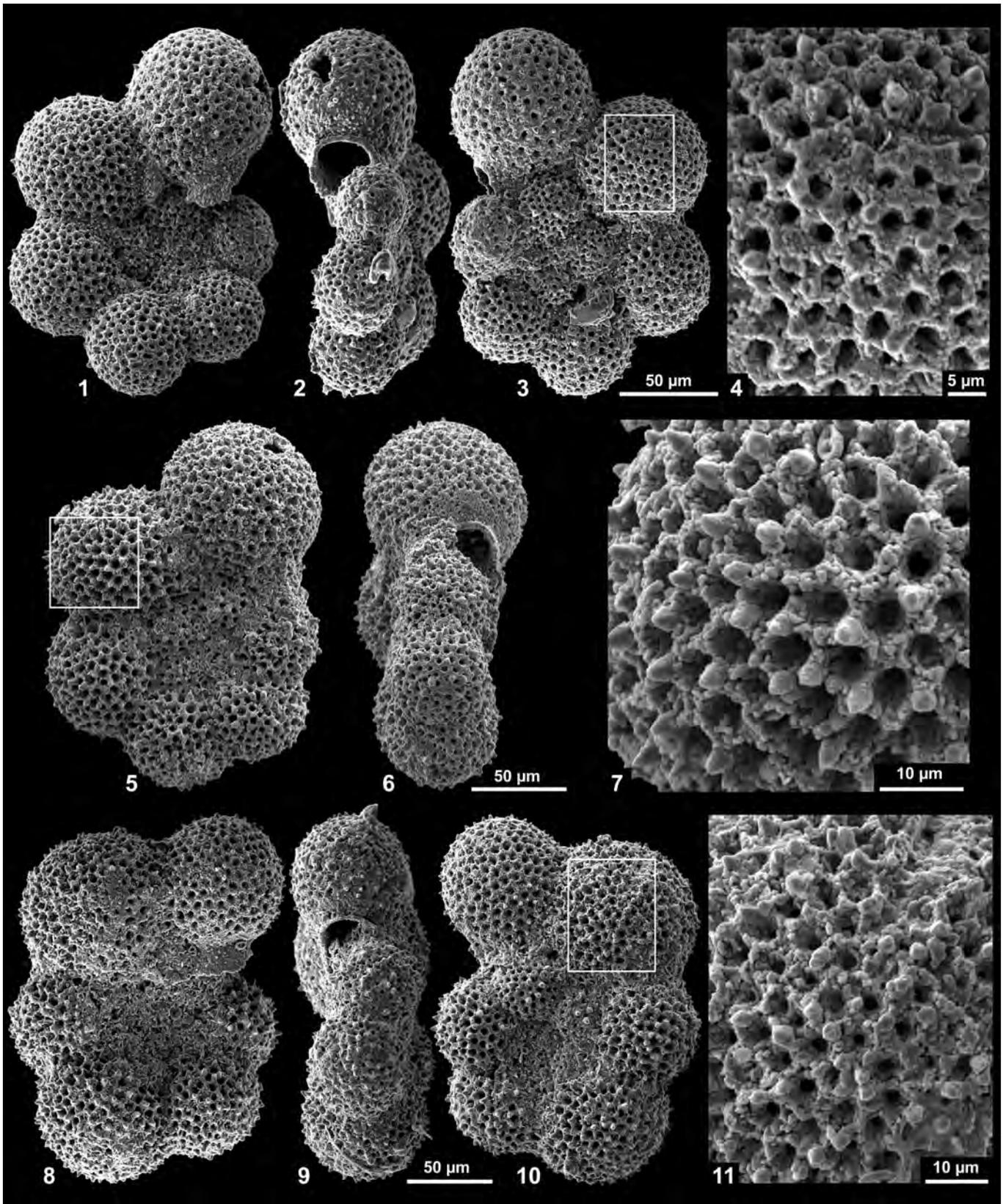


Plate 1. Specimens of *Hillsella hillsi* Georgescu, 2008a from the upper Turonian sediments (*M. schneegansi* Biozone) at the type locality, DSDP Site 150 of the Caribbean region (Venezuelan Basin). 1-4, Holotype, Sample 10-150-10-1, 58-73 cm. 5-7, Paratype, Sample 10-150-10-1, 58-73 cm. 8-11, Paratype, Sample 10-150-10-1, 58-73 cm.

