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# Palynologic delineation of the Devonian–Carboniferous boundary, West-Central Montana, USA

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## ABSTRACT

Cratonic depositional systems in the Central Montana Trough involve the Devonian–Carboniferous boundary (DCB), and reflect both subtle regional epeirogeny and significant global glacioeustatic controls. A palynologic analysis of the upper Three Forks, Sappington and lower Lodgepole formations was carried out at the classic Logan Gulch location in Horseshoe Hills. The lower Trident Member of Three Forks Formation yielded low-diversity cosmopolitan, long ranging phytoplankton and few spores species (LAs1), attributed to the middle Famennian. The upper part of the same green seaway shale yielded only leiosphaerids and *Botryococcus* (LAs2), along with an external mold of a clymenid ammonoid. Age-diagnostic spores *Retispora lepidophyta*, *Verrucosisporites nitidus* and *Vallatisporites vallatus* from middle Sappington siltstone (LAs3) indicated a Strunian LN Zone. Two more assemblages from upper Cottonwood Canyon Member (LAs4) and false Bakken (LAs5), in the lower Lodgepole Formation yielded scarce, poorly preserved spores. The presence of *Waltzispora polita* in LAs4 indicated a Tournaisian–Visean age.

## KEYWORDS

Palynology; D–C boundary; Three Forks Formation; Sappington Formation; Lodgepole Formation; Montana; USA

## 1. Introduction

The middle Famennian Three Forks and latest Famennian–Tournaisian Sappington and Lodgepole formations of west-central Montana are part of a regionally extensive sedimentary package deposited on the western part of Laurentia (Figure 1). Very similar correlatives include the Bakken Formation in eastern Montana, North Dakota and Saskatchewan, the ‘Alberta Bakken’ in northern Montana and Alberta and the Exshaw Formation in Alberta. Similar facies and stacking patterns also exist as far south as Utah. The stratigraphy is dominated by areally extensive system tracts and sequence boundaries developed in response to high-amplitude eustatic base-level changes that overprint more regional tectonic effects and make these strata appropriate for studying the Devonian–Carboniferous boundary (DCB) in western North America.

Previous work on the DCB has shown that major sequences are related to second- and third-order icehouse episodicity, and that higher order sequences correlate to regional, tectonically forced regressions (e.g. Giles et al. 1999; Sandberg et al. 2002; Simenson et al. 2011; Kumpan et al. 2014) and global eustatic drawdowns (Johnson et al. 1985; Isaacson et al. 1999, 2008; Kaiser et al. 2006, 2011; Haq & Schutter 2008). Several extinction events affected both terrestrial and marine environments during this interval, especially just prior to the DCB. De Vleeschouwer et al. (2013), Kaiser et al. (2011, 2015) and Becker et al. (2016) reviewed and discussed different aspects related to the Hangenberg crisis and agreed that more studies are needed to delineate the global DCB. Caplan & Bustin (1999), Streel et al.

(2000), Racka et al. (2010), Marynowski et al. (2012) and McGhee (2013), among others, described the global Hangenberg extinction event and distinguished high extinction and low origination rates due to major eustatic sea-level fall during a glacial advance. The globally correlated event is a sudden anoxic and transgressive pulse well known to have occurred at the end of the Devonian (Trapp et al. 2004). Caplan & Bustin (1999) mentioned that the marine ammonoids (clymeniids and goniatites), trilobites and conodonts were the most severely affected fossil groups and showed abrupt diversity and abundance declines, whereas in other groups the change was stepwise or gradual (e.g. acritarchs, agnathan fishes, foraminifera, brachiopods, echinoderms, ostracodes and some corals).

The Global Boundary Stratotype Section and Point (GSSP) for the DCB boundary defined in France is not under revision for now, but it was critically analysed by Kaiser et al. (2015) and Becker et al. (2016), concerning the appearance and extinction events of key taxa involved (invertebrates, palynomorphs, microfossils). The difficulty in recognising this boundary elsewhere stems from problems interpreting the relationship between *Siphonodella praesulcata* and *S. sulcata*. The first appearance of the conodont *S. sulcata* in the basal Tournaisian is linked to the evolutionary lineage of *S. praesulcata*; in some areas gradational morphotypes prevent the delineation of this boundary (Kaiser & Corradini 2011). Recently, Kumpan et al. (2014) recommended a ‘CKI (*costatus-kockeli*) Interregnum’ (proposed by Kaiser et al. 2009) in lieu of the lower and middle *praesulcata* zone, with *kockeli* at the very latest Famennian. On the



other hand, the extinction of *Retispora lepidophyta*'s parent plant at the end of the Devonian represents an important terrestrial global change, along with the disappearance of other spore species (e.g. *Rugospora flexuosa*, *Diducites versabilis* and *D. plicabilis*). Most of the Palaeozoic acritarchs and all of the chitinozoans that were very abundant up to the Famennian also disappeared in the latest Famennian (Streel et al. 2000). This was used to define the base of the *V. vallatus*–*R. incohatus* Zone (earliest Tournaisian) between the Lower and Upper Stockum Limestones of the type area in Europe (Higgs & Streel 1984, 1994; Higgs et al. 1993; Streel 2015).

Kaiser et al. (2015) referred to the durations of many DCB biozones that are of only 100–300 kyr (see interpolations in Becker et al. 2012). Based on geochronological dating of Polish ash layers (Myrow et al. 2014), the main event or black shale interval lasted between 50 and 190 kyr. This supports a very sudden extinction at the base of this shale, and estimates that the whole (extended) crisis interval was only between 100 to several hundred kyr in duration (Sandberg & Ziegler 1996; Becker et al. 2012; De Vleeschouwer et al. 2013).

Orbital cyclostratigraphy of the Famennian succession in Poland supports that transgressive periods of *Annulata*, *Dasberg*, *Kowala* and *Hangenberg* black shales (upper Famennian) are triggered by a maximum eccentricity cycle of 2.4 Ma, which provoked the collapse of presumably small continental Gondwanan ice-sheets (De Vleeschouwer et al. 2013). Rapid climatic cooling and a glaciation pulse follow the Hangenberg Black Shale (HBS) Event in Gondwana leading to global regression in the earliest Tournaisian (McGhee 2013). Although several works have expanded the geographic boundary of the regional glaciation, there are controversial hypotheses about cooling/warming phases through the DCB in Bolivia, Peru and Brazil (Frakes et al. 1992; Díaz-Martínez & Isaacson 1994; Crowell 1999; Isaacson et al. 1999, 2008; Caputo et al. 2008; Playford et al. 2012; Streel et al. 2013; Kaiser et al. 2015). Also, local records of alpine-type events in the USA are known in the central Appalachians (Brezinski et al. 2008; Etensohn et al. 2009; Kaiser et al. 2015).

The recognition of many diagnostic taxa from the Famennian and especially *Retispora lepidophyta* in Mississippian and Pennsylvanian units due to reworking of the Upper Devonian rocks in Bolivia has hindered the calibration of the extinction events in this region of South America (e.g. di Pasquo & Azcuy 1997; di Pasquo 2007; Streel et al. 2013; di Pasquo et al. 2015). Therefore, the timing and mode of initiation of this Gondwanan glaciation, and its mechanistic link to eustatic fluctuations, changes in continental weathering rates, and palaeo-oceanographic and climatic conditions, remain poorly resolved. Questions of reworking of microfossils at the DCB were discussed by Macke & Nichols (2007), who have also interpreted as reworking mixtures of assemblages with *Siphonodella sulcata* and *S. praesulcata* documented in western USA. Becker et al. (2016), who proposed five alternative explanations based mostly on conodont zonations, discussed these issues regarding the definition of the global DCB.

In selected North American successions, the recognition of the DCB has been relatively straightforward, because the boundary has been relatively well-defined with palynomorph- and conodont-based assemblages and concurrent range zones (e.g. Klapper 1966; Sandberg et al. 1972; Gosney 1982; Hayes & Holland 1983; Holland et al. 1987; Savoy & Harris 1993; Caplan & Bustin 1999; Johnston et al. 2010; Wicander & Playford 2013, and references therein).

