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Palynology and isotope geochronology of the Upper Ordovician–Silurian successions (Ghelli and Soltan Maidan Formations) in the Khoshyeilagh area, eastern Alborz Range, northern Iran; stratigraphic and palaeogeographic implications

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A B S T R A C T

Surface samples from the Ghelli and Soltan Maidan Formations in the Khoshyeilagh area of the eastern Alborz Range, northern Iran, were analyzed to determine the age and the stratigraphic relationship of these two units. The samples contained rich palynomorph assemblages, dominated by acritarchs (36 species recognized, distributed among 28 genera) and chitinozoans (nine species identified, distributed among seven genera). Cryptospores and scolecodonts as well as a few graptolite remains were also observed, although not studied in detail. Based on the restricted stratigraphic range of chitinozoan and acritarch species, a Late Ordovician (late Katian–Hirnantian) age is assigned to the Ghelli Formation. Palynological samples from the Soltan Maidan Formation yielded acritarch assemblages characterized by species commonly found in Upper Ordovician sediments together with typical middle Silurian forms. Considering reworking of the Upper Ordovician species, the age of the investigated part of the Soltan Maidan Formation is not younger than Gorstian (early Ludlow, early late Silurian). U–Pb zircon ages of 434.4 ± 6.4 Ma (Telychian, late Llandovery, late early Silurian) obtained for granitic clasts collected at the base of the Soltan Maidan Formation, are consistent with the inferred palynological age and indicate that granite emplacement, cooling, exhumation, erosion, transportation and deposition of its clasts took place in a time interval estimated to be 5 to 10 million year long.

Based on the presence of diagnostic chitinozoan taxa, the Ghelli Formation can be assigned to the *Armoricochitina nigerica*, *Ancyrochitina merga*, and *Spinachitina oulebsiri* chitinozoan biozones. Chitinozoan assemblages reflect a clear palaeobiogeographic affinity with the previously defined 'North Gondwana Domain'. The composition of acritarch assemblages also appears to be consistent with newly proposed hypotheses of a Late Ordovician phytoplanktonic biogeographical differentiation between a Laurentian/Baltica realm, and a Gondwanan realm to which the present assemblages belong. The presence of cryptospores together with taxonomically diverse chitinozoan, acritarch, and scolecodont assemblages, suggest a relatively shallow marine, platformal depositional environment for the Ghelli Formation.

1. Introduction

Palaeozoic sedimentary rocks crop out extensively and with great thickness in the Alborz Mountain Range of northern Iran (Fig. 1), but have received minimal biostratigraphic attention during the past decades because of the paucity of marine macrofossils. Although general stratigraphic schemes for the Palaeozoic of the Alborz Range have been proposed (e.g., Afshar-Harb, 1975, 1994), the stratigraphic relationships among the Lower Palaeozoic formations, and thus the overall stratigraphic architecture of the Alborz Mountain Range, are imprecisely known because of the lack of detailed biostratigraphic evidence. In recent years, the use of organic-walled microfossils (acritarchs, chitinozoans, miospores) has proved of primary importance for the accurate biostratigraphic dating and mapping of the Iranian Palaeozoic formations and for the analysis of their palaeogeographic affinities (e.g., Ghavidel-Syooki and Vecoli, 2007, 2008). In this paper, we focus on the stratigraphic relationships and biostratigraphic (palynological) age of the Ordovician–Silurian sedimentary successions of the Khoshyeilagh area, which is transitional between the Eastern Alborz Range, where Late Ordovician and early Silurian deposits are well developed and exposed, and the Western Alborz Range, where no rocks of this age are present. Additionally, the results will contribute to the geographic documentation of Ordovician and Silurian palynomorphs (acritarchs and chitinozoans) in the still little investigated Iranian region, and to the knowledge of palaeobiogeographic distribution of Ordovician organic-walled microplankton.

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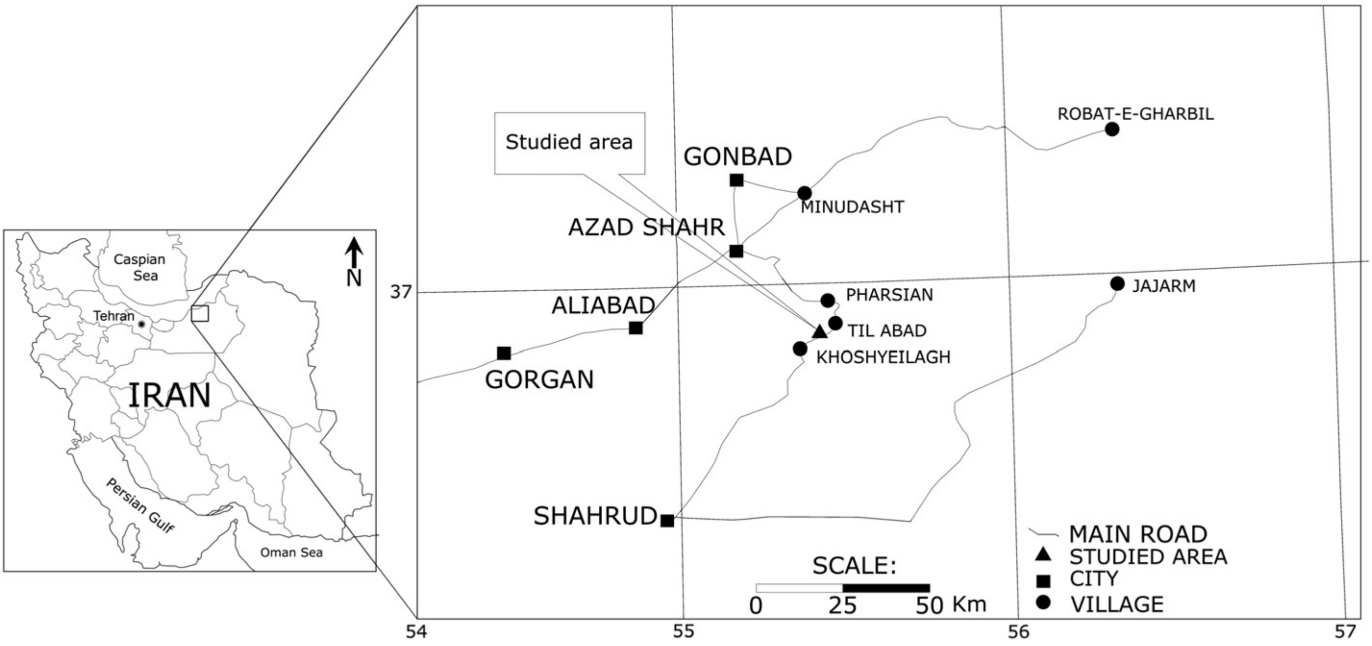

Fig. 1. Geographic setting and location of study section.

2. Regional stratigraphic setting

The Palaeozoic strata exposed in the Khoshyeilagh Pass consist of, in ascending stratigraphic order, the Ghelli (Ordovician), Soltan Maidan (Silurian?), Padeha (Upper Devonian), Khoshyeilagh (Upper Devonian), and Mobarak (Lower Carboniferous) formations. The biostratigraphic age of Upper Palaeozoic rock units (Padeha, Khoshyeilagh, and Mobarak formations) has been well established, whereas the Lower Palaeozoic rock units (Ghelli and Soltan Maidan Formations) are the subjects of major controversy because of their lack of diagnostic faunal elements.

The Ordovician sediments of the Khoshyeilagh area are known by different names such as the Shirgesht Formation (Stampfli, 1978), the Lashkarak Formation (Bozorgnia, 1973), and the Ghelli Formation (Afshar-Harb, 1994). Both the Lashkarak and Shirgesht Formations have been dated as Early Ordovician (Bozorgnia, 1973; Ghavidel-Syooki, 1995, 2000a, 2001) whereas the Ghelli Formation has been considered Middle–Late Ordovician in age at its type locality (Ghavidel-Syooki, 1997, 2000a, 2001; Ghavidel-Syooki and Winchester-Seeto, 2002). Furthermore, the Upper Ordovician Gorgan Schists have been suggested to be time-equivalent of the Ghelli Formation (Ghavidel-Syooki, 2008), notwithstanding Zanchi et al. (2009) assertion that "no thick terrigenous successions similar to the one forming the Gorgan Schists are known to the south, preventing a direct correlation with the Paleozoic units of the Iranian margin" (Zanchi et al., 2009, p. 50).

There is no agreement on the age of the Soltan Maidan Formation, which consists mainly of trachybasalt–trachyandesite rocks. Previous K–Ar dating of the Soltan Maidan volcanics (Jenny, 1977) failed because of pervasive hydrothermal alteration and/or metamorphism. However Jenny (1977) has tentatively assigned an undifferentiated Silurian age to the Soltan Maidan Formation based on its position between the Ghelli and Padeha Formations. Because of the uncertainty of the age assignments, both the Ghelli and Soltan Maidan Formations needed to be reinvestigated, in order to more precisely determine their age.

The Ghelli Formation is 301 m thick in the study area and consists mainly of olive-gray, silty shale and gray shale (Fig. 2); it is disconformably overlain by the Soltan Maidan Formation. The basal part of the Ghelli Formation consists of brown, rounded, pebbly conglomerate and red-purple shales, which do not display a clear contact with the underlying sediments due to the presence of basalt. In the Khoshyeilagh area, the Ghelli Formation contains trace fossils and symmetrical ripple marks in some intervals, but it lacks a marine macrofauna (Bozorgnia, 1973; Afshar-Harb, 1994). At the measured section along the Khoshyeilagh Pass, the Soltan Maidan Formation overlies the Ghelli Formation with a 7 m-thick basal polygenic conglomerate at its base, which is overlain by a 3 m sequence of siliciclastic sedimentary rocks and in turn by 67 m of volcanic lavas (alternation of trachybasalt and trachyandesite flows). A 2.7 m-thick conglomerate lithologically similar to that at the base of the Soltan Maidan Formation is present about 77 m from the base of the section, and it is succeeded by about 630 m of lavas (Fig. 2). Owing to the absence of macrofossils, no palaeontologic age has previously been provided. However, because of being overlain by the Devonian Padeha Formation, Jenny (1977) tentatively attributed a Silurian age to the Soltan Maidan Formation. Jenny (1977) also attempted K–Ar dating on seven samples from the Soltan Maidan volcanics attaining a wide range of isotopic ages from 501 to 173 Ma due to low-grade metamorphism/hydrothermal alteration indicated by the common occurrence of subsolidus phases including chlorite, calcite, hematite and epidote.

3. Materials and methods

3.1. Palynology

Palynological analysis was carried out on 67 outcrop samples (63 from the Ghelli Formation and four from the sedimentary portion of the basal part of the Soltan Maidan Formation; Fig. 2). Each sample was assigned a National Iranian Oil Company Code number with the prefix MG, and includes the samples MG-8204 to MG-8270.

Palynomorphs were extracted from shale and siltstone samples using the standard palynological technique of treatment in HCl and HF to remove carbonates and silicates respectively, and neutralizing the residues in distilled water after each acid treatment. Samples were not oxidized, and the organic residues were concentrated using density separation in a zinc bromide solution with a specific gravity of 1.95. The organic residue was then sieved through a 15 µm nylon mesh sieve in order to eliminate the finer debris and facilitate palynological analyses. Palynological preparations were then studied using transmitted light and scanning electron microscopy.