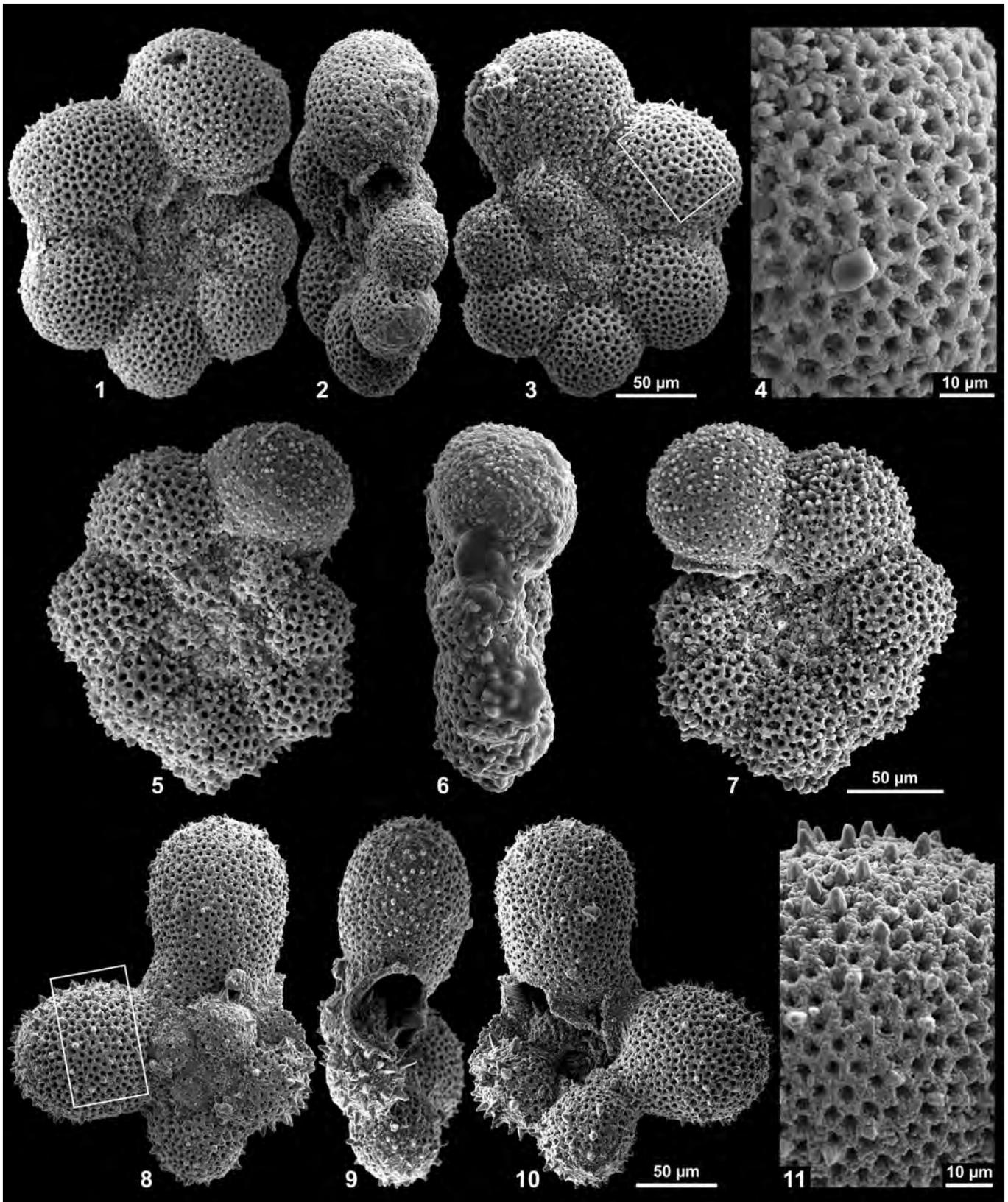


Plate 2. Specimens of *Hillisella hillsi* Georgescu, 2008a from the upper Santonian-lower Campanian (*D. asymetrica* and equivalent of *G. elevata* Biozones) sediments of DSDP Site 463 (equatorial Central Pacific Ocean, Mid-Pacific Mountains) and *H. elongans* n. sp. at the type locality, DSDP Site 150 of the Venezuelan Basin (Caribbean region). 1-4, Hypotype, Sample 62-463-26-4, 52-54 cm. 5-7, Hypotype, Sample 62-463-26-3, 52-54 cm. 8-10, Paratype, Sample 10-150-10-2, 46-62 cm.