Absolute age dates in the lower Exshaw Formation black shales support this latest Devonian age (e.g.  $363.34 \pm 0.39$  Ma) and associated conodont biostratigraphy (Richards et al. 2002). Hence, the remaining Exshaw Siltstone and overlying Banff Formation black shales are thought to be Mississippian (Richards et al. 2002; Johnston et al. 2010). Playford & McGregor (1993) studied the palynology of the three members of the Bakken Formation from cores of several boreholes in southern Saskatchewan (Figure 1), and found *Retispora lepidophyta*. However, the recognition of the *Siphonodella crenulata* Zone (Tournaisian) in its upper member led these authors to interpret *Retispora lepidophyta* (and other Famennian species) as recycled into the Early Mississippian (see also Playford 1993).

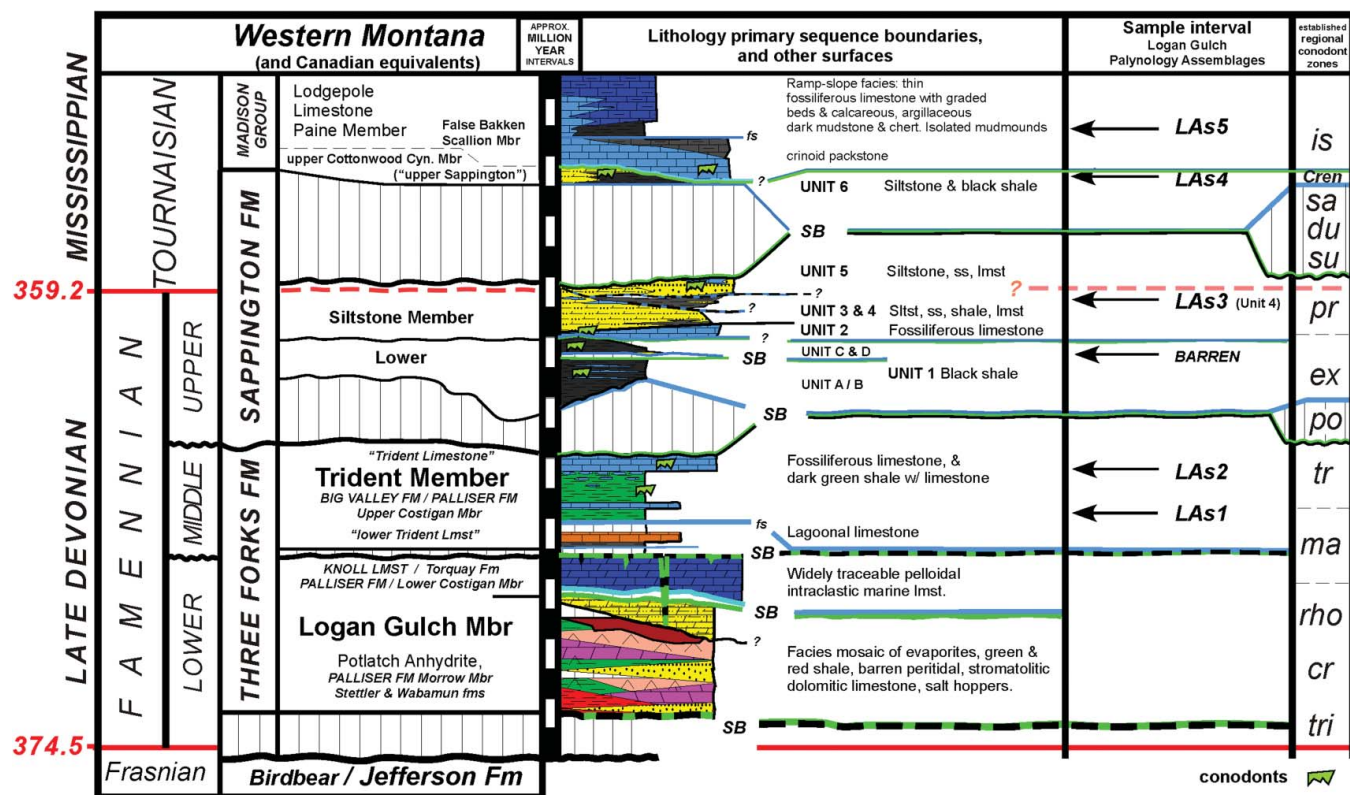
In Montana, the DCB was proposed between the upper siltstone unit of the Sappington Formation (Unit 5, Figure 2) and the Cottonwood Canyon Member of the Lodgepole Formation (also commonly known as 'Sappington Unit 6'; Sandberg et al. 1972; di Pasquo et al. 2012, 2016; Hartel et al. 2014; Warren et al. 2014). Currently, the exact position of the DCB needs to be better constrained, but there is a consensus that the organic-rich black shales of the upper Cottonwood Canyon Member, basal black shales of the Banff, and upper Bakken formations fit the global record of a major Early Mississippian sea level rise following a major fall (see Haq & Schutter 2008; Kaiser et al. 2015). Exactly when this occurred in the Tournaisian needs further study.

In order to understand palaeoenvironmental changes and their relationship to base-level changes across the DCB, we carried out a palynological analysis of the upper Three Forks, Sappington, and lower Lodgepole formations at several locations in western Montana. The results from the Logan Gulch section in the Horseshoe Hills near Three Forks, Montana (Figure 1), are presented herein. Integrated early studies and new work at these sections allow us to define lacunae and low-latitude proxy-beds that are interpreted as time-equivalent to the Gondwana glaciations. A discussion of their biostratigraphic importance for the region is extended to the Bakken Formation and other American to global DCB correlative units.

## 2. Geologic framework and previous work

The Central Montana Trough (CMT) was a structurally inherited, east–west trending cratonic feature characterised by changing seaway geometries inboard of the north–south-oriented active Antler foredeep (Figure 1). Late

**Figure 1.** Regional Devonian–Carboniferous palaeogeographic map showing the location of the Logan Gulch Devonian type section (LG, star) in the Three Forks area (modified after Sandberg & Mapel 1967; Savoy 1992; Lund 2008). Further locations mentioned in the text are: AtC = Antelope Creek; AshC = Ashbough Canyon; BBG = Brown Back Gulch; BBM Bandbox Mountain; BoG = Bouge Canyon; FrL = Frazier Lake; H = Hardscrabble (Peak 9559); IsQ = Ingleside Quarry; LCC = Lewis & Clark Caverns; RcDR = Rock Creek, Dry Range; RH = Red Hill; NG = Nixon Gulch; MC = Milligan Canyon Sappington Fm Type Section; MoC = Moose Creek; MsG = Mudspring Gulch; and the number 1 refers to the area of subsurface boreholes studied by Playford & McGregor (1993).



**Figure 2.** Stratigraphic information on the Devonian–Carboniferous boundary interval in western Montana. Famennian and lower Tournaisian stratigraphic nomenclature, lithologies, sequence stratigraphy, palynomorph recovery and description. Sappington Strata units 1 through 6 are widespread but vary in thickness from 0 to 10 to 40 m. Chronology and absolute ages 374.5 and 359.2 Ma (at the left) after Cohen et al. (2014). Abbreviations of conodonts zones (after Racka et al. 2010; Becker et al. 2016; see also Figure 7): Tri = *Palmatolepis triangularis*. Cr = *P. crepida*. Ro = *P. rhomboidea*. Ma = *P. marginifera*. Tr = *P. trachytera*. Po = *P. postera*. Ex = *P. expansa*. Pr = *Siphonodella praesulcata*. Su = *S. sulcata*. Du = *S. duplicata*. Sa = *S. sandbergi*. Cren = *S. crenulata*. Is = *S. isosticha*.

Devonian stacking patterns are relatively uniform and stratigraphic thicknesses change subtly across bordering cratonic palaeohighs, where middle and upper Famennian sequence boundaries and transgressive surfaces merge, creating unconformities (McMannis 1962; Sandberg & Klapper 1967). Deposition occurred discontinuously throughout the 15-my-long Famennian in low-accommodation settings during punctuated transgressions and regressions during an overall second-order regression (Johnson et al. 1985; Sandberg et al. 2002) and Early Mississippian transgression. Further, shale deposition, also with organic-rich black shale, is repeated, suggesting changes in weathering, humidity, nutrients and primary productivity that are indicative of changing equatorial climates (e.g. glacially induced sea-level drops after Isaacson et al. 1999, 2008). Hence, palaeoequatorial successions like the sections in Montana are important records of global events occurring during the DCB in glacial Earth history (Isaacson et al. 1999, 2008; Caplan & Bustin 1999; Streel et al. 2000; Cecil et al. 2004; Batt et al. 2007; Elrick et al. 2009; Brezinski et al. 2010; Kaiser et al. 2015).