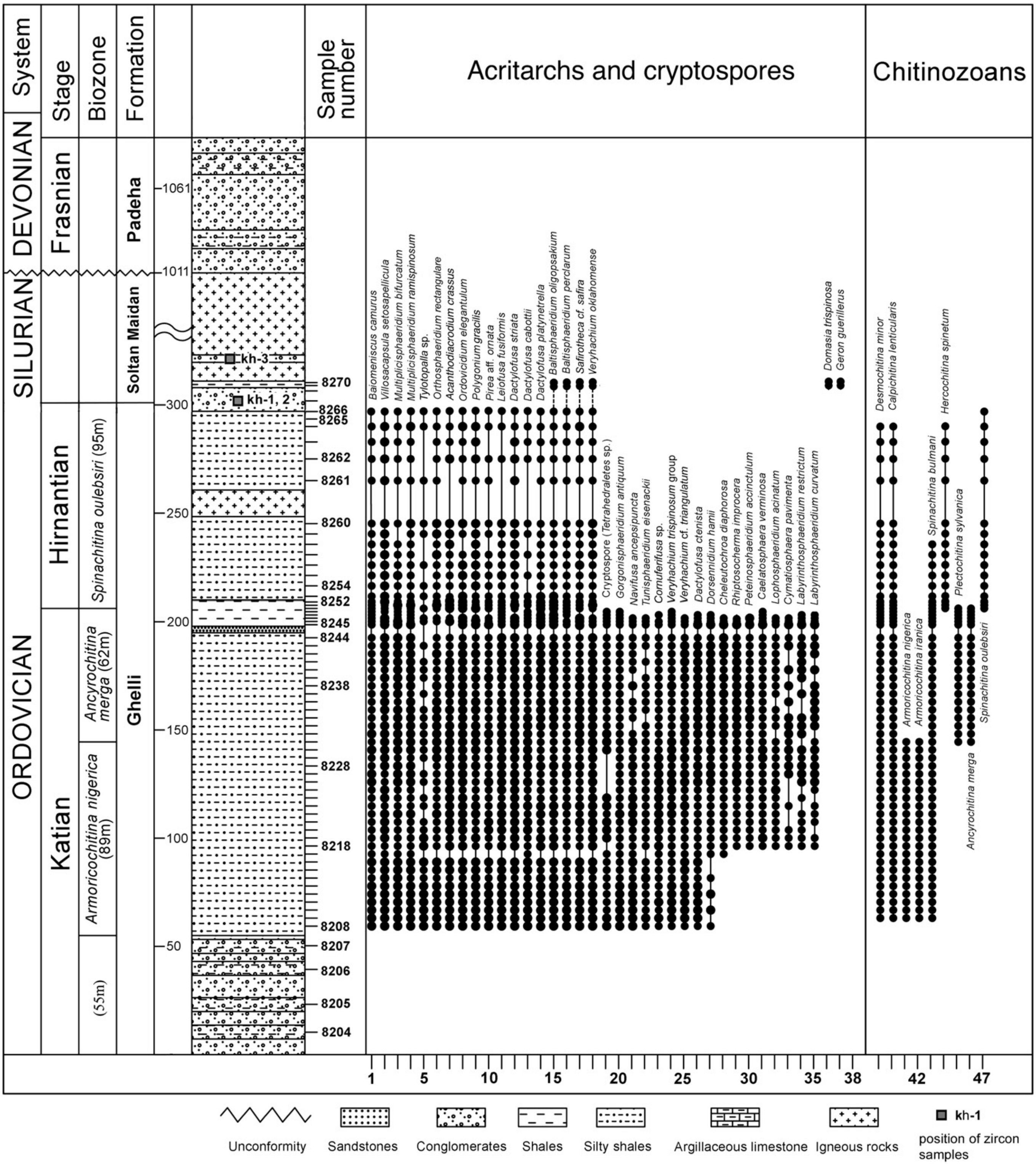

Fig. 2. Lithostratigraphy, stratigraphic distribution of palynomorphs, biozonation, and chronostratigraphic attribution of the study section.

All slides used in this study are housed in the palaeontological collections of the Exploration Directorate of the National Iranian Oil Company under the sample numbers MG-8204 to MG-8287 (only MG-8204-MG-8267 for palynological study).

3.2. Isotope geochronology

Zircon U–Pb geochronology used in the present study is the preferred method for determining crystallization ages of altered igneous rocks due to the closure temperature of the U–Pb system well in excess of the solidus of granitic melts (>700–750 °C) and the resistance of zircon in hydrothermal alteration and metamorphism. Unfortunately, Soltan Maidan volcanics have so far not yielded zircons. However, the conglomerates at the base and in the middle of this volcanic unit are dominated by rounded granitic clasts, which belong to a shallow intrusion of pink to red microgranite showing granophyric microtextures. To supplement this study, granite cobbles were collected from both conglomerate horizons for zircon separation

Table 1
Sample information for the two granite cobbles from conglomerate horizons in the Soltan Maidan Formation dated in this study.

No.	Locality Sample ID	GPS coordinates N (deg, min)	E (deg, min)	Elevation (m)	Rock type	Age (Ma)‡
Khosh Yeilagh Pass, Eastern Alborz mountains, NE Shahrud, Iran						
1	Kh 1–2					439
		36° 49.595′	55° 20.593′	1920–1960	Pink to red microgranites	
2	Kh-3					427

‡ See Table 6 for details.

and U–Pb age determinations. Table 1 provides the basic sample information and Table 2 lists the relevant isotopic analyses performed on two granite samples collected from conglomerates of the Soltan Maidan Formation (samples Kh 1–2, and Kh 3; Fig. 2).

The reported ion-microprobe zircon U–Pb ages are based on U, Pb, and Th isotopic measurements employing spot analysis using the Cameca IMS 1270 ion microprobe at the University of California, Los Angeles (UCLA). Analytical methods follow the procedures described by Quidelleur et al. (1997) and Schmitt et al. (2003a,b). Following mineral separation using conventional liquid and magnetic techniques, zircon grains were hand selected, mounted in epoxy, and coated with ~100 Å thickness of gold. The ion-microprobe spot analyses utilized a primary ion beam focused to an ~30 μm diameter spot and a secondary ion beam with a mass resolving power of 5000 and an energy window of 50 eV. Following a pre-sputtering period of ~180 s, each analysis collected data for 8–10 cycles. The sample chamber was flooded with oxygen at ~3×10^{-5} Torr to enhance secondary Pb^{+} ionization. The reported weighted-mean ages are based on ^{206}Pb/^{238}U ages calculated using zircon standard AS3 (1099 ± 0.5 Ma; Paces and Miller, 1993). Common lead corrections were made using the measured values of ^{204}Pb (Stacey and Kramers, 1975) and the values of ^{208}Pb corrected for ^{232}Th-derived ^{208}Pb (Compston et al., 1984), which are considered a proxy for common ^{206}Pb and ^{207}Pb. These corrections use the anthropogenic Pb compositions reported for the Los Angeles basin (Sanudo-Wilhelmy and Flegal, 1994). All age uncertainties are reported at the 1σ level (Table 2). As discussed below, most grains yield concordant ages, with <12% discordance.

4. Palynostratigraphy

Most of the samples examined contained well-preserved and abundant palynomorphs (e.g., acritarchs, chitinozoans, scolecodonts, cryptospores), with the exception of those collected from the basal part (first 55 m) of the Ghelli Formation, which is barren, consisting of a succession of conglomerate, sandstone, and purple shale.

In general, the acritarchs proved more abundant than the other palynomorphs. Palynomorphs and other organic debris range in color from yellow to orange brown, indicating a relatively high thermal maturity for the organic material in the Lower Palaeozoic strata in this part of the Alborz Mountain Range. A total of 36 acritarch species assigned to 28 genera, and nine chitinozoan species assigned to seven genera have been observed (Fig. 2; Plates I–VIII). In addition, rare cryptospores (e.g., *Tetrahedraletes*) and scolecodonts were also encountered, which will not be detailed in the present study. It is interesting to note that Ordovician cryptospores were previously unreported from Iran. The chitinozoans permitted the recognition of three biozones, which are well known in the so-called "North Gondwanan Domain" (Paris, 1990, 1996); these are discussed below in ascending stratigraphic order, and the biostratigraphic age of the study sequence is primarily based on the chitinozoan biozonation (Paris, 1990, 1996; Paris et al., 2000a). The chitinozoan taxa discussed herein are widely known and detailed in previous publications (e.g., Paris et al., 2000a,b; Ghavidel-Syooki and Winchester-Seeto, 2002; Ghavidel-Syooki, 2008); they are listed below according to the grouping of the classification scheme of Paris et al. (1999), which is followed herein. The acritarchs discussed and illustrated herein are treated as form species and form genera under provisions of the International Code of Botanical Nomenclature (I.C.B.N.) (Greuter et al., 1994), and are arranged alphabetically by genera under the informal *incertae sedis* "acritarch" group.

4.1. Chitinozoans

The following chitinozoan taxa were identified:

Order Operculatifera Eisenack, 1931
Family Desmochitinidae Eisenack, 1931 emend. Paris, 1981

Table 2
U–Pb zircon geochronology on Soltan Maidan Formation conglomerates, Khoshyeilagh Pass, East Alborz Mountain Range, northern Iran.

ID	^{206}Pb*/^{238}U	^{206}Pb*/^{238}U 1_	^{207}Pb*/^{235}U	^{207}Pb*/^{235}U 1_	^{207}Pb*/^{206}Pb	^{207}Pb*/^{206}Pb 1_	% Radio-genic 206Pb	Age (Ma) ^{206}Pb/^{238}U	Age (Ma) ^{206}Pb/^{238}U 1_	Age (Ma) ^{207}Pb/^{235}U	Age (Ma) ^{207}Pb/^{235}U 1_	Age (Ma) ^{207}Pb/^{206}U	Age (Ma) ^{207}Pb/^{206}U 1_	U (ppm)	Th/U
Kh 1–2, pink microgranite cobble in basal conglomerate															
g1	0,06833	0,00344	0,5225	0,0398	0,05546	0,00286	99,3	426,1	20,7	426,8	26,6	430,8	115,0	130	0,59
g4	0,06939	0,00298	0,5351	0,0363	0,05593	0,00267	99,3	432,5	18,0	435,2	24,0	449,6	106,0	159	0,62
g6	0,07043	0,00327	0,525	0,0354	0,05406	0,00244	99,0	438,8	19,7	428,5	23,6	373,4	102,0	193	0,76
g3	0,07117	0,0032	0,6199	0,0457	0,06317	0,00308	99,7	443,2	19,2	489,8	28,6	713,9	104,0	157	0,60
g5	0,07153	0,00297	0,5144	0,0312	0,05216	0,00226	99,2	445,4	17,9	421,4	20,9	292,3	99,1	269	0,66
g2	0,07228	0,00421	0,5366	0,0407	0,05384	0,00227	99,3	449,9	25,3	436,2	26,9	364,5	95,1	215	0,69
mean age*: 439.3 ± 8.8 Ma (n = 6)															
weighted mean age* = 438.9 ± 8.1 Ma, MSWD = 0.2 (n = 6)															
kh-3, pink microgranite cobble in middle conglomerate															
g3	0,06694	0,00284	0,4912	0,0359	0,05322	0,00309	98,7	417,7	17,1	405,7	24,4	338,2	132,0	162	0,56
g1	0,06826	0,00326	0,5293	0,0412	0,05623	0,00305	99,3	425,7	19,7	431,3	27,3	461,6	120,0	148	0,81
g2	0,07014	0,00298	0,4835	0,042	0,04999	0,00355	98,7	437,0	17,9	400,5	28,7	194,7	165,0	153	0,62
mean age*: 426.8 ± 9.7 Ma (n = 3)															
weighted mean age* = 426.6 ± 10.5 Ma, MSWD = 0.3 (n = 3)															
Kh 1–2 and Kh-3 combined															
mean age*: 435.1 ± 10.5 Ma (n = 9)															
weighted mean age* = 434.4 ± 6.4 Ma, MSWD = 0.3 (n = 9)															

* Mean and weighted mean calculations are for ^{206}Pb/^{238}U ages.

Subfamily Desmochitinidae Paris, 1981
Genus *Calpichitina* Wilson and Hedland, 1964
Calpichitina lenticularis (Bouché, 1965) (Plate VII, 2)
Genus *Desmochitina* Eisenack, 1931
Desmochitina minor Eisenack, 1931 (Plate VII, 1)

Subfamily Pterochitininae Paris, 1981
Genus *Armoricochitina* Paris, 1981
Armoricochitina iranica Ghavidel-syooki and Winchester-Seeto, 2002 Plate VII, 5; Plate VIII, 1)
Armoricochitina nigerica (Bouché, 1965) (Plate VIII, 2,3)

Subfamily Ancyrochitininae Paris, 1981
Genus *Ancyrochitina* Eisenack, 1955
Ancyrochitina merga Jenkins, 1970 (Plate VIII, 8)
Genus *Plectochitina* Cramer, 1964
Plectochitina sylvanica Jenkins, 1970 (Plate VIII, 9)

Order Prosomatifera Eisenack, 1972
Subfamily Spinachitininae Paris, 1981
Genus *Spinachitina* Schallreuter, 1963 emend. Paris, Grahn, Nestor, and Lakova, 1999
Spinachitina bulmani Jansonius, 1964 (Plate VII, 7)
Spinachitina oulebsiri Paris, Bourahrouh, and Le Hérissé, 2000 (Plate VII, 3,6,8; Plate VIII, 5,6)
Genus *Hercochitina* Jansonius, 1964
Hercochitina sp. cf. *H. spinetum* Melchin and Legault, 1985 (Plate VII, 4; Plate VIII, 7)

The distribution of the above taxa in the study section (Fig. 2), allows recognition of the following biozones:

Armoricochitina nigerica Biozone (Paris, 1990)

This biozone is marked by the first occurrence of *Armoricochitina nigerica* in sample MG-8209 and extends through a thickness of 89 m to sample MG-8231 (Fig. 2). According to Paris (1990, 1996) and also subsequent studies (Bourahrouh et al., 2004) the *nigerica* Biozone is a partial-range biozone corresponding to the stratigraphic interval from the first appearance of the eponymous species up to the first appearance of *Ancyrochitina merga*, the index species of the succeeding biozone. The *nigerica* Biozone is correlated to the upper part of the *complanatus* graptolite zone of the British Standard (Paris, 1990, 1996; Webby et al., 2004), thus corresponding to the early late Katian. Co-occurring chitinozoan taxa in the present assemblage are: *Armoricochitina iranica, Calpichitina lenticularis*, *Desmochitina minor*, and *Spinachitina bulmani*.