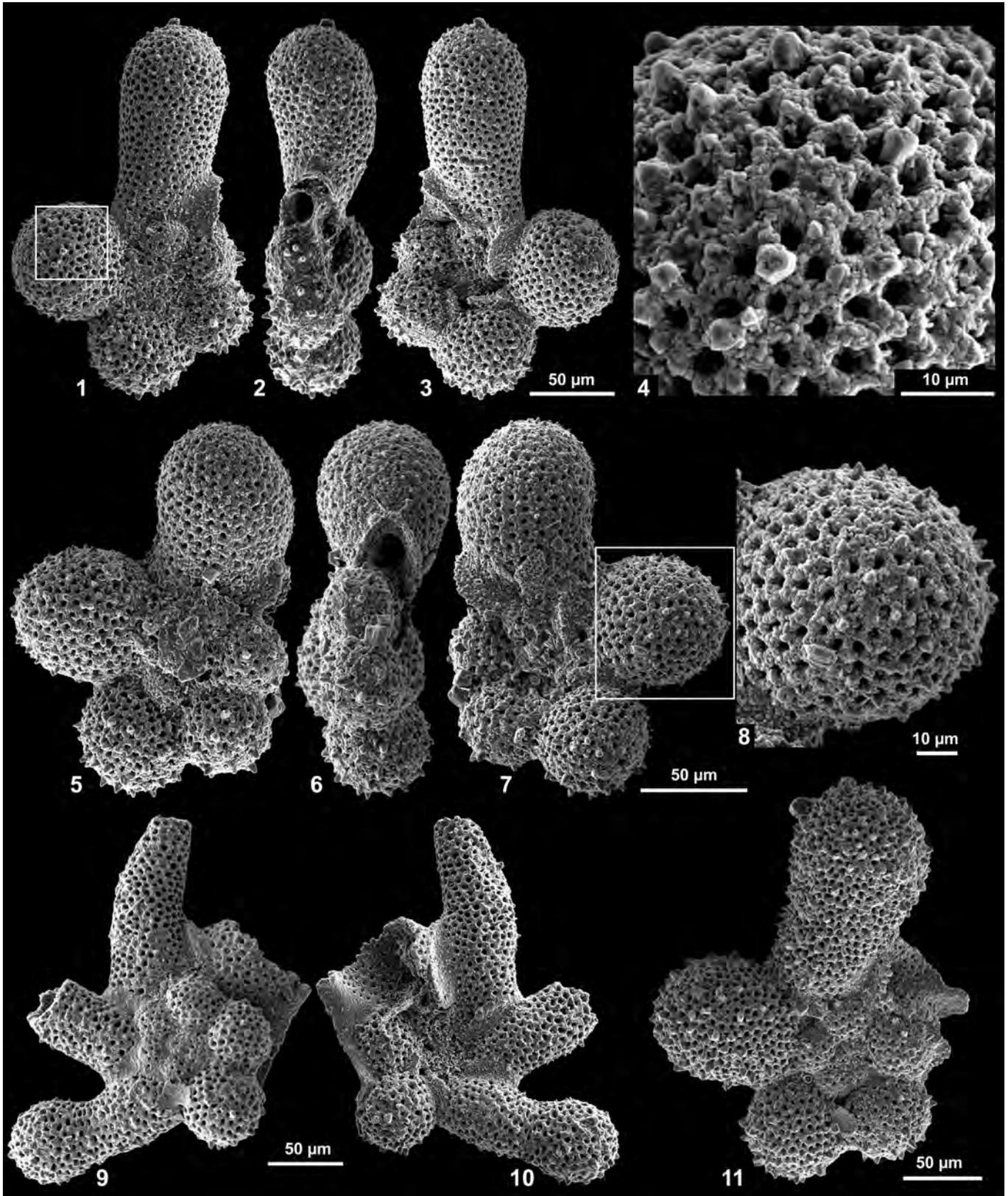


Plate 3. Specimens of *Hillisella elongans* n. sp. from the upper Turonian sediments (*M. schneegansi* Biozone) at the type locality, DSDP Site 150 of the Venezuelan Basin (Caribbean region). 1–4, Holotype, Sample 10–150–10–2, 46–62 cm. 5–8, Paratype, Sample 10–150–10–2, 46–62 cm. 9–10, Topotype, Sample 10–150–10–1, 58–73 cm. 11, Paratype, Sample 10–150–10–2, 46–62 cm.

dial and depressed. Test shape is biconvex, nearly symmetrical in edge view; periphery broadly rounded and occasionally with an imperforate band over the earlier chambers of the final whorl. Main aperture is a medium high to high arch in extraumbilical-peripheral position and bordered by an imperforate lip. Umbilicus is wide and shallow; relict apertures consistently occur in the umbilical region. Chambers are ornamented with small-sized scattered pustules, 3.2-5.5 μm in diameter. Test wall calcitic hyaline, simple-ridged and perforate; pores circular to elliptical, with a diameter of 2.0-3.2 μm .

Remarks.— *Hillsella elongans* differs from *H. hillsi* by having the last one to four chambers of the final whorl with perpendicular to tangential elongation and occasionally with a distal bulbous projection. It differs from *Pessagnoina moremani* (Cushman, 1931) by having a simple-ridged test wall rather than simple and smaller pores (2.0-3.2 μm rather than 2.7-4.1 μm) and from *Eohastigerinella watersi* by having trochospiral rather than planispiral test coiling and simple-ridged test wall rather than simple to incipiently simple-ridged. This species differs from *P. amabilis* (Loeblich and Tappan, 1961) and *E. subdigitata* (Carman, 1929) by having simple-ridged test wall and distal bulbous projection; the strong resemblances between the three species (*P. amabilis*, *H. elongans*, and *E. subdigitata*) are the result of the iterative evolution of chamber elongation from hedbergellid ancestors in the upper Albian-Santonian stratigraphic interval). Although the test morphology is similar, there are discrete differences in the test ultrastructure, porosity, and ornamentation between the three species, which demonstrate their different origins from species of the hedbergellid stock; a special mention is on the middle-late Cenomanian specimens of *P. simplex* (Morrow) reported by Leckie (1984), and assigned by Georgescu (2009b, p. 278) to *C. amabilis*. *Hillsella elongans* is the only non-schackoinid species with elongate chambers in the adult stage in the upper Turonian. Its stratigraphic occurrence appears isolated and bounded by two intervals lacking non-schackoinid taxa with elongate chambers, namely middle-lower upper Turonian and uppermost Turonian-lower Coniacian (Fig. 4).

Occurrence.— Upper Turonian of the Venezuelan Basin.

SDS: *Hillsella planispiralis* n. sp.

(Pl. 4, Figs 1-11)

2011a *Hillsella* n. sp. – Georgescu, fig. 2: 5-6.

Holotype.— Specimen WKB 010104.

Dimensions of the holotype.— $D_{\text{max}}=0.220$ mm; $D_{\text{min}}=0.176$ mm; $D_{\text{min}}/D_{\text{max}}=0.800$; $T=0.070$ mm; $T/D_{\text{max}}=0.318$; $U_D=0.070$ mm; $U_D/D_{\text{min}}=0.398$; $U_D/D_{\text{max}}=0.318$.

Paratypes.— Five specimens WKB 010105—010109.