### 2.1. Stratigraphy and palaeontology of formations straddling the DCB in western Montana, and correlative regional units

Frasnian subtidal dolostones of the Jefferson Formation underlie the global Frasnian/Famennian disconformity and are

overlain by the Three Forks Formation (Figure 2). The Three Forks Formation is subdivided into the lower Logan Gulch Member, a restricted evaporitic facies and an upper Trident Member composed in basinward settings of fossiliferous, open-marine green shales and limestones. A prominent limestone commonly forms the top of the Trident Member beneath a sequence boundary (SB) that marks the base of the Sappington Formation (Figure 2; McMannis 1962). The Three Forks successions exhibit features including lacunae and erosion, phreatic zone breccias, craton-derived sands and cryptic karst formed during subsequent early and middle Famennian transgressions in shallow seas that covered most of Montana, North Dakota and southern Canada at approximately 10°S of the palaeo-equator (Figure 1; Blakey 2008; Scotese 2014).

Although the Sappington Formation was included with the Three Forks Formation at Logan Gulch for large-scale mapping purposes (Sandberg 1962), it is clearly a separate, younger and mappable formation, and subject to further regional allostratigraphic refinements (Achauer 1959; Smith & Bustin 2000; Phelps 2015). The Sappington Formation was originally divided into widely occurring lithostratigraphic units A through I (Gutschick et al. 1962). MacQueen & Sandberg (1970) proposed and used a simpler numerical scheme (units 1–5; Figure 2), removing Unit I from the Sappington, although all early Gutschick papers and modern Bakken workers include Unit I as the upper, lithostratigraphic part of the Sappington: Unit 6. Recent workers have partly synthesised these schemes (e.g. Unit 1 with

subunits A through D or U6 A to C in Rodriguez 2014), breaking out additional packages and thereby introducing associated genetic frameworks (Phelps 2015). Sappington units consist of basal black shales (Unit 1 or Lower Sappington) and a Middle 'Siltstone Member' of silty limestone, calcareous siltstone, shale and siltstone/very fine sandstone (units 2 to 5). Simplistic interpretation suggests Sappington Unit 1 (Figure 2) is a basal transgressive relatively deep-water (shoreface or deeper) black shale overlain by the Siltstone Member that includes units 2 through 5 (equivalent to units Gutschick's units E, F, G, H). This succession was deposited in a shallow-water storm-dominated shoreface environment, and it is correlative to the Middle Bakken. Upper Sappington black shale and fine clastics (Unit 6 or 'upper black shale' of the Cottonwood Canyon Member, CCM) is correlative to the Upper Bakken Formation of North Dakota and likewise underlies a thick fossiliferous to shaley, or argillaceous and cherty, limestone succession of the Lodgepole Formation (Andrichuk 1955; Sandberg 1962; Elrick & Read 1991). In the study area Unit 6 is formally defined as the upper tongue of the CCM, a time and facies correlative to the Sappington Formation in Wyoming (Sandberg & Klapper 1967).

The base of the Lodgepole above the Bakken Formation in the Williston Basin is called the Scallion Member (cf. Stroud & Sonnenberg 2011) and occurs under the 'false Bakken' black shale; correlatives of these strata are represented in south-central Montana. The Lodgepole Formation was deposited in a moderate to deep-water distal carbonate ramp and prograded across the Sappington Formation following regional tectonic forcing and global transgression. Lodgepole-correlative units in Canada include the lower Banff Formation above the Exshaw Formation (Smith & Bustin 2000; Hartel et al. 2014; Gaswirth & Marra 2015). Diverse transgressive–regressive (T–R) sequence stratigraphic interpretations of the Sappington Formation have been proposed, based on facies shifts, stacking patterns and surfaces with erosional lags (Doughty & Grader 2010; Adiguzel et al. 2012; Schietinger 2013; Nagase 2014; Rodriguez 2014; Phelps 2015; Myrow et al. 2015). Sea-level fluctuations produced temporally discontinuous stratigraphic sections in which incomplete, punctuated biostratigraphic records occur. Early palynological contributions with illustration of palynomorphs are known from Hardscrabble (Peak 9559) in the Bridger Mountains (Achauer 1959; Sandberg et al. 1972). From Unit 1 of the Sappington Formation and the basal Lodgepole Formation, Achauer (1959) found two assemblages with acritarchs and a few spores, and Sandberg et al. (1972) recorded *R. lepidophyta*-bearing Sappington Unit 4 bracketed by the conodont *S. praesulcata* Zone found in Sappington Units 3 and 5 at Hardscrabble and other locations in south-western Montana (Figure 3).

Conodonts from the Three Forks Formation (Trident Member) indicate the presence of middle Famennian *marginifera/trachytera* zones and the *expansa* Zone in Sappington Unit 1 (see Morgridge 1954; Sandberg 1965, 1976; Klapper 1966; Sandberg 1979; Sandberg & Ziegler 1979; Sandberg et al. 1989). Although, the DCB has been placed below Unit 2 (e.g. Gutschick & Rodriguez 1967) or near the top of Unit 5 (Sandberg & Klapper 1967), others have placed the DCB within Unit 4, the Sappington Siltstone 'middle shale', based on conodonts (Klapper 1966; MacQueen & Sandberg 1970; Sandberg 1979). Early conclusions that most of the Sappington Formation was

'Early Mississippian' (Tournaisian) based on conodonts were promoted by Gutschick & Perry (1957, 1959), Gutschick (1964) and Rodriguez & Gutschick (1967). This conclusion affected ages of brachiopod occurrences of the Sappington Formation summarised by Gutschick & Rodriguez (1967). Thus, considering brachiopods have not been well studied in this region (e.g. Thrasher 1992), more detailed fieldwork is required to make collections for biostratigraphic and sequence-stratigraphic purposes. Other macrofossils described from near the DCB include ammonoids (Korn & Titus 2006), sponges (Gutschick & Perry 1959), foraminifera (Gutschick 1962) and medusoid hydrozoans (Gutschick & Rodriguez 1990).

An unconformable contact between the upper Sappington and Lodgepole formations in Montana was proposed by Sandberg & Klapper (1967), due to the record of *S. praesulcata* Zone in middle Sappington and Tournaisian *duplicata-sandbergi-crenulata* Zones in the CCM and basal Lodgepole Formation (see also MacQueen & Sandberg 1970; Sandberg et al. 1972; Sandberg 1979). Our early examinations of conodonts over this interval suggest that there are gaps in the succession. Therefore, the results presented in this contribution are part of a high-resolution biostratigraphic study of Sappington units to recalibrate their ages as a low-latitude proxy for timings of the DCB Gondwana glaciation events and to better understand correlations with other latest Devonian–Mississippian strata in the region (i.e. the Bakken).

## 2.2. Sappington Formation at the Logan Gulch reference section

The late Famennian/early Tournaisian interval (DCB) at Logan Gulch on the northern bank of the Gallatin River (Figures 1–3) is represented by the Three Forks (upper Trident Member), Sappington and Lodgepole formations (Sandberg 1965). The Three Forks Formation at its Logan Gulch type section was originally described as comprised of three members: the Logan Gulch, Trident and Sappington (Figure 2; Sandberg 1962; Klapper 1966; Sandberg & Klapper 1967; Korn & Titus 2006). However, Berry (1943) recognised that a major SB separates the Three Forks from the Sappington, which led him to define the Sappington Formation with its own type section at Milligan Canyon.

At the Sappington reference section at Logan Gulch, the basal Sappington marine shale with chert (units 1A and 1B) is missing (Figures 2–4), making Unit 1 anomalously thin. Only units 1C and 1D are present (Gutschick et al. 1962). Unit 1C is a conodont-, inarticulate brachiopod- and conchostracan (spine-caudatan)-bearing black shale overlain by a more oxidised, and crinoid-bearing, light green shale (Unit 1D). As outlined by Gutschick et al. (1962) the more proximal black shale Unit 1C usually overlies more distal marine shales (units 1A and 1B), but at Logan Gulch 1C rests on an irregular unconformity comprised of two merged sequence boundaries (SB) above the 'Trident Limestone' (competent wackestone–packstone beds at the top of the Three Forks Trident Member).