Armoricochitina iranica has been previously recorded in Iran from upper Katian to lower Hirnantian strata in the northeastern Alborz Mountain Range (upper part of the Ghelli Formation; Ghavidel-Syooki and Winchester-Seeto, 2002). *Calpichitina lenticularis* is a commonly occurring species in Upper Ordovician stratal sequences of the so-called "Northern Gondwana Domain" (Paris, 1990), including North Africa, southwestern Europe, Turkey, Saudi Arabia and Iran (Paris, 1979, 1981; Elaouad-Debbaj, 1984; Molyneux and Paris, 1985; Al-Hajri, 1995; Oulebsir and Paris, 1995; Ghavidel-Syooki, 2000a,b; Paris et al., 2000a; Ghavidel-Syooki and Winchester-Seeto, 2002). *Desmochitina minor* is a very long-ranging and cosmopolitan species, ranging from the Floian to Katian worldwide (e.g., Grahn, 1984). *Spinachitina bulmani* is a common accessory species of the *nigerica* Biozone in the North Gondwana Domain (Paris, 1990); it is known to occur from the Sandbian of Scotland and Shropshire (Jansonius, 1964; Jenkins, 1970), the Upper Ordovician of Morocco (Elaouad-Debbaj, 1986), the Katian of Anticosti, Québec, Canada (Achab, 1978), Norway (Grahn et al., 1994), and Libya (Molyneux and Paris, 1985).

Ancyrochitina merga Biozone (Paris, 1990)

This biozone has been identified on the basis of the occurrence of the eponymous species between samples MG-8231 and MG-8250, through a total thickness of 62 m within the Ghelli Formation (Fig. 2). The species *Armoricochitina nigerica* last occurs at the base of this zone, in sample MG-8231. According to Paris (1990) and subsequent studies (e.g., Bourahrouh et al., 2004) the *merga* Biozone is an interval-range zone between the FAD of *Ancyrochitina merga* and the FAD of *Tanuchitina elongata*, the index species of the succeeding biozone. Arguments concerning the biostratigraphic age of this biozone have been presented in Paris (1990) and Bourahrouh et al. (2004), supporting a late Katian age. Co-occurring species are *C. lenticularis*, *D. minor*, and *S. bulmani*, which range through the present biozone from the underlying *A. nigerica* Biozone. In addition, *Plectochitina sylvanica* makes its first inception at the base of the present biozone. This latter species is a commonly co-occurring species of the *merga* Biozone in the northern Gondwana region (Elaouad-Debbaj, 1984, 1986; Molyneux and Paris, 1985; Paris, 1990; Al-Hajri, 1995; Ghavidel-Syooki, 2000a,b; Ghavidel-Syooki and Winchester-Seeto, 2002; Bourahrouh et al., 2004). *Hercochitina* sp. cf. *H. spinetum* occurs in the uppermost part of the biozone; *H. spinetum* is a typical Upper Ordovician (Katian) species previously recorded from the Simcoe Group in southern Ontario, Canada (Melchin and Legault, 1985), and the Upper Thumb Mountain Formation, Little Cornwallis Island, Canada (Achab and Asselin, 1995).

Spinachitina oulebsiri Biozone (Paris et al., 2000a)

This chitinozoan biozone has been recognized in the uppermost part of the Ghelli Formation; it coincides with the occurrence of *S. oulebsiri* in samples MG-8250 to MG-8266, through a 95 m-thick stratigraphic interval in the study sequence (Fig. 2). This species has been originally established from records in the Upper Member of the M' Kratta Formation, northeast Algerian Sahara, Bordj Nili area (Paris et al., 2000a), and was indirectly correlated with the *persculptus* graptolite zone, of latest Hirnantian age (Webby et al., 2004). Accordingly, the *oulebsiri* chitinozoan biozone is of particular interest for correlation of latest Ordovician strata and for the identification of the Ordovician–Silurian boundary. Recently, however, Butcher (2009) has questioned the taxonomic validity of this species, and hence the usefulness of the associated biozone, demonstrating that it is impossible to find quantitative criteria to clearly distinguish *Spinachitina oulebsiri* Paris et al., 2000a, from *Spinachitina fragilis* (Nestor, 1980); this latter was then considered by Butcher (2009) as senior synonym of the former. According to Paris (in Paris et al., 2000a) *S. oulebsiri* could be regarded as an early morphotype stage of the *S. fragilis* lineage, characterized by "the increase of the vesicle's length, by a progressive differentiation of the flexure, and by the development of a crown of spines on the margin while the diameter remains more or less stable" (Paris in Paris et al., 2000a, p. 101). Vandenbroucke et al. (2009) also recognized the objective difficulties in differentiating species within the *S. oulebsiri*–*S. fragilis* lineage, especially in moderately preserved material as it is in the case of the present study. However, Vandenbroucke et al. (2009) decided to retain the two species separately, pending more comprehensive analyses. They considered the split between the two morphotypes as being supported by subtle differences (conical vs. cylindrical basal spines, and overall stouter vs. slender chamber appearance in *S. oulebsiri* and *S. fragilis*, respectively). The present material is certainly not adapted to form a basis for a taxonomical analysis and revision.

Based on the overall morphological aspect, we tentatively attribute our specimens to *S. oulebsiri*, and accordingly we correlate the associated strata to the Hirnantian *oulebsiri* biozone. We should also highlight the fact that our specimens are unusually small for *S. oulebsiri*; they fall at the lower end of the dimensional range accepted for the species, in some cases even falling outside the up to

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Plate I. Scale bar 10 μm.

1. *Leiofusa fusiformis* (Eisenack) Eisenack, 1938.
2. *Safirotheca* sp. cf. *S. safira* Vavrdová, 1989.
3,5. *Dactylofusa platynetrella* (Loeblich and Tappan) Fensome, Williams, Barss, Freeman, and Hill, 1990.
4. *Dactylofusa cabottii* (Cramer) Fensome, Williams, Barss, Freeman, and Hill, 1990.
6. *Navifusa ancepsipuncta* Loeblich, 1970 *ex* Eisenack, Cramer, and Díez, 1979.

Plate II. Scale bar 10 μm. (see on page 258)

1. *Pirea* sp. aff. *P. ornata* (Burmann) Eisenack, Cramer, and Díez, 1976.
2,3,5. *Dorsennidium hamii* (Loeblich) Sarjeant and Stancliffe, 1994.
4. *Geron guerillerus* Cramer, 1966 ex Cramer, 1969.
6. *Cheleutochroa diaphorosa* Turner, 1984.
7. *Orthosphaeridium rectangulare* (Eisenack) Eisenack, 1968.

Plate III. Scale bar 10 μm. (see on page 259)

1. *Cornuferifusa sp.*
2. *Veryhachium* sp. cf. *V. triangulatum* Konzalova-Mazancova, 1969.
3. *Tunisphaeridium eisenackii* Loeblich and Tappan, 1978.
4. *Domasia trispinosa* Downie, 1960 emend. Hill, 1974.
5. *Baiomeniscus camurus* Loeblich, 1970.
6. *Dactylofusa striata* (Staplin, Jansonius, and Pocock) Fensome, Williams, Barss, Freeman, and Hill, 1990.
7. *Acanthodiacrodium crassus* (Loeblich and Tappan) Vecoli, 1999.
8. *Cheleutochroa diaphorosa* Turner, 1984.

Plate IV. Scale bar 10 μm. (see on page 260)

1. *Polygonium gracilis* Vavrdová, 1966.
2. *Veryhachium trispinosum* group *sensu* Servais et al., 2007.
3. *Multiplicisphaeridium ramispinosum* Staplin, 1961.
4. *Peteinosphaeridium accinctulum* Wicander, Playford, and Robertson, 1999.
5. *Ordovicidium elegantulum* Tappan and Loeblich, 1971.
6. *Gorgonisphaeridium antiquum* Loeblich and Tappan, 1978.
7. *Lophosphaeridium acinatum* Wicander, Playford, and Robertson, 1999.
8. *Multiplicisphaeridium bifurcatum* Staplin, Jansonius, and Pocock, 1965.

Plate V. Scale bar 10 μm. (see on page 261)

1. *Cymatiosphaera pavimenta* (Deflandre) Deflandre, 1954.
2. *Villosacapsula setosapellicula* (Loeblich) Loeblich and Tappan, 1976.
3,6. *Lophosphaeridium acinatum* Wicander, Playford and Robertson, 1999.
4. *Veryhachium trispinosum* group *sensu* Servais et al., 2007.
5. Cryptospore (*Tetrahedraletes* sp.).
7. *Multiplicisphaeridium* sp. cf. *M. ramispinosum* Staplin, 1961.
8. *Labyrinthosphaeridium curvatum* Uutela and Tynni, 1991.
9,10. *Tylotopalla* sp.

Plate VI. Scale bar 10 μm. (see on page 262)

1. *Dactylofusa ctenista* (Loeblich and Tappan) Fensome, Williams, Barss, Freeman, and Hill, 1990.
2. *Caelatosphaera verminosa* Wicander, Playford, and Robertson, 1999.
3. *Labyrinthosphaeridium restrictum* Uutela and Tynni, 1991.
4. *Veryhachium oklahomense* Loeblich, 1970.
5. *Baltisphaeridium perclarum* Loeblich and Tappan, 1978.
6. *Rhiptosocherma improcera* (Loeblich) Loeblich and Tappan, 1978.
7. *Baltisphaeridium oligopsakium* Loeblich and Tappan, 1978.

Plate VII. Scale bar 10 μm. (see on page 263)

1. *Desmochitina minor* Eisenack, 1931.
2. *Calpichitina lenticularis* (Bouché, 1965).
3,6,8. *Spinachitina oulebsiri* Paris, Bourahrouh, and Le Hérissé, 2000.
4. *Hercochitina* sp. cf. *H. spinetum* Melchin and Legault, 1985.
5. *Armoricochitina iranica* Ghavidel-syooki and Winchester-Seeto, 2002.
7. *Spinachitina bulmani* Jansonius, 1964.

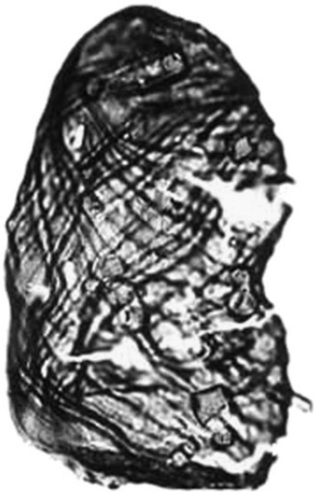
Plate VIII. Scale bar 10 μm. (see on page 264)

1. *Armoricochitina iranica* Ghavidel-syooki and Winchester-Seeto, 2002.
2,3. *Armoricochitina nigerica* Bouché, 1965.
4. Scolecodont.
5,6. *Spinachitina oulebsiri* Paris, Bourahrouh, and Le Hérissé, 2000.
7. *Hercochitina* sp. cf. *H. spinetum* Melchin and Legault, 1985.
8. *Ancyrochitina merga* Jenkins, 1970.
9. *Plectochitina sylvanica* Jenkins, 1970.

1

2

3



4

5

6

Plate I.