Dimensions.— $D_{\text{max}}=0.203-0.225$ mm; $D_{\text{min}}=0.143-0.182$ mm; $D_{\text{min}}/D_{\text{max}}=0.704-0.827$; $T=0.070-0.079$ mm; $T/D_{\text{max}}=0.318-0.369$; $U_D=0.060-0.076$ mm; $U_D/D_{\text{min}}=0.398-0.447$; $U_D/D_{\text{max}}=0.318-0.338$. Ranges are based on the average measurements of the holotype and paratypes.

Material.— Circa 100 specimens.

Type locality.— DSDP Site 463 (equatorial Central Pacific Ocean, Mid-Pacific Mountains); geographical coordinates: 21°21'N and 174°40'E.

Type level.— Lower Campanian sediments (*G. elevata* Biozone equivalent), Sample 62-463-26-2, 52-54 cm.

Derivation.— Species name is derived from the planispirally coiled test.

Diagnosis.— *Hillsella* with planispiral test.

Description.— Test is planispiral, consisting of 10 to 12 globular chambers that increase in size as added; there are five to seven chambers in the final whorl. Sutures are distinct, radial and depressed. Test shape is biconvex and symmetrical; periphery is broadly rounded, occasionally with a wide imperforate band on the earlier chambers of the final whorl. Main aperture is a low to medium high arch, in peripheral-equatorial position at the base of the last formed chamber. Umbilicus is shallow and variable in size, its diameter representing one fourth to one third of the maximum test diameter; relict apertures occur in the umbilicus. Chamber surface is ornamented with small and scattered pustules, 3.1-6.0 μm in diameter. Test wall is calcitic hyaline, simple-ridged and perforate; pores are cir-

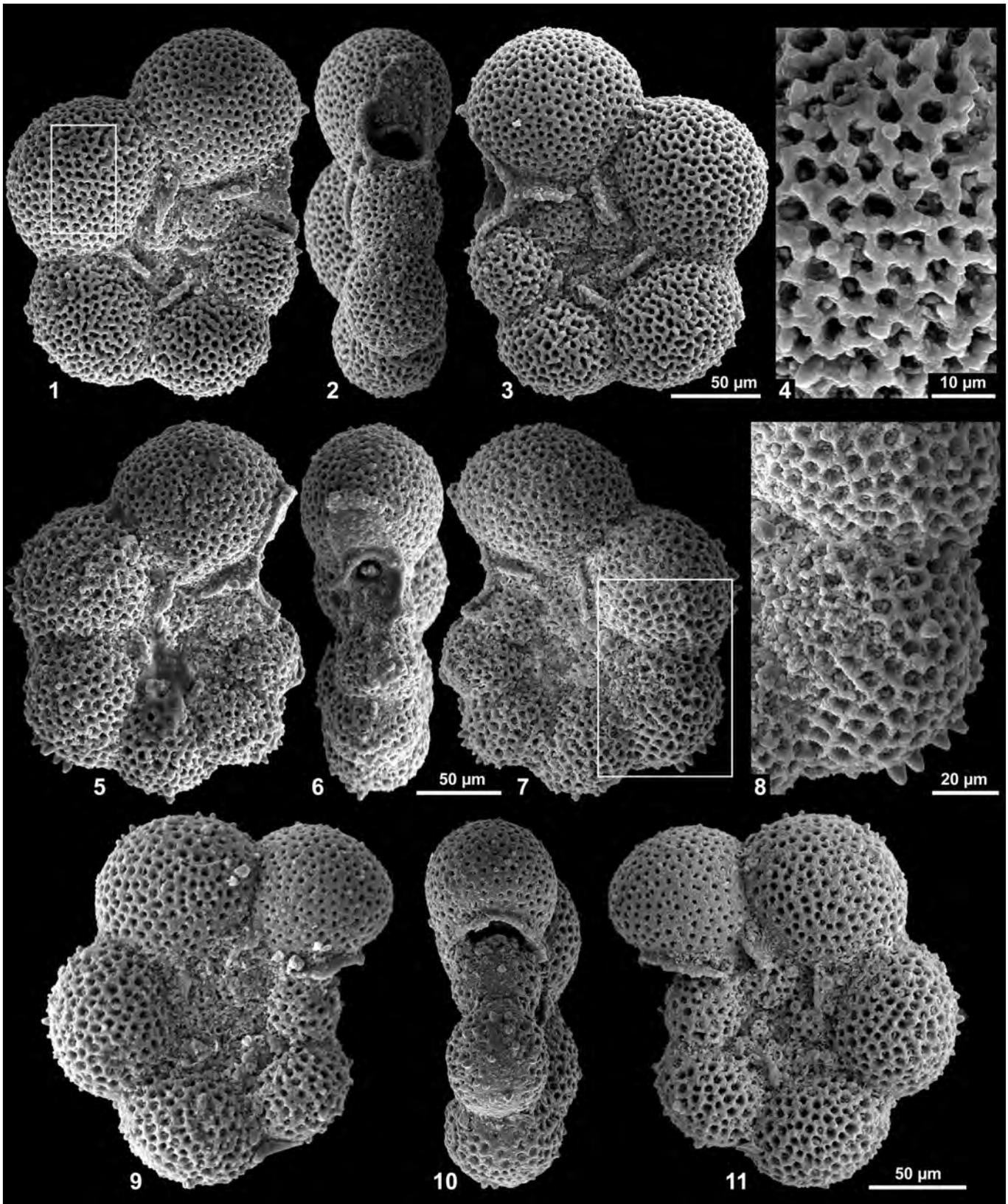


Plate 4. Specimens of *Hillsella planispiralis* n.sp. from the lower Campanian sediments (equivalent of the *G. elevata* Biozone) at the type locality, DSDP Site 463 (equatorial Central Pacific Ocean, Mid-Pacific Mountains). 1-4, Holotype, Sample 62-463-26-2, 52-54 cm. 5-8, Paratype, Sample 62-463-26-2, 52-54 cm. 9-11, Topotype, Sample 62-463-26-3, 52-54 cm.

cular to subpolygonal in shape, with diameters of 2.8-5.2 µm.