Unit 1D regionally underlies Unit 2 carbonate and siltstone with conspicuous large, ferruginous, microbially wrapped (oncolitic) bioclasts and sponges. This limestone has large brachiopods, crinoids, rare bryozoans and corals. These shales of different environments and different age (i.e. lower *expansa*

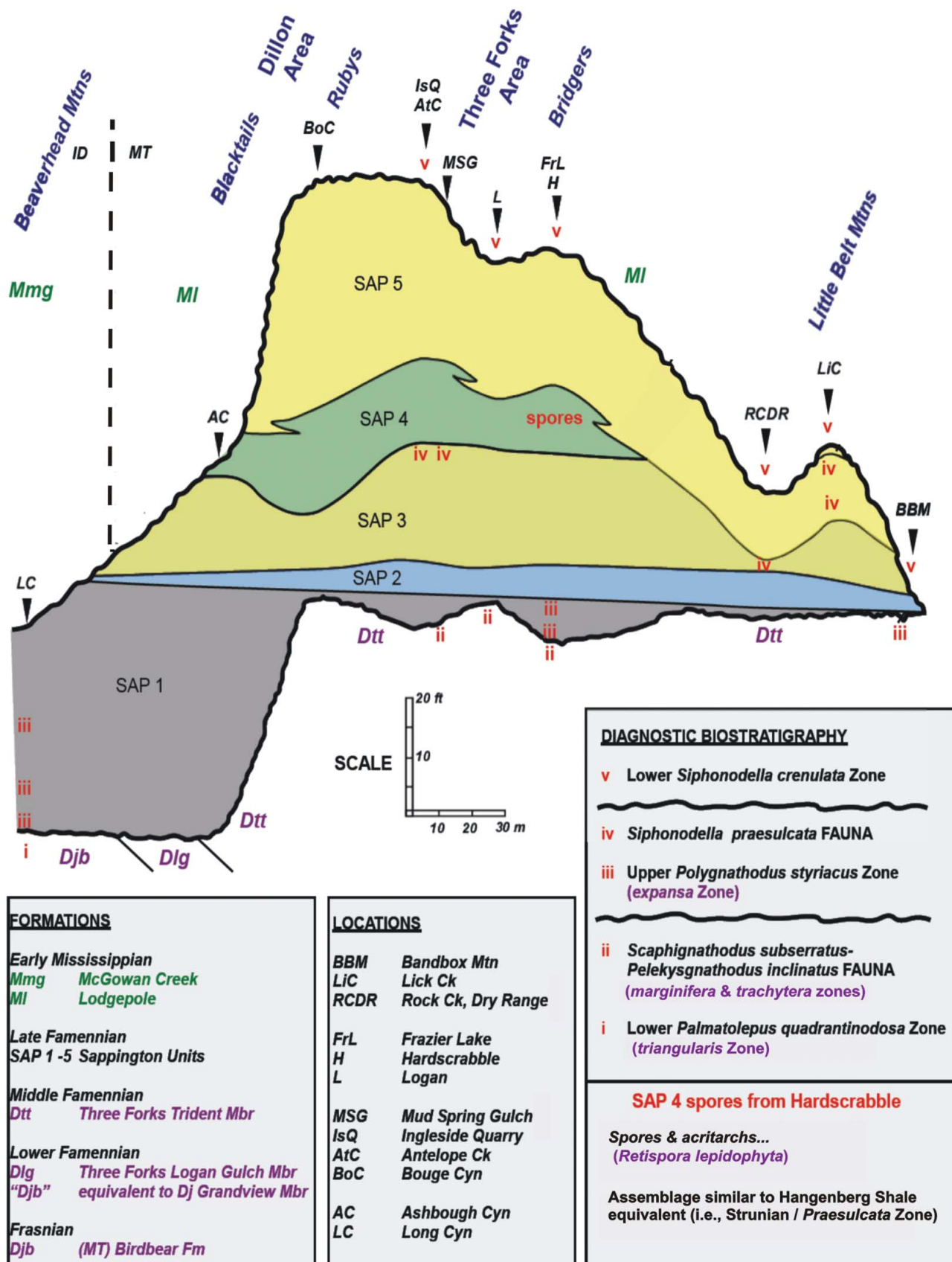


Figure 3. Cross-section showing conodont and spore collections and depositional/erosional lacunae in the Sappington Formation from the Beaverhead Mountains (ID) to the Little Belt Mountains (MT). Modified from Sandberg et al. (1972).

Zone units 1A and B versus upper *expansa* Zone units 1C and D; Sandberg et al. 1972) define an SB (Figure 2). Shale units 1C and 1D are widely present at nearby locations, where underlying black shales Unit 1A and B are also quite thick (e.g. Milligan Canyon and Nixon Gulch). The intra-black shale contact between units 1A/B and Unit 1C is known to have a phosphatic, very thin, sandstone lag (Klapper 1966; Myrow et al. 2015). Such lags are known in the Lower Bakken and are rarely observed in Sappington outcrop (e.g. Frazier Lake), and represent condensed deposition above an SB. A transgressive surface of non-erosion is defined between the silty oncolitic limestone of Unit 2 and the top of Unit 1 (Myrow et al. 2015), although most Bakken workers identify this surface as an SB because of an interpreted basinward shift in facies (calcareous silt over black shale). Up-section, a similar regional surface under the basal Lodgepole Formation Scallion packstones has also been given different interpretations (i.e. SB, transgressive surface, or maximum flooding surface), needing further biostratigraphic effort to clarify (Figure 2).

Unit 3 overlies Unit 2 gradationally, coarsens very slightly upward, and is interpreted as a lower shoreface siltstone, with distal storm lenses, burrows, shelly marine fossils and limestone beds. Unit 4 sharply overlies Unit 3 and is comprised primarily of bioturbated to burrowed green shale with thin rippled silt lenses and beds. Unit 4 is relatively thin at Logan Gulch (3 m) compared to thicker occurrences immediately west and north (8 m); it is dominated by an *Arenicolites* ichnofacies characterised by *Bifungites fezzanus* (Gutschick & Lanborn 1975; or *Diplocraterion* in some current publications) with abundant *Teichichnus* upwards where the unit has more siltstone, with less starved ripples, and larger silty, wave-rippled bedforms with muddy interbeds. This ichnofacies also occurs in units 3 and 5 (Figure 2), where siltstone and sandstone occur with much larger sedimentary structures and bedforms, marine fossils and more diverse burrows including *Planolites*, *Helminthopsis*, *Nereites* and *Lockeia*. At most central CMT locations, Unit 3 represents more distal shoreface environments that shallow upward.

Unit 5 is comprised of storm beds and wave rippled siltstones in more proximal shoreface environments. Foreshore and eolian backshore environments are present to the east. At Logan Gulch, Unit 5 sharply overlies Unit 4 and represents relatively continuous deposition of middle shoreface silt and very fine sandstone in stacked trough ripples (Figure 4F). This relationship remains consistent through much of the central and eastern portions of the CMT. The Sappington Formation thickens towards the west, ultimately losing the well-defined Unit 4, as units 3 and 5 become finer grained and less easily differentiated.

The overlying black shale of the upper CCM of the Lodgepole Formation (Figures 2, 4E) occurs with a basal sand lag with chert, bone, teeth and conodonts. CCM as used in this paper is synonymous with Sappington Formation 'Unit 1' (Gutschick et al. 1962), or Unit 6 after Doughty & Grader (2010). This primarily shale unit is more burrowed and has a higher silt content up-section. *Zoophycus* traces are common in the upper part of this interval and do not occur lower in the Sappington Formation. Basal grey crinoidal packstone–grainstone with glauconite of the Lodgepole Scallion Member overlies the CCM and thins

to the west. Some black shale ('false Bakken') occurs above the Scallion Member and is correlative laterally with thin-bedded calcareous slope deposits with abundant *Helminthopsis* in the overlying argillaceous Lodgepole Formation.