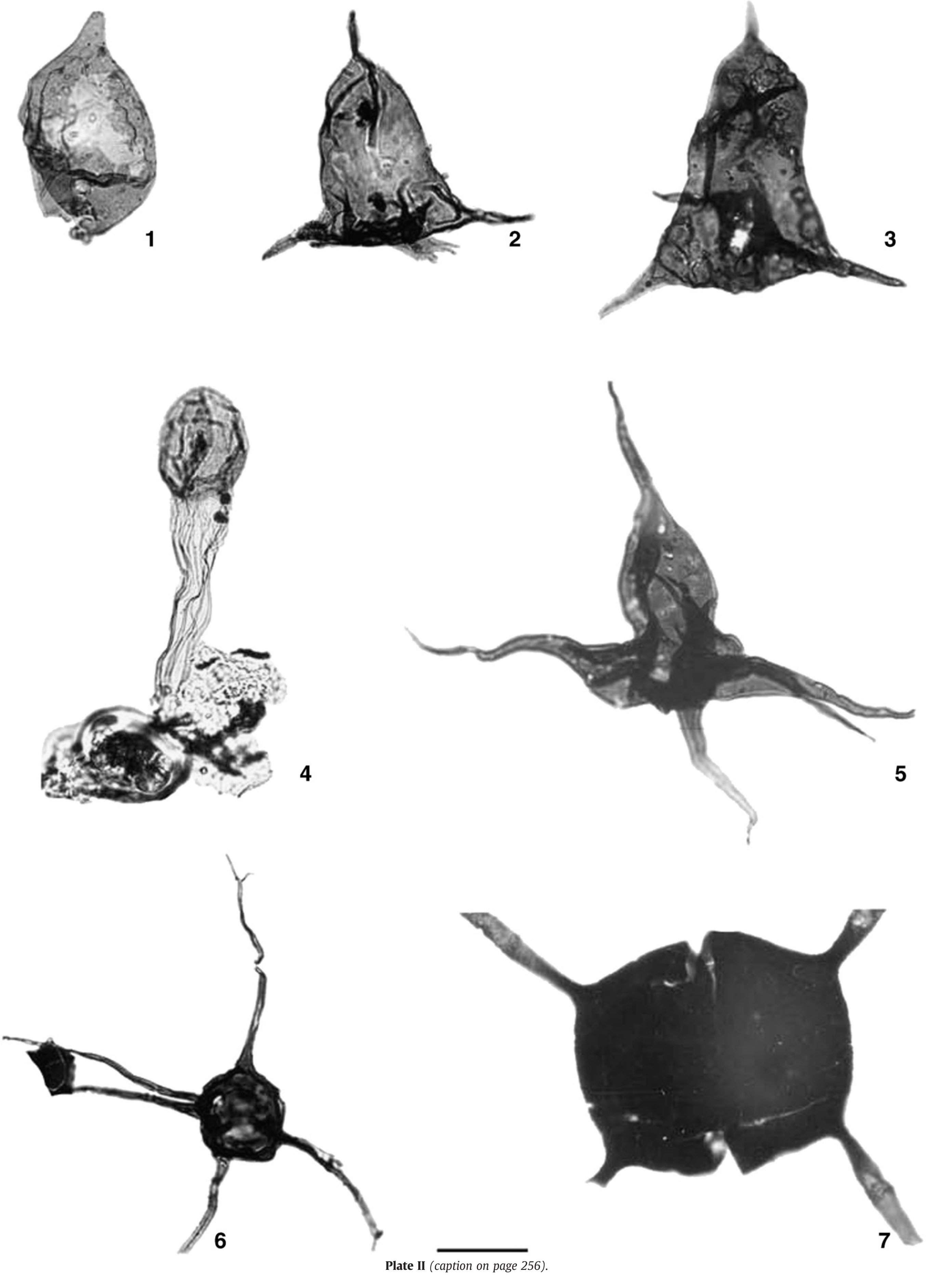


Plate II *(caption on page 256).*

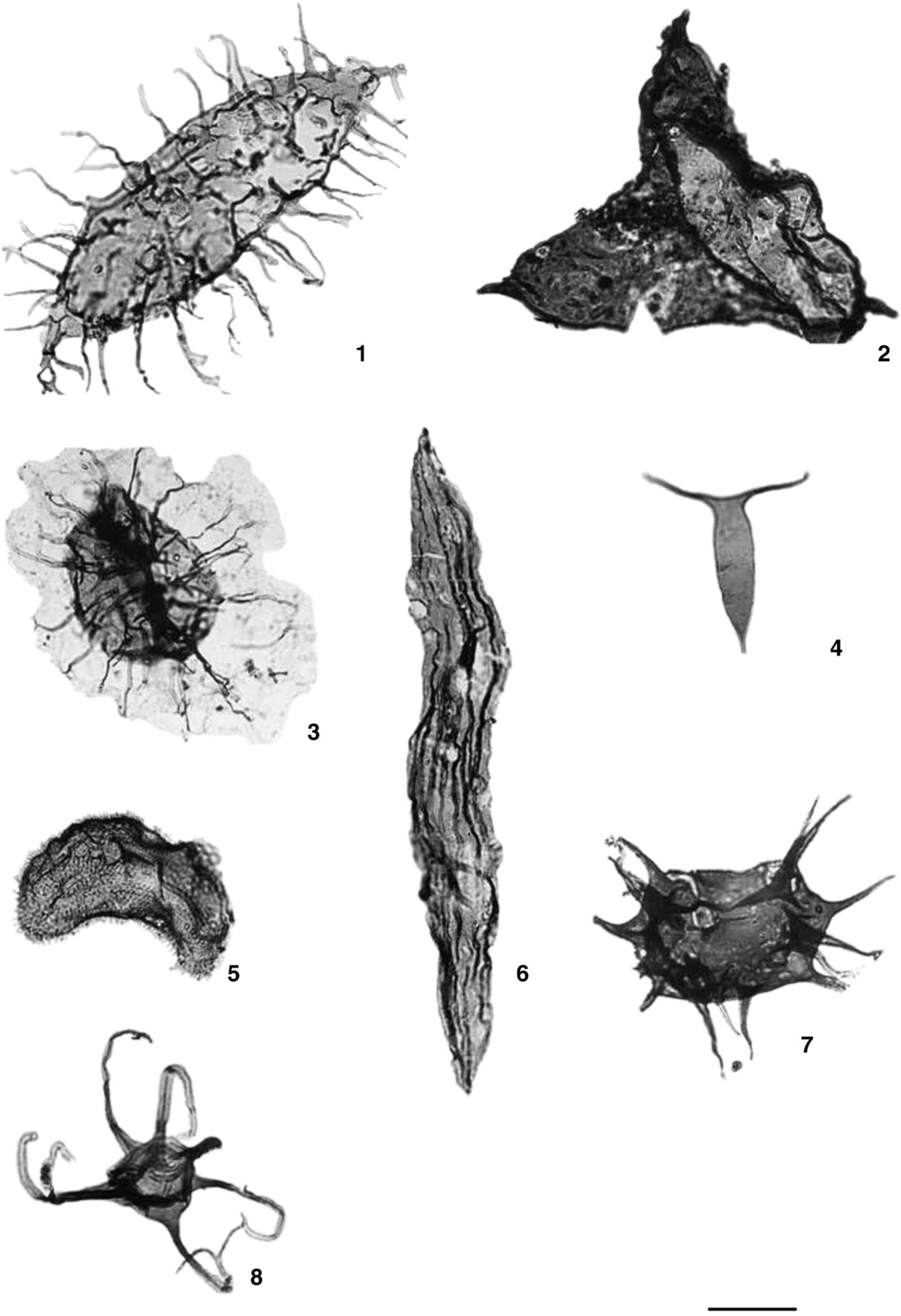


Plate III (caption on page 256).

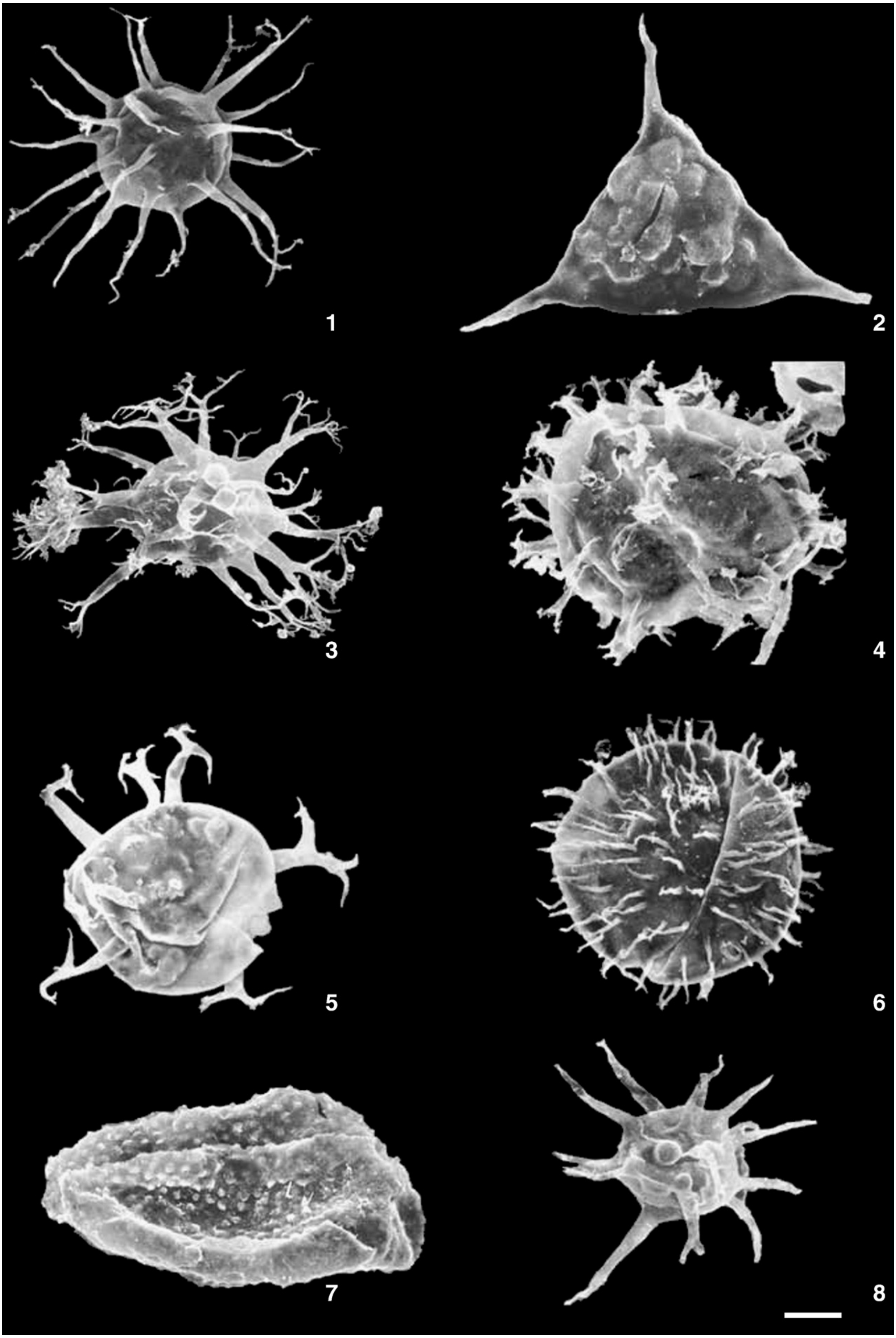


Plate IV *(caption on page 256)*.