Remarks.— *Hillsella planispiralis* differs from other species with a planispiral coil by having simple-ridged test wall. It differs from *H. hillsi* and *H. elongans* by having planispi-

ral rather than low trochospiral test and peripheral–equatorial rather than extraumbilical-peripheral main aperture. Specimens with intermediate morphological features between the ancestor *H. hillsi* and *H. planispiralis*, its descendant, are known from the lowermost Campanian (Pl. 4, Figs 9-11). These specimens are assigned to *H.*

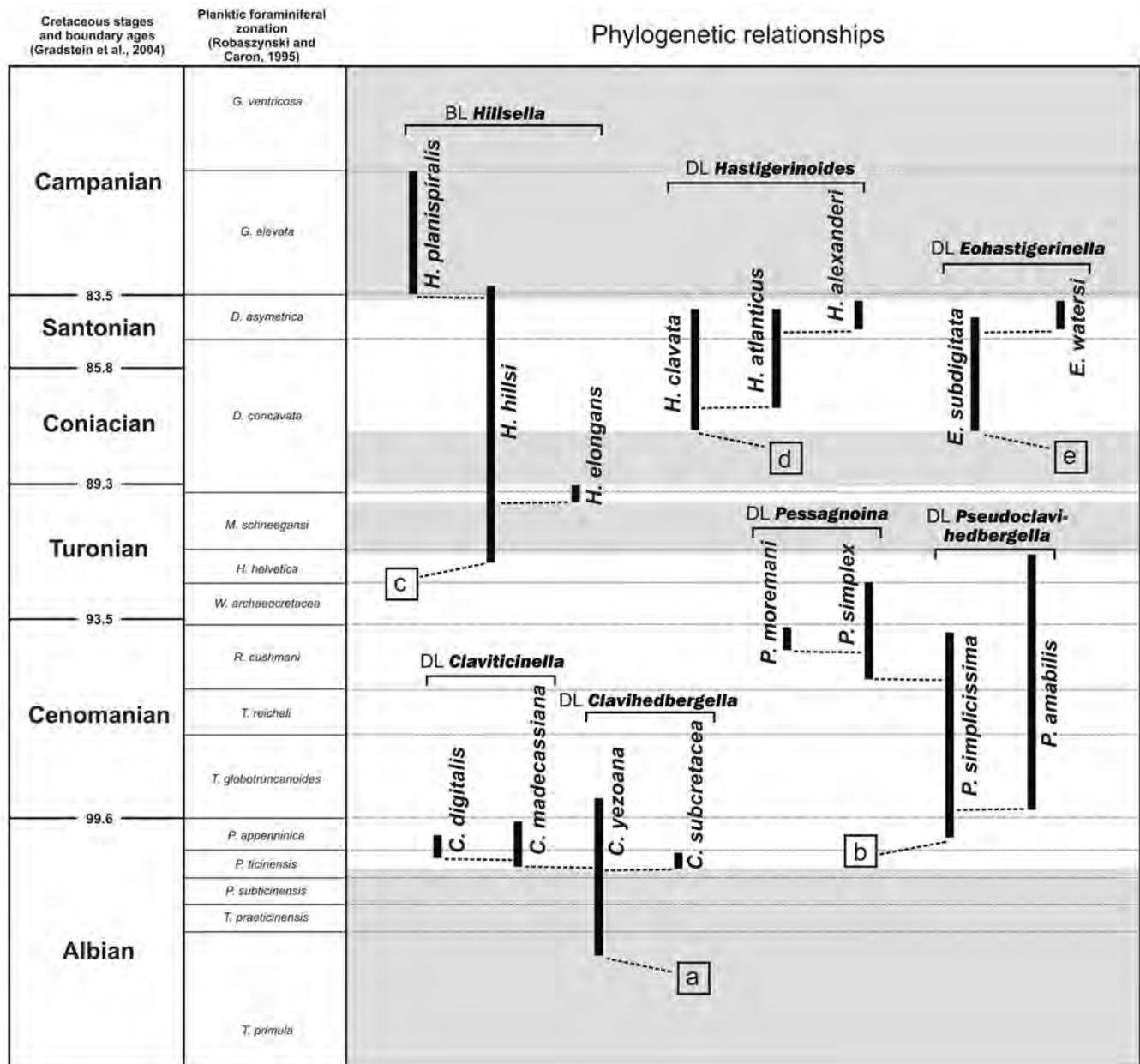


Figure 4. Diagram presenting the multiple origins of the late Albian-Maastrichtian non-schackoinid lineages that developed radially elongate chambers. Stratigraphic intervals marked with light grey indicate the absence from the fossil record of non-schackoinid species with radially elongate chambers. Lineage kinds: DL-directional lineage, BL-branched lineage. Lineage origins: a-the ancestor of *Clavihedbergella yezoana* (Takayanagi and Iwamoto, 1962) is not precisely known, but probably is among the smooth species of the hedbergellid stock (Georgescu, 2009b); b-*Pseudoclavihedbergella* evolved from *Hedbergella delrioensis* (Carsey, 1926), a species with pustulose ornamentation (Georgescu, 2009b); c-the ancestor of the *Hillsella* is probably *H. planispira*; d-*Hastigerinoides* Brönnimann, 1952 evolved from the globular-chambered planispiral species *Globigerinelloides escheri* (Kaufmann in Heer, 1865) (Georgescu and Huber, 2008); e-the ancestral species of the *Eohastigerinella* Morozova, 1957 directional lineage is unknown.