### 3. Materials and methods

#### 3.1. Prior palynological sampling

Since 2011, an extensive sampling for palynological analysis of the DCB from different locations in western Montana was carried out by the University of Idaho working together with PRI-SEM Geoconsulting, specifically to establish the age of the Three Forks, Sappington (units 1, 4) and Lodgepole formations (di Pasquo et al. 2012; Warren et al. 2014; Warren 2015). In 2011, the Trident Member (Three Forks Formation), units 1 and 4 of the Sappington and the basal Lodgepole (CCM) formations were sampled at Logan Gulch (eight samples), Milligan Canyon (three), Milligan Canyon (East) (seven), Red Hill (four), Brown Back Gulch (two) and Lewis and Clark Caverns State Park (two) (Figure 1). The results from four productive samples of the Trident Member and two of the Lodgepole Formation at Logan Gulch, and from two samples of the Trident Member at Milligan East, were summarised by di Pasquo et al. (2012). Unfortunately, at that time two barren samples were obtained from Unit 1 and Unit 4, Sappington Formation, whilst a cephalopod mold was collected (MdP in 2011) from the upper Trident Member at Logan Gulch (Figure 4B).

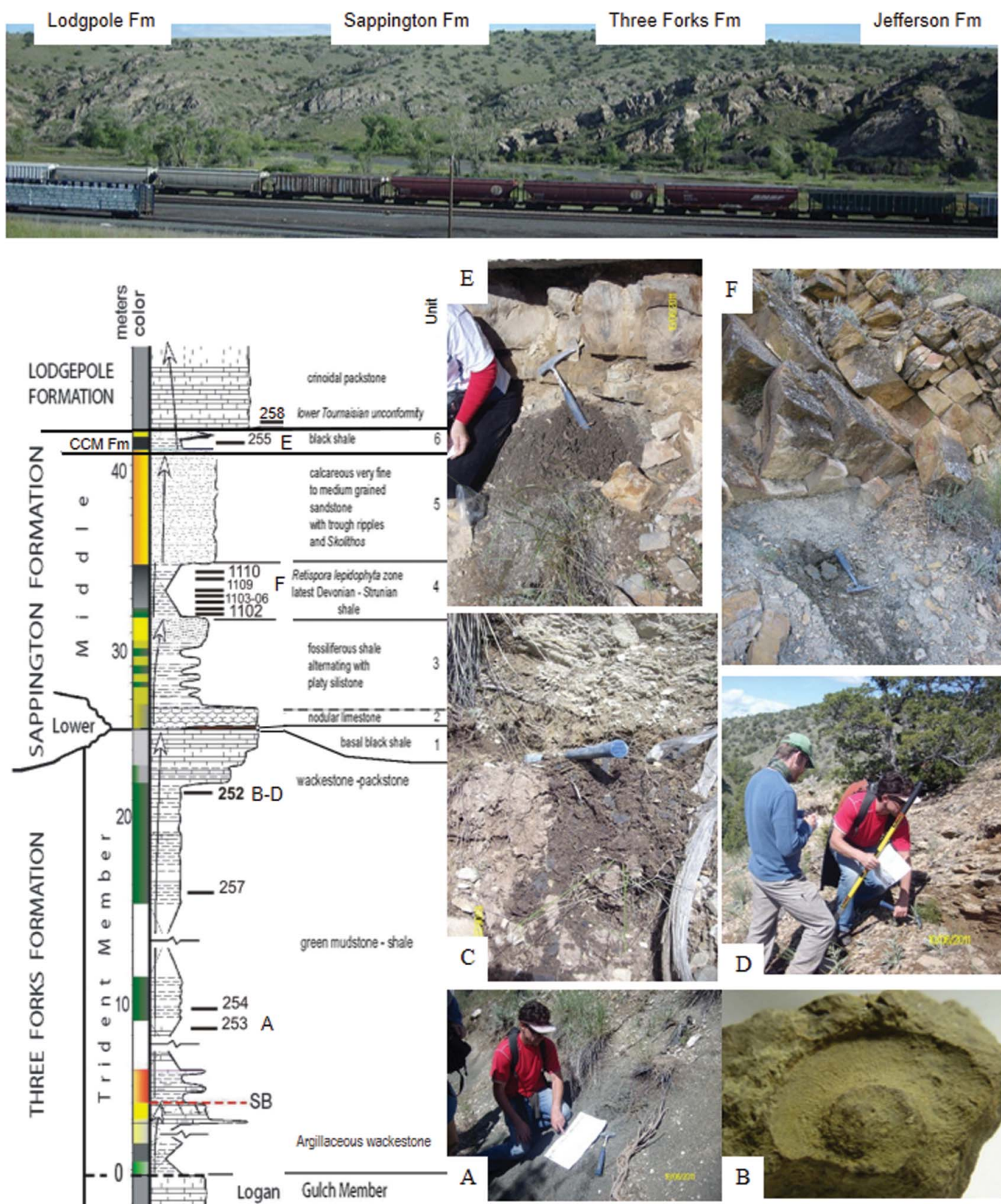
In 2013, 23 samples were collected systematically at Logan Gulch from units 1 and 4 (15 samples), and CCM (seven). Unit 4 was sampled at 20-cm intervals; of them, seven samples from above 60 cm from its base were productive, whereas the remaining samples from this unit and the other two were barren. Unit 1 was systematically sampled at Antelope Creek (10 samples) and Nixon Gulch (15) but all samples were barren. These samples were processed and studied by Mercedes di Pasquo in 2014 and the results were summarised by Warren et al. (2014). In 2014, five samples were collected at Moose Creek – Unit 1 (one sample), Unit 4 (one), CCM (three) – and seven samples at Trident – Unit 1 (three samples), Unit 4 (three), CCM (one). Preliminary results were presented by Warren (2015) indicating that samples from Unit 4 and CCM at Moose Creek were productive, and samples from Trident were barren.

#### 3.2. Palynologic samples studied from Logan Gulch

The stratigraphic column illustrating the lithostratigraphic nomenclature, weathering profile and rock types, which are described in Section 2.2, and the productive levels studied in detail herein are indicated in Figure 4.

#### 3.3. Processing methods and repository

Samples were processed using standard palynological methods (hydrochloric acid [HCl] and hydrofluoric acid [HF], with distilled water neutralisation) at the Laboratory of Palynostratigraphy and Paleobotany of the CICYTTP-CONICET in 2011 and 2014. All the residues were sieved with 25- $\mu$ m and 10- $\mu$ m mesh and slides were mounted with glycerin jelly. Some of the residues



**Figure 4.** A panoramic photograph of the Logan Gulch railroad exposure and a stratigraphic column illustrating the lithostratigraphic nomenclature, weathering profile, sampling levels and a brief description of rock types.

A. Green shales of Trident Member ca. at 10 m (CICYTTP-PI 253). B. Clymenid (external mold, maximum diameter 4 cm) from green shale/fine sandstone at ca. 20 m (sample CICYTTP-PI 252) of Trident Member (detail in C). C–D. Two shale beds of the Trident Member at ca. 20 m. C. In detail, the reddish bed is above the green shale (CICYTTP-PI 252). D. General picture with people for scale. E. Black shale of the CCM Lodgepole Formation (CICYTTP-PI 255). F. Green shale of Unit 4 of the Sappington Formation showing the contact with sandstone Unit 5.

were also treated with HF a second time to eliminate extra quartz particles (equivalent to separation with heavy liquids as zinc chloride [ZnCl<sub>2</sub>]) and boiled in HCl (to eliminate amorphous organic matter). Two residues (a half part of CICYTTP-PI 255 and 258; Figure 4E) were oxidised with nitric acid for 2 minutes; the reaction was stopped with ammonium hydroxide and the residue immediately washed with distilled water several times in the centrifuge (see Traverse 2007).

Slides and residues are housed at the Palynostratigraphy and Palaeobotany Laboratory of the CICYTTP and catalogued under the CICYTTP-PI acronym (England Finder<sup>TM</sup> coordinates for position of palynomorphs). Identification and illustration of palynomorphs were achieved using transmitted light microscopes and digital video cameras (Nikon Eclipse E200 with a Labomed (5.0 MP) and a Leica DM500 with a fluorescent LED lamp (cold light) attached with a fluorescence (ca. 450 nm) filterblock, and a Leica EC3 3.0 MP video camera) at the CICYTTP. A brief analysis of the autofluorescence of palynomorphs was carried out on the slides. The palynomorph images were improved (i.e. the background was cleaned up) with Photoshop®.