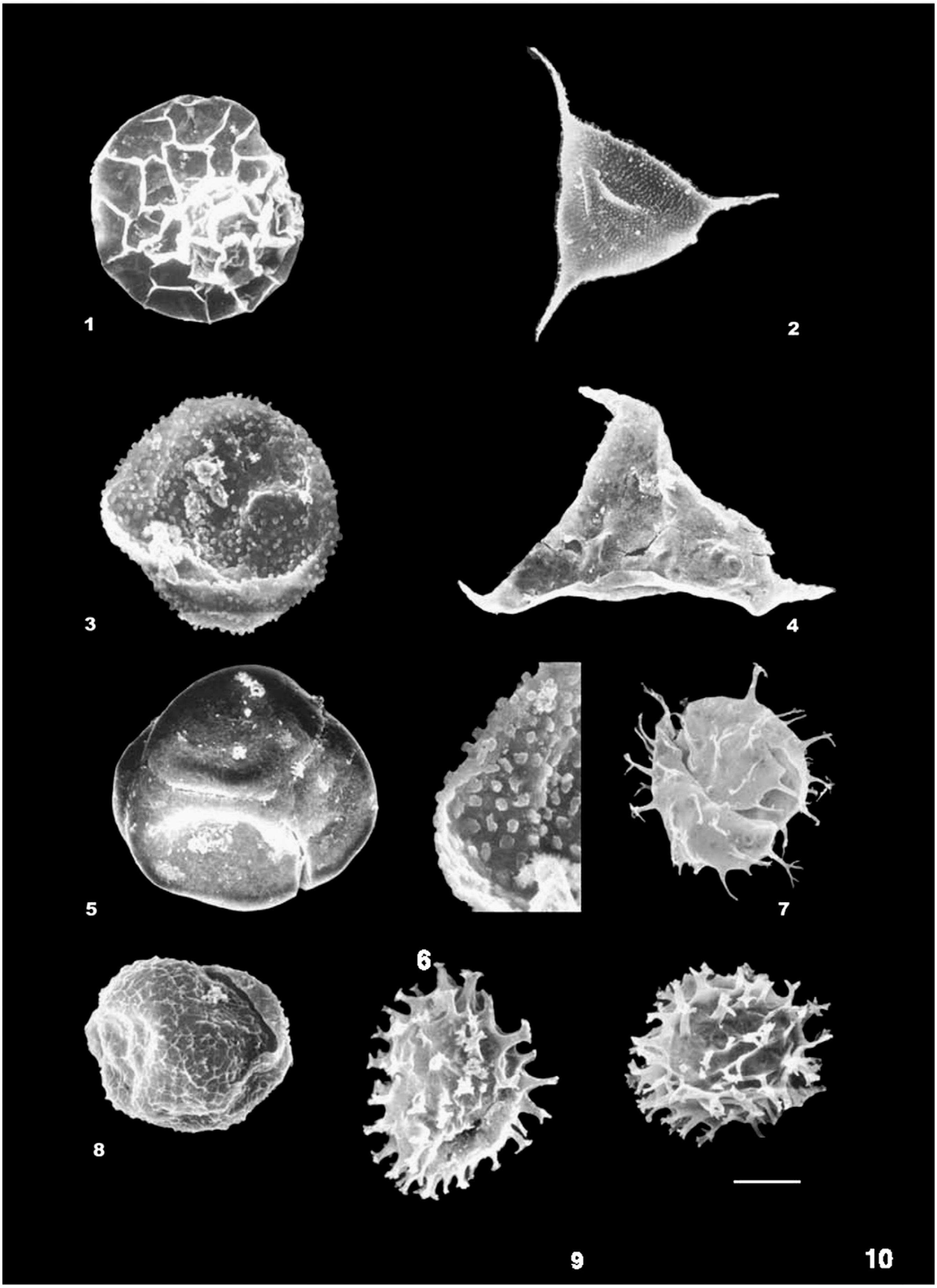


Plate V *(caption on page 256).*

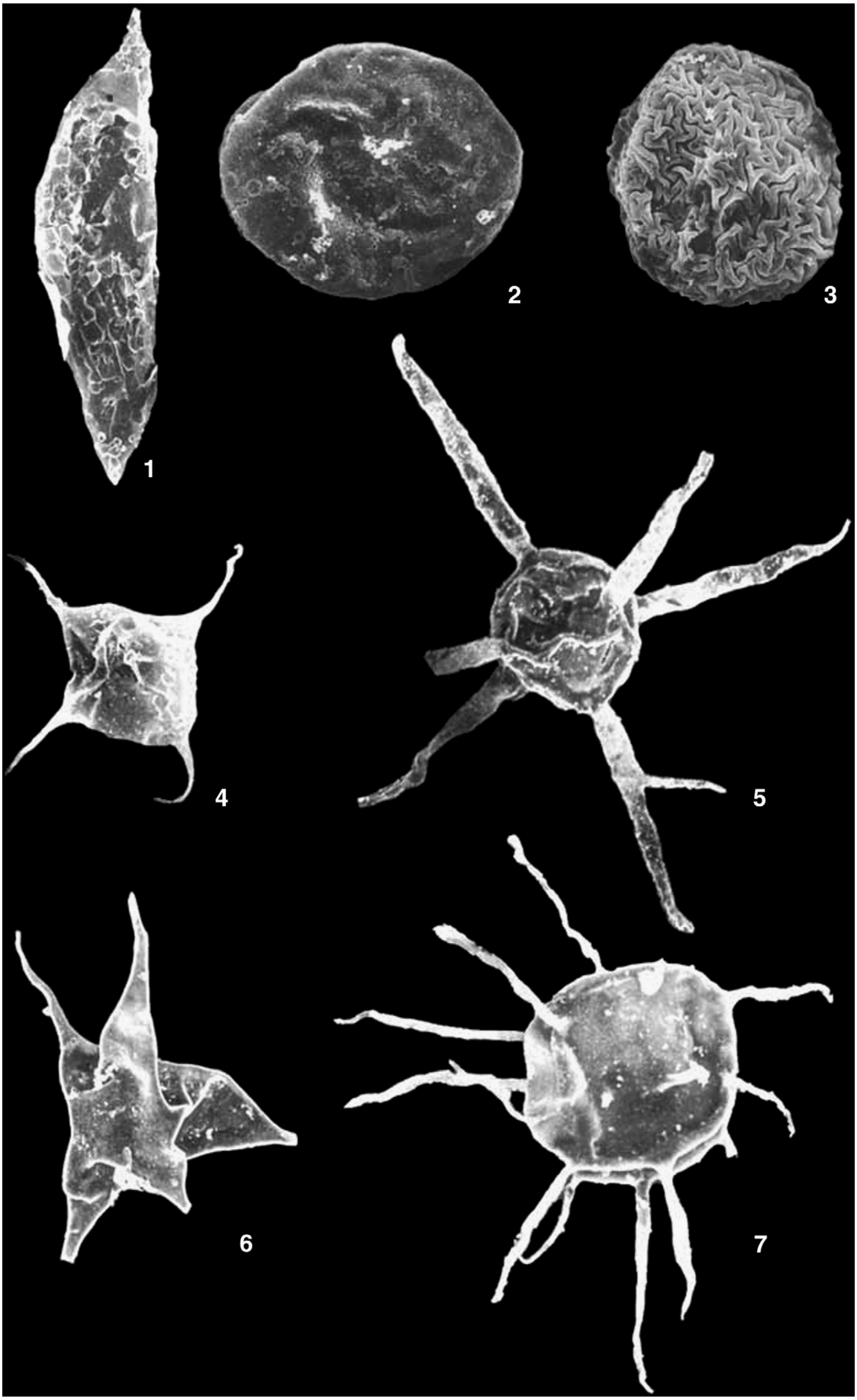


Plate VI *(caption on page 256).*

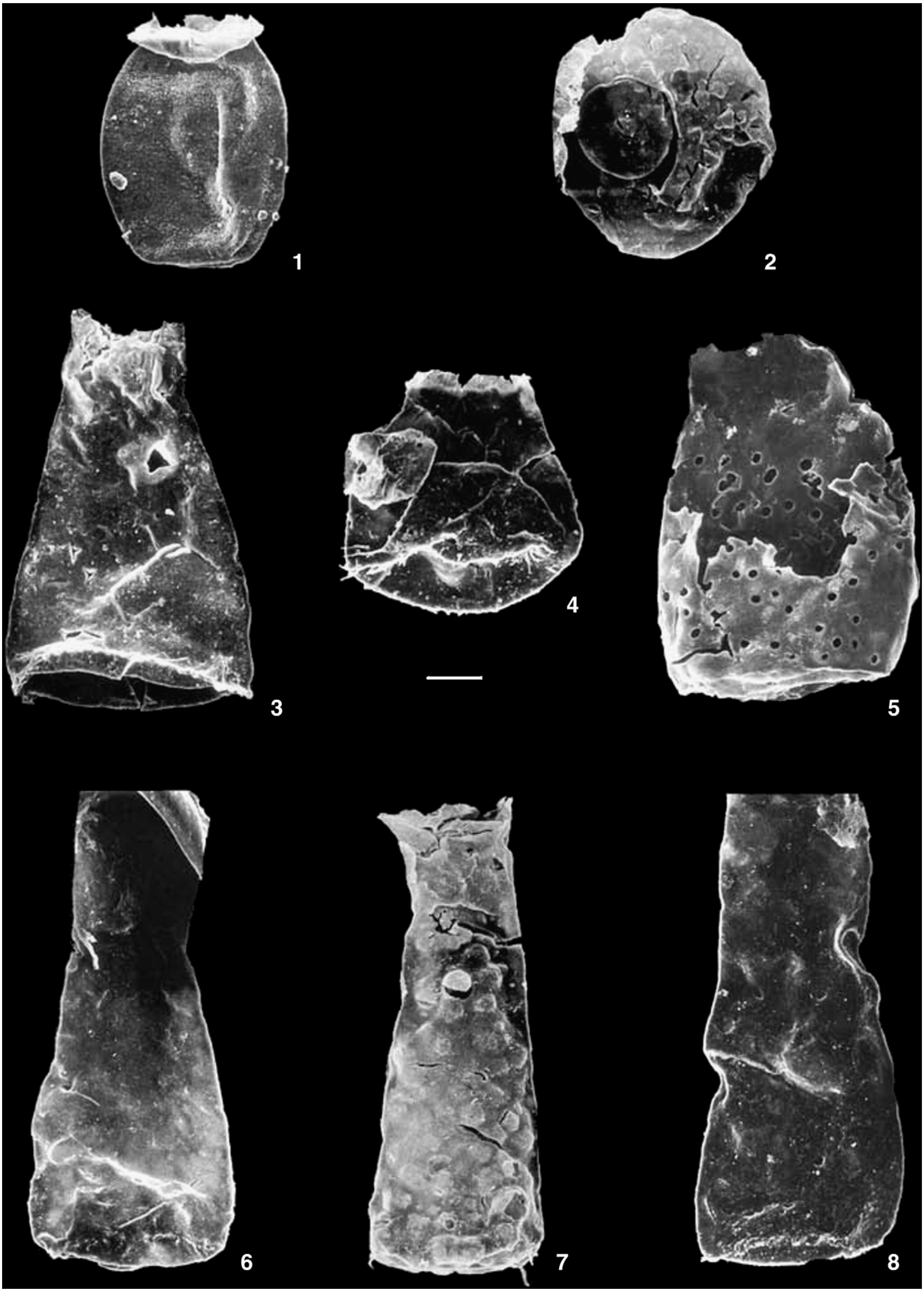


Plate VII *(caption on page 256)*.

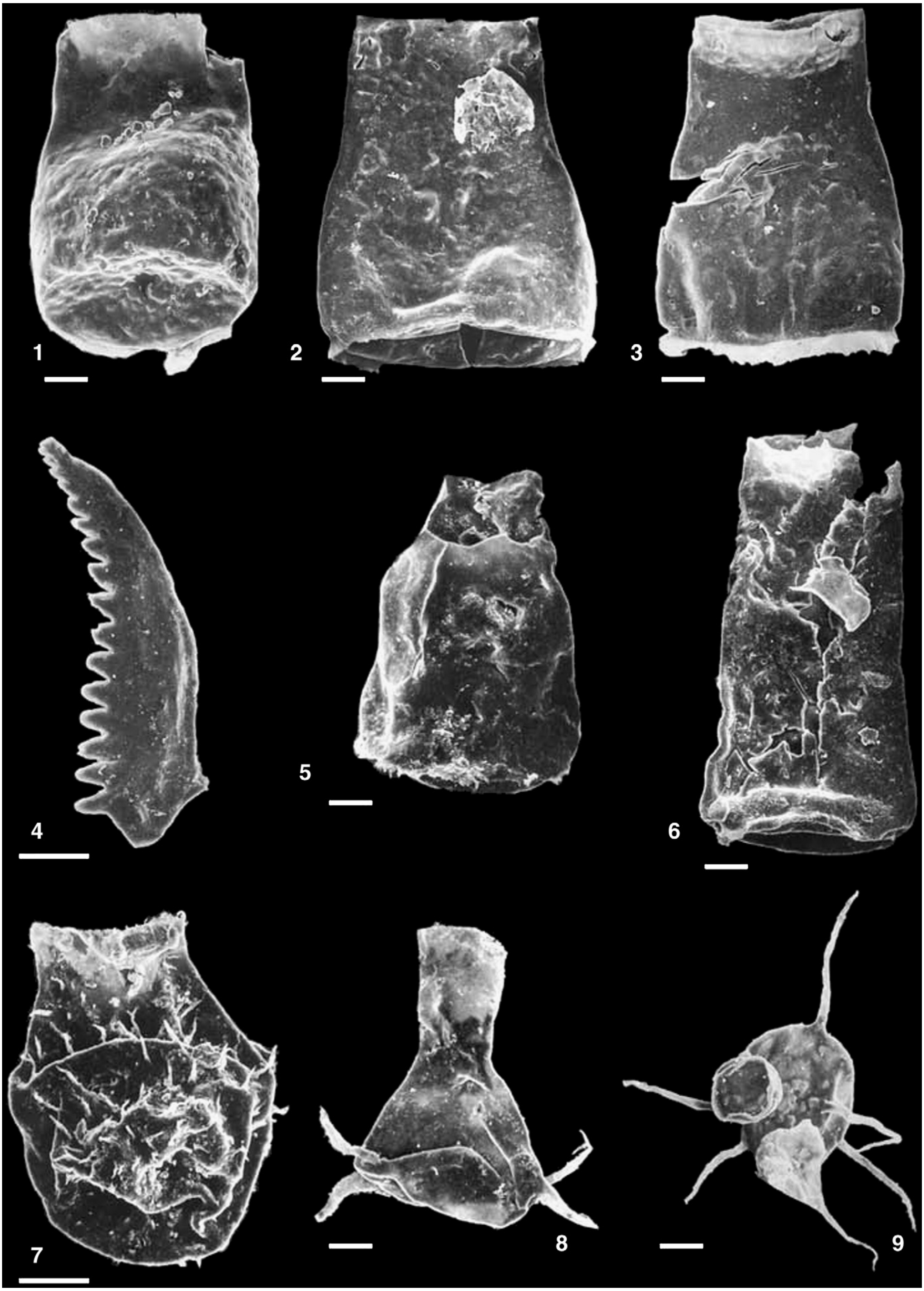


Plate VIII *(caption on page 256)*.

know recorded range of variability. A final comment is necessary concerning the chronostratigraphic significance of the *oulebsiri* biozone; in fact, even if correlation with the *persculptus* graptolite biozone seems to be accepted and demonstrated (Bourahrouh et al., 2004; Paris et al., 2007; Le Heron et al., 2008), there is no published evidence for such a direct correlation (see also discussion in Delabroye and Vecoli, 2010).