planispiralis on the basis of the quasi-planispiral test and peripheral-equatorial main aperture. These tests are monoubiliculate, with relict apertures on the umbilical side only and occasionally exhibit imperforate peripheral band on the earlier chambers of the final whorl.

Occurrence.— Lower Campanian of the Mid-Pacific Mountains.

4. CONCLUSIONS AND DISCUSSION

Hillsella is redefined in the evolutionary classification as a branched lineage of Turonian-early Campanian age (Fig. 5). The species of this branched lineage are characterized by simple-ridged test walls and exhibit significant variability with respect to the chamber shape and test coiling. The evolutionary history of *Hillsella* shows that it initiated with a trochospiral species, *H. hillsi*, in the early Turonian, namely the upper part of the *H. helvetica* Biozone.

Hillsella hillsi is a long ranging species (approximately 7.1 m.y.), its last occurrence being in the lower Campanian sediments of DSDP 463, immediately above the Santonian/Campanian boundary. Its ancestor is not precisely known but the low trochospire and test general aspect seemingly point out towards *Hedbergella planispira* (Tappan, 1940), a small species of the upper Albian-Turonian with ornamentation consisting of pustules (Fig. 6). Last chamber ornamentation and porosity features of some specimens of *H. hillsi* (Pl. 4, Figs 5, 7) apparently indicate that the ancestral species was probably a smooth or pustulose hedbergellid with simple test wall and pores with diameter of 1.4-2.2 μm , features that match those of *H. planispira*. The similarities between the ornamentation and porosity features of the ancestral species and those of the last formed chamber in its descendant were demonstrated in various lineages of Cretaceous planktic foraminifera (Georgescu, 2007, 2009a, 2009b, 2009c, 2010a, 2010b; Georgescu and Huber, 2008, 2009; Georgescu and Abramovich, 2008, 2009; Georgescu et al., 2008).

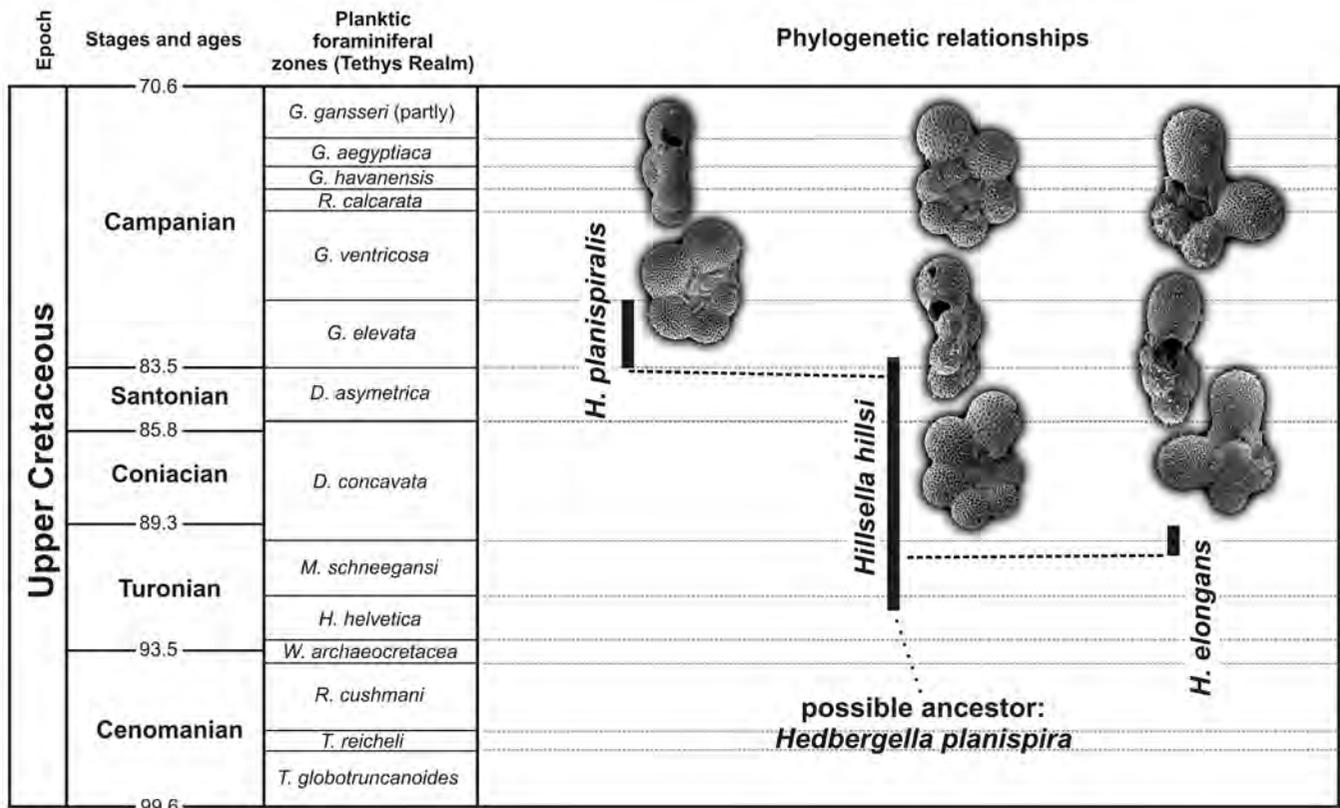


Figure 5. Evolutionary relationships between the species with simple-ridged test wall of the *Hillsella* branched lineage.