Taxonomic analysis and stratigraphic distribution (SD) of palynomorphs were obtained from specific literature, mainly from the Northern Hemisphere (e.g. Wicander 1974; Wicander & Loeblich 1977; Playford & McGregor 1993; Wicander & Playford 2013, and references therein). The cephalopod mold (Figure 4B) is housed in the Invertebrate collection at the CICYTTP (CICYTTP-In acronym) and was illustrated using a Panasonic camera (12 MP).

## 4. Results

Thirteen samples from the Trident Member (Three Forks Formation), Unit 4 of the middle Sappington and the basal Lodgepole formations (Figure 4) yielded 87 total spore and phytoplankton species and some scolecodonts. Of these, 27 species are prasinophytes and acritarchs (marine origin), 59 are trilete spores, and one is the chlorophycean *Botryococcus braunii* (fresh/brackish origin). Seven samples of the lower Lodgepole yielded abundant phytoclasts and/or amorphous organic matter (di Pasquo et al. 2012; Warren et al. 2014; Warren 2015). Microplankton species (*Botryococcus*, acritarchs and leiosphaerids) exhibited light orange autofluorescence, whereas spores showed dark grey to black.

Five Logan assemblages (LAs1 to LAs5) are characterised by the SD of the species documented in the Logan Gulch section (Figure 5). A taxonomic list of species is included below, under each assemblage. Taxonomic revision of the species (i.e. synonymy and combinations) will be appearing in a future contribution. Their occurrences, as given herein, are based on detailed descriptions along with illustrations from the literature (Figure 6). Selected species are illustrated in Plates 1–5. The species found in the Hardscrabble section (Figure 1) by Sandberg et al. (1972) shared with LAs3 are cited in this section under SD, and a list of all the species illustrated and not illustrated by those authors is included in Appendix 1 at the end of this work. The chronostratigraphic, stratigraphic and biostratigraphic correlation of the Late Famennian–Early Tournaisian palynoassemblages at Logan Gulch (Montana) is shown in Figure 7. The usage of the Strunian as equivalent to the uppermost Devonian was not formally

accepted yet (see Geological Time Scale after Cohen et al. 2014; Becker et al. 2016). Nevertheless, we prefer to use this name in the SD of the list of species as corresponding to uppermost Famennian. The first appearance of *R. lepidophyta* in the Ardennes–Rhein basins where this name originated defines the base of the Strunian, and its base is at the base of the upper *expansa* conodont zone (Streel & Loboziak 2000). Streel et al. (2006) reiterated the proposal to divide the Famennian into four substages including the Strunian, presenting a correlation between microspore zones from neritic to terrestrial environments.

### 4.1. LAs1 Assemblage – the Trident Member of the Three Forks Formation

The Trident Member (Figure 4A) yielded a low-diversity palynomorph assemblage (LAs1), dominated by cosmopolitan, stratigraphically long-ranging phytoplankton species (see below) that exhibit reasonably good preservation and a –2/+2 thermal alteration index (TAI, Utting & Wielens 1992). The most abundant species (frequency of specimens) in all levels are *Gorgonisphaeridium ohioense*, *G. plerispinosum* and leiosphaerids (many different morphotypes as in the LAs2), whereas *Ammonidium loriferum/garrasinoi* (poorly preserved due to fragmentation) and *Unellium lunatum* are less abundant (Plate 1). Scolecodonts were also found (Plate 2). Spores were scarce at one level (Figure 5).

#### SPORES

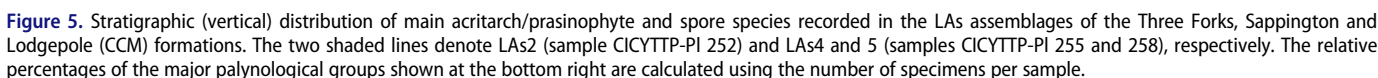
*Auroraspora macra* Sullivan 1968. SD: Famennian (Playford 1976; Coleman & Clayton 1987; Utting 1987a, 1987b; Higgs et al. 1988; McNestry 1988; McGregor & Playford 1992; Filipiak 2004; di Pasquo 2007; Marynowski & Filipiak 2007; Amenábar et al. 2009; Racka et al. 2010; Wicander & Playford 2013; di Pasquo et al. 2015); Frasnian–Visean (Playford & Melo 2012); Mississippian (Higgs et al. 1988; Filipiak 2004; Azcuy & di Pasquo 2005, 2006).

*Grandispora echinata* Hacquebard 1957. Remarks: *G. clandestina* Playford 1976 is here considered part of this taxon due to the presence of transitional forms between them. Vigran et al. (1999) illustrated *Spelaeotriletes resolutus* (pl. 1, fig. 7) and *Spelaeotriletes microspinosus* (pl. 1, fig. 12) that are also considered herein to be *Grandispora echinata*. SD: Mid-Famennian–Mississippian (Sandberg et al. 1972; Higgs 1975; Playford 1976, 1993; Molyneux et al. 1984; Coleman & Clayton 1987; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Lu & Wicander 1988; McGregor & Playford 1992; Avchimovitch 1993; Playford & McGregor 1993; Vigran et al. 1999; Filipiak 2004; Marynowski & Filipiak 2007; Filipiak & Racka 2010; Racka et al. 2010; Higgs et al. 2013).

*Retusotriletes incohatus* Sullivan 1964. SD: Famennian–Tournaisian (Sandberg et al. 1972; Coleman & Clayton 1987; Utting 1987a, 1987b; Avchimovitch et al. 1988; Higgs et al. 1988; McNestry 1988; Playford 1993; Playford & McGregor 1993; Vigran et al. 1999; Filipiak 2004; Ghavidel-syooki & Owens 2007; Marynowski & Filipiak 2007; Playford & Melo 2012; Wicander & Playford 2013).

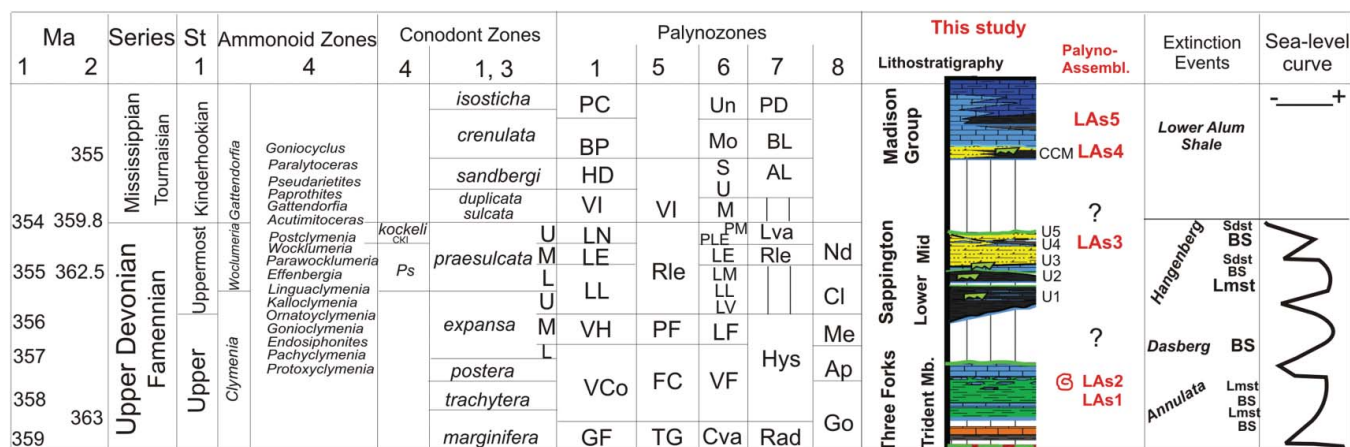
#### ACRITARCHS/PRASINOPHYTES

*Ammonidium loriferum* (Deunff) Hashemi & Playford 1998–A. *garrasinoi* Ottone 1996 (morphon) (Plate 1, figure 1).



processes with distally multifurcate branches up to third order. SD: Givetian–Famennian (Ottone 1996; Hashemi & Playford 1998; Racka et al. 2010; Noetinger & di Pasquo 2011; di Pasquo et al. 2015).