In the present material, associated chitinozoans are *C. lenticularis*, *D. minor*, *H.* cf. *spinetum*, and *S. bulmani*, all ranging from underlying strata.

4.2. Acritarchs

The following acritarch taxa have been identified:

Genus *Acanthodiacrodium* Timofeev, 1958 emend. Deflandre and Deflandre-Rigaud, 1962
  Acanthodiacrodium crassus (Loeblich and Tappan) Vecoli, 1999 (Plate III, 7)
Genus *Baiomeniscus* Loeblich, 1970
  Baiomeniscus camurus Loeblich, 1970 (Plate III, 5)
Genus *Baltisphaeridium* Eisenack, 1958 ex Eisenack, 1959 emend. Eiserhardt, 1989
  Baltisphaeridium oligopsakium Loeblich and Tappan, 1978 (Plate VI, 7)
  Baltisphaeridium perclarum Loeblich and Tappan, 1978 (Plate VI, 5)
Genus *Caelatosphaera* Wicander, Playford, and Robertson, 1999
  Caelatosphaera verminosa Wicander, Playford and Robertson, 1999 (Plate VI, 2)
Genus *Cheleutochroa* Loeblich and Tappan, 1978 emend. Turner, 1984
  Cheleutochroa diaphorosa Turner, 1984 (Plate II, 6; Plate III, 8)
Genus *Cornuferifusa* Jacobson and Achab, 1985
  Cornuferifusa sp. (Plate III, 1)
Genus *Cymatiosphaera* Wetzel, 1933, *ex* Deflandre, 1954
  Cymatiosphaera pavimenta (Deflandre) Deflandre 1954 (Plate V, 1)
  Genus *Dactylofusa* Brito and Santos, 1965 emend. Cramer, 1971
  Dactylofusa cabottii (Cramer) Fensome, Williams, Barss, Freeman and Hill, 1990 (Plate I, 4)
  Dactylofusa ctenista (Loeblich and Tappan) Fensome, Williams, Barss, Freeman and Hill, 1990 (Plate VI, 1)
  Dactylofusa platynetrella (Loeblich and Tappan) Fensome, Williams, Barss, Freeman and Hill, 1990 (Plate I, 3,5)
  Dactylofusa striata (Staplin, Jansonius and Pocock) Fensome, Williams, Barss, Freeman and Hill, 1990 (Plate III, 6)
Genus *Domasia* Downie, 1960 emend. Hill, 1974
  Domasia trispinosa Downie, 1960 (Plate III, 4)
Genus *Dorsennidium* Wicander, 1974 emend. Sarjeant and Stancliffe, 1994
  Dorsennidium hamii (Loeblich) Sarjeant and Stancliffe, 1994 (Plate II, 2,3,5)
Genus *Geron* Cramer, 1966 ex Cramer, 1969
  Geron guerillerus Cramer, 1966 ex Cramer, 1969 (Plate II, 4)
Genus *Gorgonisphaeridium* Staplin, Jansonius and Pocock, 1965
  Gorgonisphaeridium antiquum Loeblich and Tappan, 1978 (Plate IV, 6)
Genus *Labyrinthosphaeridium* Uutela and Tynni, 1991
  Labyrinthosphaeridium curvatum Uutela and Tynni, 1991 (Plate V, 8)
  Labyrinthosphaeridium restrictum Uutela and Tynni, 1991 (Plate VI, 3)
Genus *Leiofusa* Eisenack, 1938 emend. Combaz, Lange and Pansart, 1967
  Leiofusa fusiformis (Eisenack) Eisenack, 1938 (Plate I, 1)
Genus *Lophosphaeridium* Timofeev, 1959 ex Downie, 1963
  Lophosphaeridium acinatum Wicander, Playford and Robertson, 1999 (Plate IV, 7; Plate V, 3,6)
Genus *Multiplicisphaeridium* Staplin, 1961 emend. Turner, 1984
  Multiplicisphaeridium bifurcatum Staplin, Jansonius and Pocock, 1965 (Plate IV, 8)
  Multiplicisphaeridium ramispinosum Staplin, 1961 (Plate IV, 3; Plate V, 7)
Genus *Navifusa* Combaz, Lange and Pansart, 1967 ex Eisenack, 1976
  Navifusa ancepsipuncta Loeblich, 1970 ex Eisenack, Cramer, and Díez, 1979 (Plate I, 6)
Genus *Ordovicidium* Tappan and Loeblich, 1971
  Ordovicidium elegantulum Tappan and Loeblich, 1971 (Plate IV, 5)
Genus *Orthosphaeridium* Eisenack, 1968 emend. Kjellstrom, 1971
  Orthosphaeridium rectangulare (Eisenack) Eisenack, 1968 (Plate II, 7)
Genus *Peteinosphaeridium* Staplin, Jansonius, and Pocock 1965 emend. Playford, Ribecai and Tongiorgi, 1995
  Peteinosphaeridium accinctulum Wicander, Playford and Robertson, 1999 (Plate IV, 4)
Genus *Pirea* Vavrdová, 1972
  Pirea sp. aff. *P. ornata* (Burmann) Eisenack, Cramer and Diez, 1976 (Plate II, 1)
Genus *Polygonium* Vavrdova 1966 emend. Sarjeant and Stancliffe, 1994
  Polygonium gracilis Vavrdová, 1966 (Plate IV, 1)
Genus *Rhiptosocherma* Loeblich and Tappan, 1978
  Rhiptosocherma improcera (Loeblich) Loeblich and Tappan, 1978 (Plate VI, 6)
Genus *Safirotheca* Vavrdová, 1989
  Safirotheca sp. cf. *S. safira* Vavrdová, 1989 (Plate I, 2)
Genus *Tunisphaeridium* Deunff and Evitt, 1968
  Tunisphaeridium eisenackii Loeblich and Tappan, 1978 (Plate III, 3)
Genus *Tylotopalla* Loeblich, 1970
  Tylotopalla sp. (Plate V, 9,10)
Genus *Veryhachium* Deunff, 1954 ex Downie 1959 emend. Turner, 1984
  Veryhachium oklahomense Loeblich, 1970 (Plate VI, 4)
  Veryhachium trispinosum group *sensu* Servais et al., 2007 (Plate IV, 2; Plate V, 4)
  Veryhachium sp. cf. *V. triangulatum* Konzalova-Mazancova, 1969 (Plate III, 2)
Genus *Villosacapsula* Loeblich and Tappan, 1976
  Villosacapsula setosapellicula (Loeblich) Loeblich and Tappan, 1976 (Plate V, 2)

4.3. Integrated acritarch and chitinozoan biostratigraphy

Despite the fact that recent studies have demonstrated the potential for a refined biozonation that would allow the recognition of several acritarch biozones at least for the Katian–Hirnantian interval (Vecoli and Le Hérissé, 2004; Vecoli, 2008; Delabroye et al., in press), currently, no formal acritarch zonation exists to subdivide the Upper Ordovician. It is only by means of further detailed studies of Upper Ordovician sedimentary successions worldwide in which independent age evidence is present that progress towards a stable acritarch biozonation can be achieved (Vecoli, 2008). The present results are to be considered as a contribution towards such a goal, and the distribution of the acritarchs are mainly discussed in the context of the chronostratigraphic framework suggested by the chitinozoans, where they are present (i.e., in the Ordovician part of the sequence). The vertical distribution of acritarch species highlights two major changes of the palynoflora throughout the study sequence, which in turn permit the identification of three different assemblages. The first conspicuous change is the extinction of many species at a stratigraphic level coinciding with the base of the *oulebsiri* chitinozoan

Biozone as recognized in our study section (Hirnantian; Fig. 2). This extinction level hence allows the distinction of two assemblages: a relatively high diversity assemblage (Assemblage 1, 34 species) correlated to the *nigerica* and *merga* chitinozoan biozones (late Katian), and a lower diversity assemblage (Assemblage 2, 18 species) correlated to the *oulebsiri* zone (previously considered as late Hirnantian in age, but see discussion below). The second palynofloral change occurs at a level corresponding to the top of the *oulebsiri* Biozone and is defined by the extinction of the large majority of taxa of Assemblage 2; it defines the third assemblage (Assemblage 3, 6 species).

According to standard Ordovician chitinozoan biozonations (Paris, 1990, 1996), a hiatus corresponding to the *elongata* chitinozoan Biozone (early Hirnantian) might occur between the top of the *merga* Biozone and the base of the *oulebsiri* Biozone. However, recent results (Vandenbroucke et al., 2009; and see also discussion in Delabroye and Vecoli, 2010) seem to indicate that the occurrence and stratigraphic extension of the *elongata* Biozone might be facies-dependent, so that the stratigraphic relationships between the *elongata* and *oulebsiri* biozones are not clear, nor is it the precise age of the latter biozone. In any case, it is possible to affirm that the extinction level observed in the study section between Assemblage 1 and Assemblage 2 corresponds at least to a change in palaeoecological conditions, probably marking the onset of glacial-related palaeoenvironments.

Assemblage 1 is characterized by a group of long-ranging species, which range through the stratigraphically younger Assemblage 2, and a second group of species which undergo an abrupt extinction at a level coinciding with the base of the *oulebsiri* chitinozoan Biozone (Fig. 2). This second group comprised many species that are well-reported from generally Upper Ordovician (Caradoc–Ashgill: Sandbian–Hirnantian) strata, and commonly occurring in North American localities (e.g., *Caelatosphera verminosa*, *Cornuferifusa* sp., *Dactylofusa ctenista*, *Dorsennidium hamii*, *Navifusa ancepsipuncta*, *Peteinosphaeridium accinctulum*, *Rhiptosocherma improcera*), but also from Baltican localities (e.g., species of *Labyrinthosphaeridium*). All these species become extinct at the base of the *oulebsiri* chitinozoan Biozone; this sudden change in palynoflora composition is consistent with the possible presence of a hiatus, also indirectly evidenced by the absence of the *elongata* Biozone. This extinction level has most probably a local, or regional, significance and is probably also related to a change in palaeoenvironmental conditions: in North Africa, the *oulebsiri* chitinozoan Biozone is associated to glacial-related sedimentary settings (Paris et al., 2000a; Bourahrouh et al., 2004). The effects of the well-known Hirnantian glaciation on acritarch assemblages might contribute to the first order palynoflora change observed here. In this regard, it is interesting to note that triangular morphotypes of *Veryhachium* (*V. trispinosum* group, and *V.* cf. *triangulatum*) also go extinct at the boundary between Assemblage 1 and Assemblage 2. In previous studies, it has been highlighted that a marked change in the relative proportion of abundance of quadrangular vs. triangular *Veryhachium* is observed in coincidence with the onset of glacial-related palaeoenvironmental conditions during Hirnantian times, with quadrangular *Veryhachium* largely dominating over triangular morphotypes (Le Hérissé in Paris et al., 2000a; Le Hérissé and Vecoli, 2003; Bourahrouh et al., 2004; Vecoli et al., 2009). Generally, no abrupt extinction is recorded in the fossil record of acritarchs in conjunction with the onset of the Hirnantian glaciation in Gondwana, as it has been shown by several studies (Le Hérissé in Paris et al., 2000a; Vecoli and Le Hérissé, 2004). Vecoli (2008) on the basis of a detailed review on acritarch data across the Ordovician–Silurian transition, concluded that an “acritarch turn-over” might be a more appropriate term to describe the significant but gradual changes of the acritarch palynoflora occurring through the end-Ordovician glaciation. According to the discussion above, it is evident that the first extinction level observed in the study section is most probably due to the combined effects of change in palaeoecological conditions and to the presence of a sedimentary hiatus, the stratigraphical extension of which is difficult to evaluate at the present state of knowledge, although it is not larger than the early Hirnantian (Paris, 1996; Paris et al., 2000a; Bourahrouh et al., 2004), and is possibly less (Vandenbroucke et al., 2009; Delabroye and Vecoli, 2010).