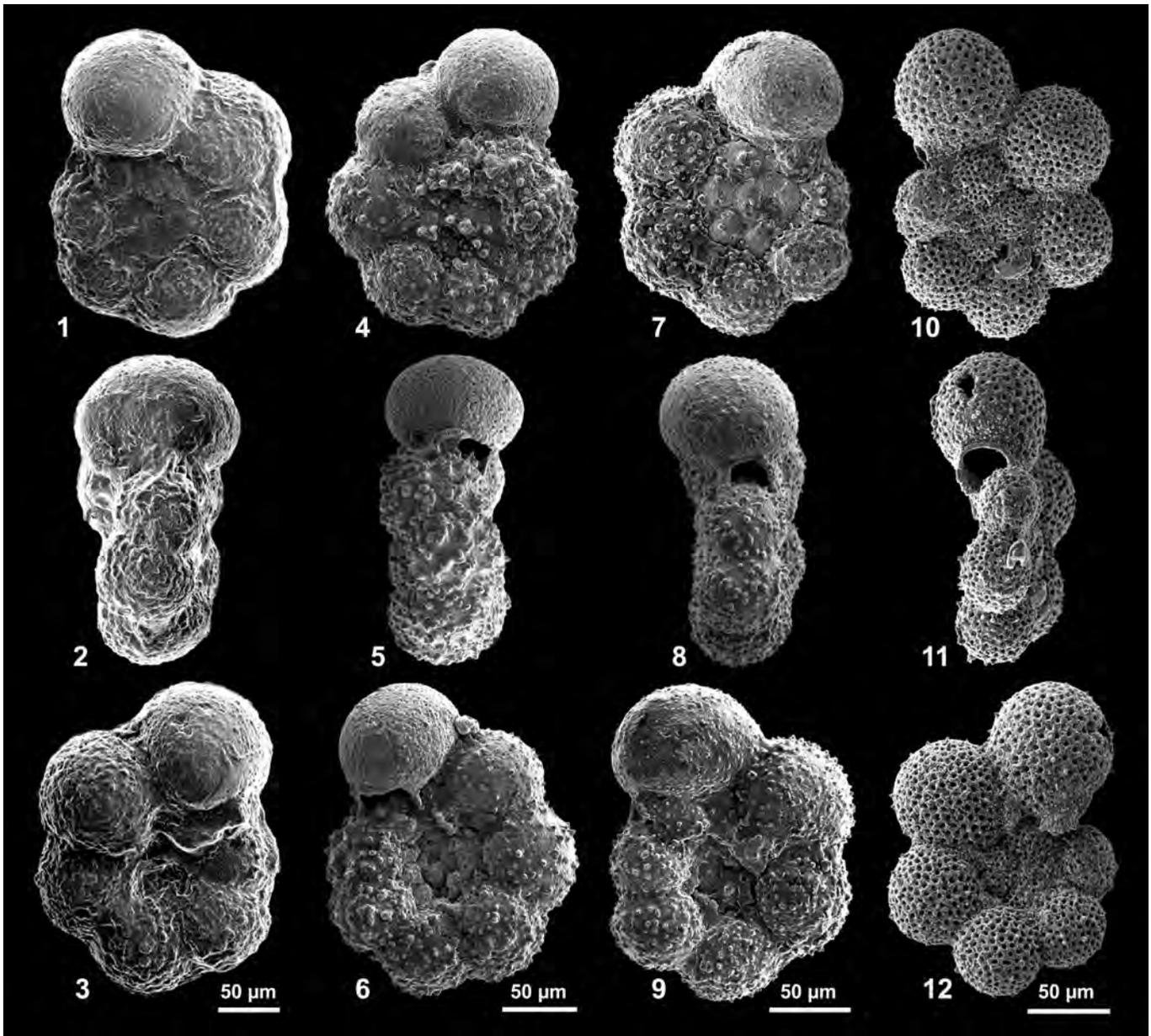


Figure 6. Specimens of *Hedbergella planispira* and *Hillsella hillsii*, the former being considered the presumed ancestor of the latter. Note the similarities in the test general appearance, chamber shape, test shape in edge view, umbilical size, etc. 1-3. Holotype of *H. planispira* (USNM 25113) from the Grayson Bluff of Texas, originally figured by Tappan (1940, pl. 19, fig. 12); ESEM illustrations from the CHRONOS Taxonomic Database (www.chronos.org). 4-6. Hypotype of *H. planispira* from the Britton Clay (Dallas County, Texas), Loeblich and Tappan Topotype Collection, NMNH (USNM 473262). 7-9. Hypotype of *H. planispira* from the Britton Clay (Dallas County, Texas), Loeblich and Tappan Topotype Collection, NMNH (USNM 473262). 10-12. Holotype of *H. hillsii*, Sample 10-150-10-1, 58-73 cm. Note the similarities in the gross test architecture (e.g., trochospire height, chamber shape and side increase rate, aperture shape and position, etc) and ornamentation consisting of scattered pustules; the test wall is simple in *H. planispira* (1-9) and simple-ridged in *H. hillsii*.