**Figure 6.** Global stratigraphic ranges of selected microplankton and spore species present at the Logan Gulch locality based on selected literature (refer to those below and additional references in the text). Those species highlighted are of particular biostratigraphic significance. The shaded band indicates the latest Devonian corresponding to Las3. Latest Devonian: Sandberg et al. (1972); Playford (1976, 1981); Wicander & Loeblich (1977); Molyneux et al. (1984, 1996); Wicander & Playford (1985, 2013); McGregor & Playford (1992); Hartkopf-Fröder & Streehl (1994); Ottone (1996); Melo et al. (1999); Le Hérisse et al. (2000); Maziane et al. (2002); Hashemi & Playford (2005); Racka et al. (2010); Marynowski et al. (2012). Devonian–Carboniferous: Wicander (1974, 1975); Higgs (1975); Lu & Wicander (1988); Higgs et al. (1988); Avchimonvitch et al. (1988, 1993); Playford (1993); Playford & McGregory (1993); Vavrdová et al. (1996); Díaz Martínez et al. (1999); Melo & Loboziak (2003); Filipiak (2004, 2005); González et al. (2005a, 2005b); Ghavidel-syooki & Owens (2007); Amenábar et al. (2009). Mississippian (Carboniferous): Smith & Butterworth (1967); Utting (1987a, 1987b); Azcuay & di Pasquo (2005, 2006); di Pasquo (2007); Heal & Clayton (2008); Amenábar et al. (2009); Melo & Playford (2012).



**Figure 7.** Chronostratigraphic, stratigraphic and biostratigraphic correlation chart of the Late Famennian–Early Tournaisian at Logan Gulch (Montana). Erosional unconformity at the boundary is shown. Sea level curve modified here from Johnson et al. (1985, in Streele et al. 2000). For extinction events, see Racka et al. (2010); Kaiser et al. (2011, 2015); Marynowski et al. (2012); Ma et al. (2014); Becker et al. (2016). References: 1. Coleman & Clayton (1987), Streele et al. (1987, 2000), Playford & McGregor (1993), Maziane et al. (2002), González et al. (2005a, 2005b), Kaiser et al. (2015). 2. Trapp et al. (2004), Cohen et al. (2014). 3. Hartel et al. (2014). 4. Kaiser et al. (2011, 2015), Becker et al. (2016). Note that Sandberg & Klapper (1967) placed *Wocklumeria* in the Strunian (D–C boundary) in the absence of conodonts (Streele et al. 2006). Spore zonation: 5. Richardson & Ahmed (1988). 6. Avchimovitch in Filipiak (2004). 7. Melo & Loboziak (2003), Melo & Playford (2012). Acritarch zonation: 8. Molyneux et al. (1996), Le Hérisse et al. (2000). Lithostratigraphic section not to scale. Abbreviations: palynozones: GF = *Grandispora gracilis*–*G. famenensis*. VCo = *Diducites versabilis*–*Grandispora cornuta*. PF = *Vallatisporites pusillites*–*Apiculiretusispora fructuosa*. Rle = *R. lepidophyta*. Cva = *Cornispora varicornata*. VF = *Diducites versabilis*–*G. famenensis*. LF = *R. lepidophyta*–*G. facilis*. LV = *R. lepidophyta*–*A. verrucosa*. LMB = *R. lepidophyta*–*Tumulispora mirabilis*. PLE = *V. pusillites*–*R. lepidophyta*–*K. explanatus*. PM = *V. pusillites*–*T. malevkensis*. M = *T. malevkensis*. U = *Grandispora upensis*. S = *Apiculiretusispora septalia*. Mo = *Potonieisporites monotuberculatus*. Un = *Pustulatisporites uncinatus*. Rad = *Rugospora radiata*. Hys = *Vallatisporites hystricosus*. Lva = *R. lepidophyta*–*V. vallatus*. AL = *Radiizonates arcuatus*–*Waltzispora lanzonii*. BL = *Spelaotrilites balteatus*–*Neoraistrickia loganensis*. PD = *Spelaotrilites pretiosus*–*Colatisporites decorus*. Go = *Gorgonisphaeridium ohioense*. Ap = *Aciora petala*. Me = *Muraticavea enteichia*. CI = *Cymatiosphaera labyrinthica*. Nd = *Navifusa drosera*. CKI = *costatus*–*kockeli* interregnum. Ps = *Praesulcata*. Note: The Uppermost Famennian is equivalent to the informal ‘Strunian’ name not formally accepted yet (see Geological Time Scale after Cohen et al. 2014; Becker et al. 2016).

1993; Wicander & Loeblich 1977; Lu & Wicander 1988; Filipiak 2005; Wicander & Playford 2013).

*Gorgonisphaeridium evexispinosum* Wicander 1974 (Plate 1, figure 6). SD: Frasnian–Famennian (Wicander 1974, 1975).

*Gorgonisphaeridium ohioense* (Winslow) Wicander 1974 (Plate 1, figure 2). SD: upper Frasnian–Strunian (Wicander 1975; Playford 1976, 1993; Wicander & Loeblich 1977; Molyneux et al. 1984; Lu & Wicander 1988; Playford & McGregor 1993; Vavrdová et al. 1993; Hashemi & Playford 1998; Le Hérisse et al. 2000; Filipiak 2005; González et al. 2005a; Marynowski & Filipiak 2007; Marynowski et al. 2010; Racka et al. 2010; Wicander & Playford 2013; di Pasquo et al. 2015).

*Gorgonisphaeridium plerispinosum* Wicander 1974 (Plate 1, figure 5). SD: upper Frasnian–Famennian (Wicander 1974, 1975; Wicander & Loeblich 1977; Lu & Wicander 1988; Playford 1993; Playford & McGregor 1993; Hashemi & Playford 1998; González et al. 2005a; Wicander & Playford 2013).

*Leiosphaeridia*–Tasmanitid group (Plate 2, figures 1–6). See description in Section 4.2.

*Michrystidium coronatum* Stockmans & Willièrè 1963 (Plate 1, figure 11). SD: Silurian–Devonian (Wicander & Loeblich 1977; McNestry 1988).

*Stellinium micropolygonale* (Stockmans & Willièrè) Playford 1977 (Plate 1, figure 10). SD: Emsian–Strunian (Wicander 1974; Molyneux et al. 1984; Wicander & Playford 1985, 2013; Lu & Wicander 1988; McNestry 1988; Playford 1993; Playford & McGregor 1993; Ottone 1996; Hashemi & Playford 1998; Le Hérisse et al. 2000; González et al. 2005a; di Pasquo et al. 2009; Marynowski et al. 2010; Noetinger & di Pasquo 2011). Lowermost Carboniferous (McNestry 1988).

*Unellium lunatum* (Stockmans & Willièrè) Eisenack, Cramer & Díez 1979 (Plate 1, figure 14). Remarks: *U. winslowiae* is its junior synonym in González et al. 2005a. SD: Frasnian–Strunian (Wicander & Playford 1985, 2013; McNestry 1988; Lu & Wicander 1988 (lowermost Carboniferous); Playford 1993; Playford & McGregor 1993; Hashemi & Playford 1998; Le Hérisse et al. 2000; Filipiak 2005; González et al. 2005a; Marynowski et al. 2010, VF and LV zones).

*Unellium piriforme* Rauscher 1969 (Plate 1, figure 13). Remarks: *Unellium ampullium* Wicander 1974 is probably synonymous with this taxon. SD: Frasnian–Strunian (Wicander 1974; Wicander & Playford 1985; Lu & Wicander 1988; McNestry 1988; Playford 1993; Playford & McGregor 1993; Hashemi & Playford 1998; Le Hérisse et al. 2000; Filipiak 2005; Marynowski et al. 2010; Racka et al. 2010). Lowermost Carboniferous (McNestry 1988).