It is interesting to note that all species that characterize Assemblage 2 also occur in Assemblage 1, except in a few levels at the base of the study sequence (Fig. 2). The long-ranging species which co-occur in Assemblages 1 and 2 are, in their majority, very well known to occur in many Late Ordovician sequences worldwide: e.g., *Acanthodiacrodium crassus*, *Ordovicidium elegantulum*, *Multiplicisphaeridium bifurcatum*, *M. ramispinosum*, the small quadrangular *Veryhachium* (*V. oklahomense*), and the large *Baltisphaeridium* morphotypes (e.g., *B. perclarum*). Netromorphic acritarchs, such as the characteristic and ubiquitous Upper Ordovician species of *Dactylofusa* (*D. platynetrella*, *D. striata*), and *Leiofusa* (*L. fusiformis*) are also common as relatively long-ranging species. Detailed discussions on the global stratigraphic occurrences of the above taxa can be found in Vecoli (1999) and Playford and Wicander (2006) and do not need to be repeated here. An exception is the occurrence of a morphotype of *Tylotopalla* (Fig. 2; Pl. 5, figs. 9, 10), a genus which has been only rarely reported in Upper Ordovician strata; this morphotype possesses the characteristic features of *Tylotopalla* Loeblich, 1970 such as small size, subcircular vesicle, short processes with small distal furcations, and presence of ridges on the vesicle surface, converging towards process bases. The presence of *Tylotopalla* in Upper Ordovician (Hirnantian) strata, has been recently confirmed in a study of latest Ordovician sedimentary sequences in Anticosti Island, Québec, Canada, where it only occurs in pre-glacial related sediments (Delabroye, 2010).

The base of Assemblage 3 is well defined by the disappearance of the majority of the species constituting Assemblage 2, with the exception of *Baltisphaeridium oligopsakium*, *B. perclarum*, *Safirotheca* cf. *safira*, and *Veryhachium oklahomense*. The extinction level corresponds to the disconformity between the Ghelli and Soltan Maidan Formations, marked by an abrupt lithological change from the shaly sediments of the Ghelli Formation to the pebbly conglomerate at the base of the Soltan Maidan Formation. Assemblage 3 is completed by two Silurian species *Domasia trispinosa* and *Geron guerillerus*, in a thin shaly layer between the basal pebbly conglomerate and the thick sequence of volcanoclastic sediments of the Soltan Maidan Formation (Fig. 2). The low diversity of Assemblage 3 is clearly associated with a lithofacies change and also probably to a sedimentary hiatus coinciding with the disconformity between the Ghelli and Soltan Maidan Formations. The co-occurrence of acritarchs characteristically common in Upper Ordovician sediments (e.g., *Baltisphaeridium perclarum*, *Safirotheca* cf. *safira*, and *Veryhachium oklahomense*), with typical Silurian morphotypes suggests the possibility that the acritarch assemblage is affected by reworking. *Geron guerillerus* has a long albeit not well constrained stratigraphic range, usually taken as to encompass the entire Silurian and part of the Devonian (Diez and Cramer, 1977; Rodriguez-Gonzalez, 1983). Known occurrences of *Domasia trispinosa* define a better constrained stratigraphic range from the late Llandovery (Telychian) to early Ludlow (Gorstian), with most common occurrences recorded in strata of Wenlock age (Le Hérissé, 1989 and in Molyneux et al., 1996). Post-Wenlock (Gorstian, early Ludlow) occurrences of *D. trispinosa* have been proposed but not adequately documented from South American localities (Oliveira and Lima, 1990; Cardoso, 2005). Accordingly, the last appearance datum of *D. trispinosa* is most probably near the Wenlock/Ludlow boundary, or at most within the early Ludlow (Gorstian). If the entire palynoflora is reworked, the age of the Soltan Maidan Formation is not older than Gorstian (early Ludlow, late Silurian); on the other hand if we consider that only the Upper Ordovician forms are reworked and the Silurian species are *in situ*, then the palynology indicates that the

Soltan Maidan Formation is not younger than Gorstian (early Ludlow, early late Silurian), and most probably of Wenlock age (428–423 Ma).

5. Isotopic age constraints for the Soltan Maidan Formation

Sample Kh 1–2 is a rounded cobble-sized pink microgranite clast from the basal conglomerate of the Soltan Maidan Formation. Six of the separated zircons from this specimen were individually analyzed with ion microprobe and yielded concordant U–Pb ages with weighted mean $^{206}Pb/^{238}U$ age of 438.9 ± 8.1 Ma and MSWD = 0.2 (Table 2). Sample Kh 3 from the upper conglomerate level in the same formation (Fig. 2) is also a pink cobble size microgranite clast and petrographically identical to sample Kh 1–2. Three of the separated zircons from this clast were analyzed to support the results attained for sample Kh 1–2, and gave comparable concordant U–Pb ages with weighted mean $^{206}Pb/^{238}U$ age of 426.6 ± 10.5 Ma and MSWD = 0.3 (n = 3) (Table 2). Because the two microgranite samples gave overlapping ages, it is desired to combine the results of the nine zircon crystals to strengthen the statistics. The joined data give weighted mean $^{206}Pb/^{238}U$ age of 434.4 ± 6.4 Ma and MSWD = 0.3 (n = 9) (Table 2). Fig. 3 illustrates the concordant ages of these zircons.

Considering the specified uncertainty, crystallization age of the microgranite source of the analyzed cobbles encompasses the early Silurian (Llandovery) time interval. This finding implies that the Soltan Maidan Formation is younger than Llandovery but older than Devonian because of the age of the overlying Padeha Formation (Late Devonian). The other implication of this isotopic age determination is that the Ghelli Formation must be older than the Llandovery, which is consistent with the biostratigraphic age evidence provided by the acritarchs. It is noteworthy that plutons with this age are entirely unknown throughout Iran.

6. Palaeobiogeographical implications of palynological assemblages

Since the seminal paper published by Vavrdová (1974), who firstly introduced the concept of palaeolatitudinally-controlled acritarch biogeographical provincialism during Ordovician times, several studies have been published confirming not only the validity of such a concept, but also the successful application of acritarch biogeographic distribution for palaeogeographic reconstructions (Cramer and Diez, 1974a,b; Li, 1987; Servais and Fatka, 1997; Tongiorgi and Di Milia, 1999; Vecoli and Samuelsson, 2001a,b; Vecoli and Le Hérissé, 2004; and see also Playford et al., 1995 and Servais et al., 2003 for exhaustive historical reviews). These models show that biogeographic differentiation of acritarch assemblages started in the Early Ordovician and reached a maximum in mid-Ordovician (Darriwilian) times, evidencing a clear distinction between a peri-Gondwanan acritarch "province" (*sensu* Playford et al., 1995, pp. 45–46: "a broad, latitudinally extensive, circumpolar, cold to cool-temperate, palaeogeographic belt along the northern Gondwana border from Argentina through eastern Newfoundland, North Africa, central and southern Europe, and southern Turkey to South China") and a Baltican acritarch "province" (*sensu* Tongiorgi and Di Milia, 1999, p. 299 diffused "primarily in Baltica, including Norway, Sweden, Estonia and Russia, and Poland"). Differentiation of cold-water and warm-water acritarch assemblages has been proposed also for the Tremadoc–Arenig interval (Tremadocian to early Dapingian in terms of modern stratigraphic nomenclature). The "*messaoudensis–trifidum*" assemblage has been defined as a cold-water assemblage in high-latitude peri-Gondwana localities (Molyneux et al., 2007). The *Aryballomorpha–Athabascaella–Lua* "warm water" assemblage has been identified in low-latitude localities of Laurentia and China (Volkova, 1997; Servais et al., 2003). The factors controlling acritarch bioprovincialism are not yet entirely known and evaluated, and even if the palaeolatitudinal gradient plays a role, the effects of oceanic currents (e.g., Tongiorgi et al., 1998) and continental physiography (e.g., Servais et al., 2003) also have a major influence on the palaeobiogeographical differentiation on ancient oceanic microphytoplankton.

So far, evidence for a Late Ordovician acritarch palaeoprovincialism has been considered meager. Insignificant provincialism of acritarch assemblages during Late Ordovician times has been suggested in many previous publications. Hill and Molyneux (1988) noted that Late Ordovician (Ashgill) acritarch assemblages from northeastern Libya displayed some similarity to those recorded from coeval sediments in North America. Uutela and Tynni (1991) reported that 20% of the Late Ordovician acritarch species of Estonia were common with those known from Baltica and the North America regions. Both the late Ordovician acritarch assemblages of Hill and Molyneux (1988; Libya) and Uutela and Tynni (1991; Estonia) included significant proportions of previously unrecorded species with unclear biogeographic significance. Le Hérissé (in Molyneux et al., 1996) noted that Late Ordovician microphytoplankton reveal conspicuous similarity at the generic and specific levels throughout the world, indicating insignificant provincialism during this time. Similarly, Playford and Wicander (2006; U.S.A.) have stated that undoubted provincialism existed during the Early and Middle Ordovician, but this condition was diminished during the Late Ordovician (Ashgill). More recent papers of Li et al. (2006) from the Late Ordovician of China and Ghavidel-Syooki (2006) from Iran supported little microphytoplankton provincialism during Late Ordovician time. The lack of evidence for a biogeographic differentiation of acritarch assemblages might also result from an insufficient data set and hence a difficulty in the comparison between assemblages from different palaeocontinents (Servais et al., 2004). Notably, North American Upper Ordovician localities were largely under-investigated for organic-walled microphytoplankton assemblages, compared to North African or Baltica localities. Wicander (in Servais et al., 2004) noted that North American Late Ordovician acritarch studies did not include entire assemblages but only a few selected species; many assemblages were based on one or only a few samples from a single locality and most of the reported occurrences were not from continuous sections. This situation has started to change more recently with the publication of monographs on acritarch assemblages from well-dated sections from North America (e.g., Playford and Wicander, 2006; Wicander and Playford, 2008, in addition to Wicander et al., 1999), including the first detailed study of the rich and diverse acritarch flora from the well-known Upper Ordovician sections of Anticosti Island, Québec, Canada (Delabroye, 2010). These recent data on Upper Ordovician acritarch assemblages from Laurentian localities and their comparison with Baltican data, allow a new model of acritarch palaeobiogeographical differentiation during

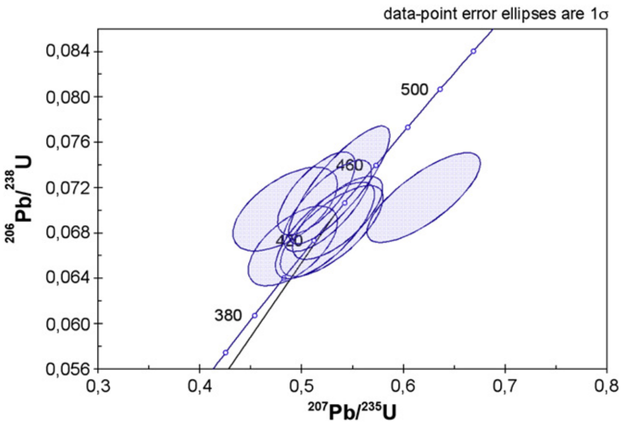


Fig. 3. Concordant ages of zircons from granite pebbles in conglomerate horizons of the Soltan Maidan Formation, Khoshyeilagh Pass, East Alborz Mountain Range, northern Iran.

Late Ordovician times (Delabroye et al., in press). According to this model two phytoplanktonic palaeoprovinces can be distinguished between late Katian and Hirnantian times:

a) An eastern Laurentian–Baltican province, characterized by large *Baltisphaeridium*; of "giant" acritarchs such as *Hoegklintia* spp. or *Estiastra* spp.; of large species of the *Goniosphaeridium–Stellechinatum–Polygonium* plexus; as well as by the presence of *Dilatisphaera* species, such as *D. wimani* (Eisenack) Le Hérissé, 1989;
b) a peri-Gondwanan province, dominated by netromorph acritarchs (e.g., *Dactylofusa cucurbita* jardiné et al., 1974; *Dactylofusa striatogranulata* Jardiné et al., 1974; *Dactylofusa striatifera* (Cramer) Fensome et al., 1990; *Safirotheca* Vavrdová, 1989) and other characteristic forms such as *Tylotopalla* spp. (particularly of *T. caelamenicutis* Loeblich, 1970); *Beromia clipeata* Vavrdová, 1986; and *Neoveryhachium* spp.