Two adaptive radiations from *H. hillsii* are herein recognized, namely *H. elongans* n. sp. from the late Turonian and *H. planispiralis* n. sp. from the early Campanian. The former is a short range species (0.3-0.4 m.y.) characterized by perpendicularly to tangentially elongate chambers, occasionally with distal bulbous projection; it is the

only Cretaceous planktic foraminiferal species with last formed elongate chambers and simple-ridged test wall. The latter is recorded only in the lower Campanian sediments of the equatorial Central Pacific and is characterized by planispiral coil and simple-ridged test wall; its stratigraphic range spans approximately 2.2-2.3 m.y.

The new occurrences of *Hillsella* result in better understanding the evolutionary history of this group of planktic foraminifera, allowing upgrading it from typological monospecific genus (Georgescu, 2008a) to branched lineage, a unit with significance in evolutionary classification. Homogeneity in test ultrastructure and porosity, the stratigraphical ranges of the three species of the branched lineage and occurrences of specimens with intermediate morphological features between *H. hillsi* and *H. elongans* in the late Turonian and *H. hillsi* and *H. planispiralis* in the early Campanian suggest that the group is a monophyletic supraspecific unit. This conclusion is also supported by uniqueness of the simple-ridged wall of *H. elongans*, among the non-schackoinid Cretaceous planktics with elongate chambers in the adult, and *H. planispiralis* among the planispirally coiled taxa.

The evolutionary relationships between the three species (Fig. 5) question the classical Linnaean classification, in which accepting the planispiral and trochospiral tests should be separated at the superfamily level and the trochospiral and globular chambered species and those with elongate chambers in the adult, should be included into distinct subfamilies. A natural classification framework, that takes into consideration the resemblances between the species of a monophyletic group as the result of common ancestry and the differences as a result of evolutionary process in all of its aspects (e.g., morphology, stratigraphical range, paleoecology, etc), can be achieved only by grouping the composite paleontological species according to the data derived from the stratigraphical record. Application of the classical Linnaean principles of classification in the case of *Hillsella* would result in an unnatural classification framework. *Hillsella hillsi*, the stem species of the branched lineage, would be classified as: genus *Hillsella* Georgescu, 2008a, subfamily Hedbergellinae Loeblich and Tappan, 1961, family Hedbergellidae Loeblich and Tappan, 1961, superfamily Rotaliporacea Sigal, 1958. Its first descendant species, *H. elongans* would be included into the genus *Clavihedbergella* Banner and Blow, 1959, subfamily Ticinellinae Longoria, 1974, family Rotaliporidae Sigal, 1958, superfamily Rotaliporacea Sigal, 1958. The second descendant, *H. planispiralis*, would be classified into genus *Globigerinelloides* Cushman and ten Dam, 1948, subfamily Globigerinelloidinae Longoria, 1974, family Globigerinelloididae Longoria, 1974, superfamily Planomalinae Bolli *et al.*, 1957. This hypothetical clas-

sification shows that the classical Linnaean units, although objective, have small chances to be natural, namely consisting of monophyletic taxa that can document the evolution process (Georgescu, 2009a, 2009c, 2010a).

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ÍNDICE DE AUTORES, VOLUMEN 44 (2012) *Authors index*

Borghi, L.	1	Lobato, G.....	1
Carrigy, C.....	79	Melo, J.H.G.	1
Georgescu, M.D.....	79	Playford, G.....	1
Guillem, J.	45	Salazar Ramírez, R.W.....	23
Gutiérrez-Marco, J.C.	23	Sarmiento, G.N.	23
Ibrahim, M.I.A.	57	Tahoun, S.S.	57
Kholeif, S.	78	Usera, J.	45

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- GEOFFREY PLAYFORD, LEONARDO BORGHI, GUSTAVO LOBATO, AND JOSÉ HENRIQUE G. MELO
Palynological dating and correlation of Early Mississippian (Tournaisian) diamictite sections, Parnaíba Basin, northeastern Brazil..... 1-22
- ROSELIS W. SALAZAR RAMÍREZ, GRACIELA N. SARMIENTO Y JUAN CARLOS GUTIÉRREZ-MARCO
Taxonomía y bioestratigrafía de conodontos silúricos en el pliegue acostado del Courel (Lugo, España) [*Taxonomy and biostratigraphy of Silurian conodonts from the Courel recumbent fold (province of Lugo, Spain)*]..... 23-44
- JORDI GUILLEM AND JUAN USERA
A new benthic foraminifer from the Recent coastal lagoon of Torreblanca (Castellón, Western Mediterranean)..... 45-56
- SAMEH S. TAHOUN, MOHAMED I.A. IBRAHIM, AND SUZAN KHOLEIF
Palynology and paleoenvironments of Middle Jurassic to Cenomanian successions, Alamein-IX well, north Western Desert, Egypt..... 57-78
- M. DAN GEORGESCU AND CAITLIN CARRIGY
Evolutionary classification of the coiled Upper Cretaceous (Turonian-Lower Campanian) planktic foraminifera with simple-ridged test wall 79-98



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