This Late Ordovician microphytoplankton biogeographical differentiation has been tentatively attributed to the blocking of exchange between water masses due to the bathymetric ridge-rise associated with the opening of the Rheic Ocean in conjunction with the sea level drawdown occurring during the end-Ordovician glaciation (Delabroye et al., in press). This model is not in contrast with a general perception of diminishing acritarch palaeobioprovinciality in post-Darriwilian times as noted by many previous authors, but it only introduces the possibility of a newly increased bioprovincialism during latest Ordovician times after the generally observed breakdown in palaeobiogeographical differentiation during the post-Darriwilian (Caradoc–Ashgill).

It is noteworthy that many of the typical "Laurentian–Baltican" index taxa do not occur in the present assemblages, with the exception of some large specimens of *Baltisphaeridium* spp. In particular the absence of *Dilatisphaera*, as well as of the "giant" acritarchs *Hoegklintia* and *Estiastra* are noteworthy. The absence of these "giant" acritarchs also reflect ecological, in addition to biogeographical, control on the distribution of these species, which are most often found to occur in carbonate platform environments in low to middle palaeolatitudes (e.g., Delabroye et al., in press; Le Hérissé, 1989). Conversely, the presence and diversity of netromorphic acritarchs (*Dactylofusa*, *Safirotheca*, etc.) as well as the occurrence of a morphotype of *Tylotopalla* suggest similarities with "peri-Gondwanan" assemblages. The change in acritarch assemblages marking the boundary between Assemblage 1 and Assemblage 2, and corresponding to the base of the *oulebsiri* chitinozoan Zone, is also interesting because it coincides with the disappearance of many typical taxa which are found in Laurentian and/or Baltican localities (*Caelatosphera verminosa*, *Cornuferifusa* sp., *Dactylofusa ctenista*, *Dorsennidium hamii*, *Labyrinthosphaeridium* sp., *Navifusa ancepsipuncta*, *Peteinosphaeridium accinctulum*, *Rhiptosocherma improcera*). This change is thus consistent with the hypothesis of the development of a latest Ordovician biogeographic differentiation of the acritarchs between an Eastern Laurentian–Baltican province and a peri-Gondwanan province, linked to the palaeoclimatic changes leading to the end-Ordovician glaciation (Delabroye et al., in press). We are aware that some fusiform–netromorphic taxa, which characteristically occur in the Hirnantian peri-Gondwanan assemblages, also occur in pre-Hirnantian, Late Ordovician assemblages from Laurentian localities; however this is not in contradiction with the proposed biogeographical differentiation developing in latest Ordovician (Hirnantian) times following the palaeoenvironmental and palaeoceanographic consequences of the glaciation, and the observation that high-latitude, Hirnantian, acritarch assemblages are dominated by netromorphs (some of them effectively first described from North American localities), in opposition to Laurentian associations that are largely characterized by totally different acritarch morphotypes, as illustrated above (e.g., large acanthomorphs and "giant" acritarchs; see also details in Delaboye et al., in press).

Chitinozoan research during the past twenty years has resulted in well-documented information for three major palaeogeographic provinces: Laurentia (Achab, 1989), Baltica (Grahn, 1984), North Gondwana (Oulebsir and Paris, 1995; Paris, 1990; Paris et al., 2000a,b; Al-Hajri, 1995; Ghavidel-Syooki, 2000b, 2008; Ghavidel-Syooki and Winchester-Seeto, 2002). In spite of increasing number of studies, fundamental differences have not been revealed in the composition of chitinozoan assemblages between major palaeogeographic domains. So far, five genera (Paris, 1990) appear to be restricted to the North Gondwana realm during the Ordovician when this realm was located in relatively high latitudes (e.g., *Armoricochitina*, *Eremochitina, Velatachitina*, *Sagenochitina,* and *Siphonochitina*). Even if many cosmopolitan species are known from both equatorial and circum-polar palaeolatitudes (e.g., *Acanthochitina symmetrica*, *Cyathochitina campanulaeformis*, *Desmochitina minor*, *Lagenochitina baltica*, *Plectochitina sylvanica*), the biozonations of Achab (1989: Laurentia), Paris (1990: Gondwana), and Webby et al. (2004: Baltic and Gondwana Domains) highlight some differences, especially at the species level, between the three major palaeogeographic domains. All of the chitinozoan species found in the Late Ordovician of

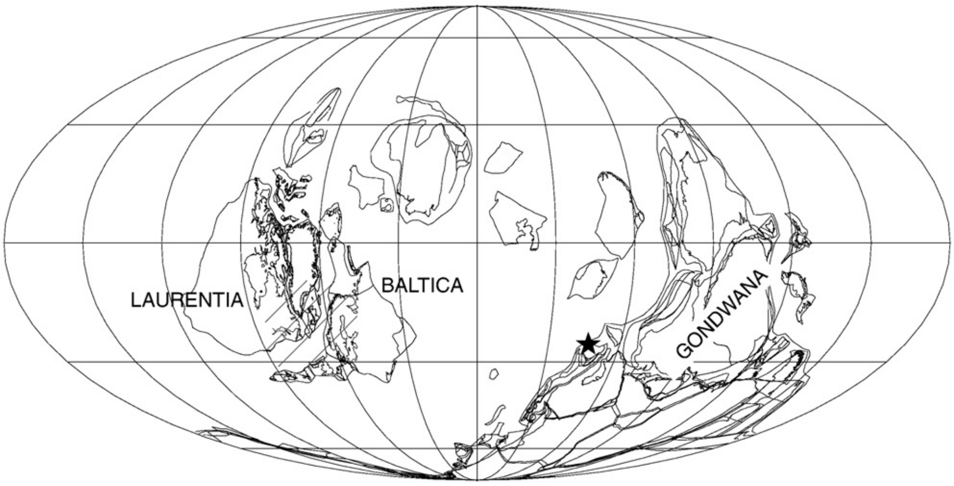


Fig. 4. Palaeogeographical model of continental mass distribution during Late Ordovician (Hirnantian) times, based on BugPlates software (courtesy of Center for Geodynamics, Geological Survey of Norway). Shaded area indicates the eastern Laurentia–Baltican Late Ordovician acritarch province. Star indicates the palaeoposition of the study area, assigned to the North Gondwanan Province.

the Khoshyeilagh area have been recorded from the North Gondwana Domain (including North Africa: Morocco, Algeria, Tunisia, Libya, and Nigeria; the Middle East: Saudi Arabia, Syria, Jordan, Iran; southwestern Europe: Italy, France, Spain, and Portugal; and Central Europe: Czech Republic). In particular, *Armoricochitina nigerica*, *Ancyrochitina merga*, and *Spinachitina oulebsiri* have never been recorded outside the North Gondwana Domain (Paris, 1981, 1990; Webby et al., 2004).

In conclusion, evidence from acritarch and chitinozoan biogeographic affinity indicates that the study locality belongs to a peri-Gondwanan palaeogeographical domain; this is clearly consistent with recent palaeogeographic reconstructions which place northern Iran at the periphery of the Gondwana supercontinent, at intermediate palaeolatitudes (Fig. 4), palaeogeographically well separated by a large ocean from the Laurentian–Baltican palaeocontinents.

7. Conclusions

In this paper, we have presented the evidence for the biostratigraphic and isotopic dating of the mixed siliciclastic and volcanic successions of the Ghelli and Soltan Maidan Formations outcropping in the Khoshyeilagh area in the eastern Alborz Range, northern Iran, which previously had been assigned to a general Ordovician to Silurian age, based only on indirect correlation (Ghelli Formation), or on stratigraphic relationships (Soltan Maidan Formation). Moreover, some implications for the palaeobiogeographic differentiation of Late Ordovician oceanic microplankton have been discussed. Biostratigraphic age evidence comes from rich and well-preserved palynological assemblages, i.e., chitinozoans and acritarchs. Three chitinozoan biozones have been identified in the Ghelli Formation, namely the *Armoricochitina nigerica* Biozone (early late Katian), the *Ancyrochitina merga* Biozone (late Katian), and the *Spinachitina oulebsiri* Biozone (late? Hirnantian). The absence of the *Tanuchitina elongata* Biozone, normally recorded in the North Gondwana Domain between the *merga* and *oulebsiri* biozones and generally attributed to the early Hirnantian, possibly indicates the presence of a hiatus encompassing part of the Hirnantian Stage. The acritarch palynoflora recovered from the study section can be subdivided into three successive assemblages here informally indicated Assemblages 1, 2, and 3. Assemblages 1 and 2 are found in the Ghelli Formation and Assemblage 3 occurs in the lowermost Soltan Maidan Formation. Assemblage 1 is correlated to the *nigerica* and *merga* chitinozoan biozones and it is distinguished from Assemblage 2 by the disappearance of numerous taxa in correspondence to a level at the top of the *merga* Biozone. This abrupt palynofloral change can be attributed to a change in palaeoenvironmental conditions (onset of glaciation in Gondwana) together with the possible presence of a sedimentary hiatus. Acritarch Assemblages 1 and 2 are characterized by the presence of many taxa that are diagnostic for a Late Ordovician age, which is entirely consistent with the age indications provided by the chitinozoans. Acritarch Assemblage 3 occurs in the shaly interbeds of the basal part of the mainly volcanoclastic and volcanic succession of the Soltan Maidan Formation; it is characterized by an impoverished and mixed assemblage including some characteristic Upper Ordovician and Silurian taxa, thus evidencing sediment reworking. In the case of reworking of the entire palynoflora, the age of the Soltan Maidan Formation would be not older than Gorstian (early Ludlow, late Silurian); on the other hand if we consider that only the Upper Ordovician forms are reworked and the Silurian species are *in situ*, then the palynology indicates that the Soltan Maidan Formation is not younger than Gorstian, and very possibly even Wenlock in age. We favor this latter hypothesis because of the absence of post late-Silurian reworked palynomorph within the assemblage.

The proposed biostratigraphic age attributions are supported by the isotopic analysis of zircons extracted from granite pebbles collected in the lower part of the Soltan Maidan Formation, and indicating mean radiometric ages of 434.4 ± 6.4 Ma (Telychian). A Wenlock–early Ludlow (Gorstian) depositional age for the Soltan Maidan Formation is consistent with the radiometric ages of the granitic clasts, allowing sufficient time (in the order of five to ten million years) for the unroofing, erosion and redeposition of the granite conglomerates. Rate of exhumation of crustal blocks is highly variable and changes considerably from place to place as a function of tectonic setting, palaeoclimate conditions, and palaeogeographic situation. In tectonically active belts, shallow plutons can reach the Earth surface in less than a million years. Examples for such rapid uplift rates are the late Pliocene Eldzhurtinskiy granite (Caucasus: Hess et al., 1993) and the Pliocene–Quaternary Takidani Granodiorite in central Japan (Harayama, 1992). Our results then suggest that our study area in the Eastern Alborz was indeed tectonically very active in Silurian times.

From a palaeobiogeographic point of view, acritarch Assemblage 1 comprises many species with a worldwide distribution, but also commonly found in Laurentian and Baltican localities. Conversely, Assemblage 2 is mainly characterized by taxa which are now considered as typical of a Late Ordovician peri-Gondwanan bioprovince or realm, such as abundant and diverse netromorphs, presence of *Tylotopalla*, coupled to the absence of all the typical taxa abundantly occurring in the Hirnantian of eastern Laurentia and Baltica, such as the giant acritarchs *Hoegklintia* and *Estiastra* or species of *Dilatisphaera*. The present results, thus, are consistent with a recently proposed model of palaeobiogeographical differentiation of latest Ordovician acritarchs assemblages into two provinces, or realms: Laurentian/Baltican and peri-Gondwanan (Delabroye et al., in press). The “North Gondwana” affinity of the chitinozoan assemblages from the Ghelli Formation. corroborates this conclusion. The palaeobiogeographic indications given by the palynomorphs are also entirely consistent with the recent palaeogeographic reconstructions, which place the northern Iran block at the margin of Gondwana, in intermediate palaeolatitudes, and geographically well separated by the Laurentia/Baltica palaeocontinents. Finally, the presence of cryptospores together with taxonomically diverse chitinozoan, acritarch, and scolecodont assemblages, suggest a relatively shallow marine, platformal depositional environment for the Upper Ordovician sediments of the Ghelli Formation. This latter can be considered as time-equivalent and non-metamorphic analog of the Gorgan Schist sequence.

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