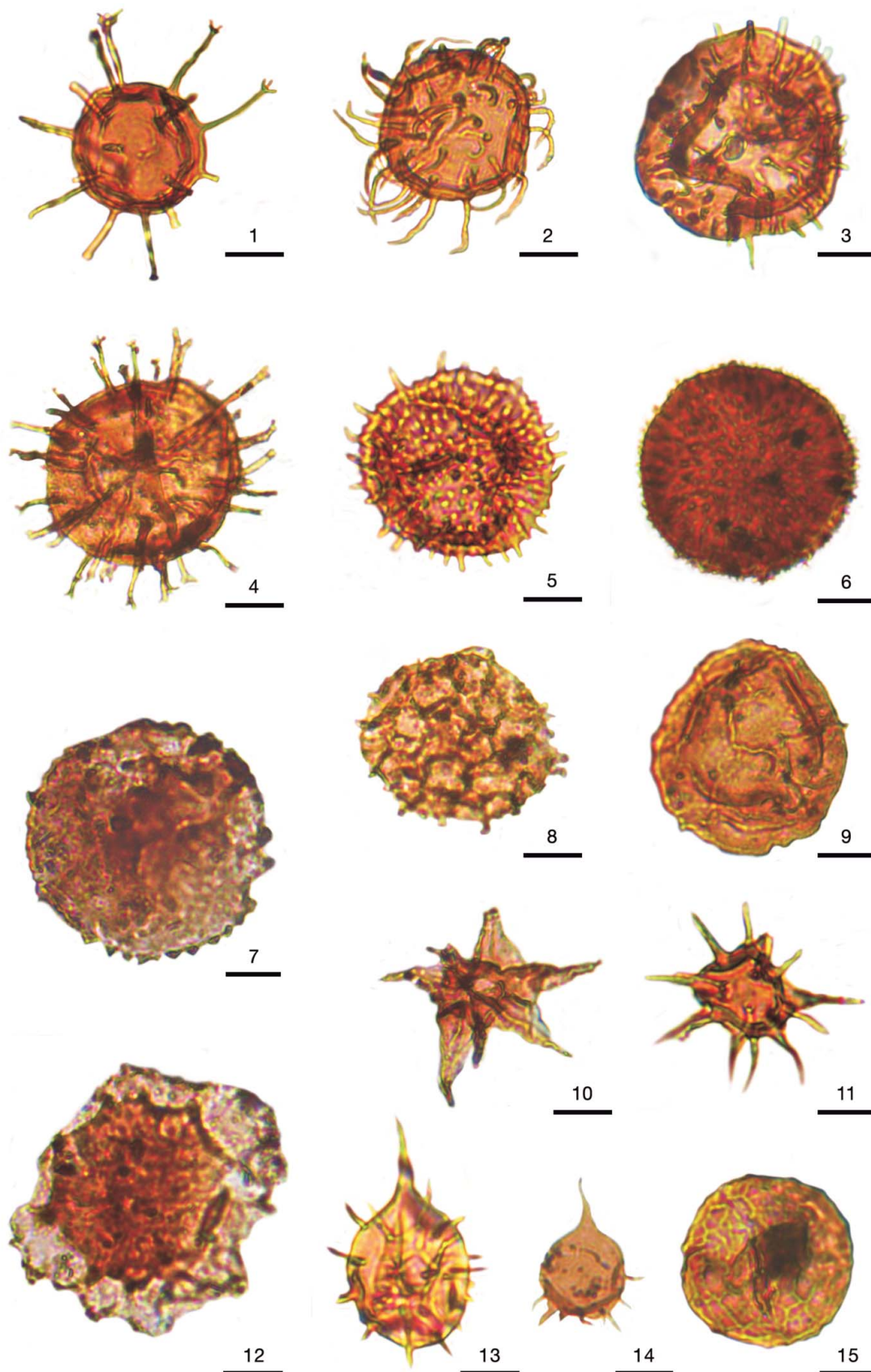
#### TAXA PRESENT, BUT NOT ILLUSTRATED

*Cymatiosphaera perimembrana* Staplin 1961. SD: Givetian–Tournaisian (Wicander & Playford 1985, 2013; Lu & Wicander 1988; Playford 1993; Playford & McGregor 1993; Ottone 1996; Hashemi & Playford 1998; Le Hérisse et al. 2000; Filipiak 2005; Amenábar et al. 2009; di Pasquo et al. 2009; Marynowski et al. 2010; Noetinger 2010).

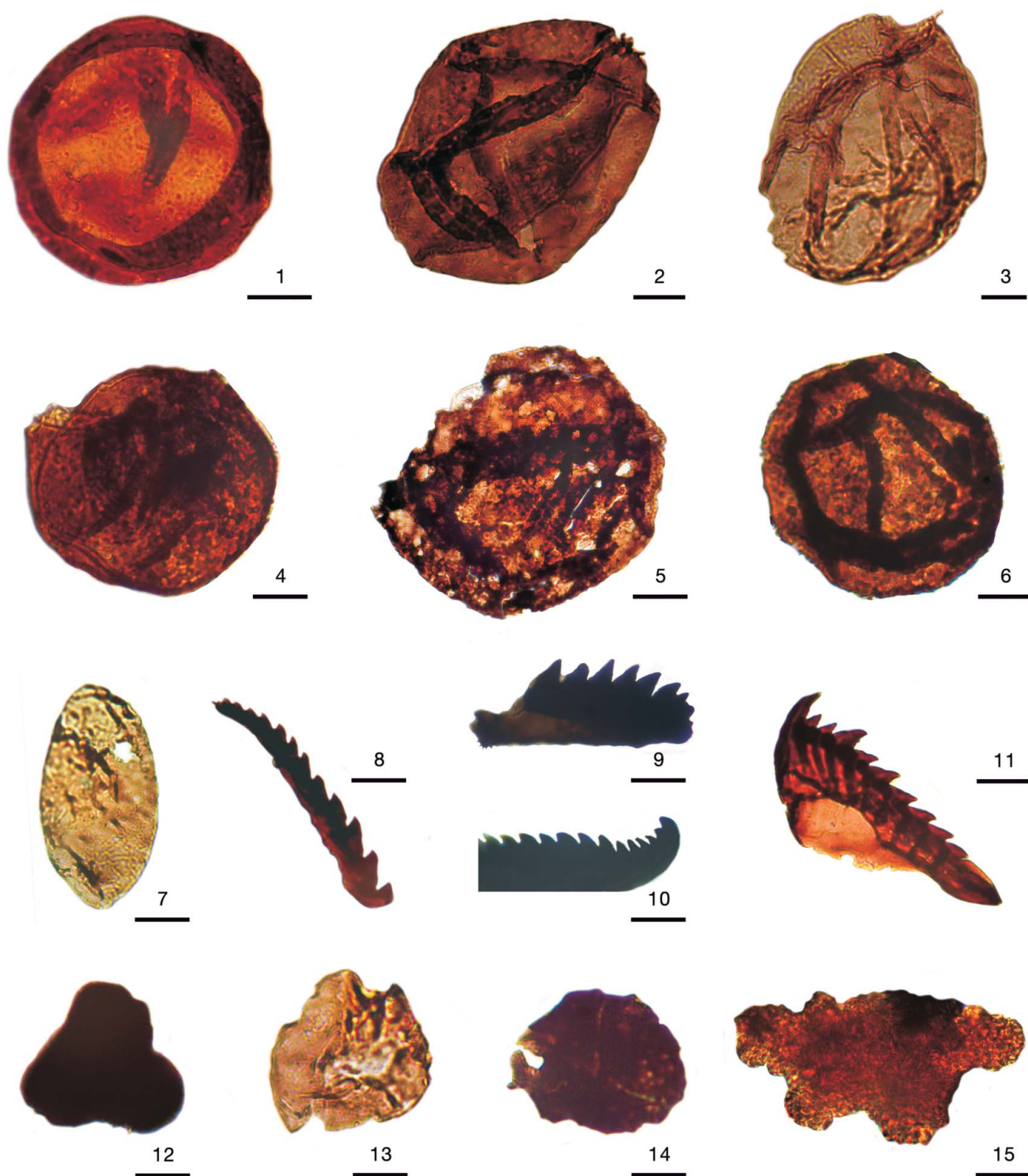
*Solisphaeridium astrum* Wicander 1974. SD: Givetian–Famennian (Wicander 1974, 1975; Wicander & Loeblich 1977; Playford & McGregor 1993).

*Veryhachium downei* Stockmans & Willièrè 1962

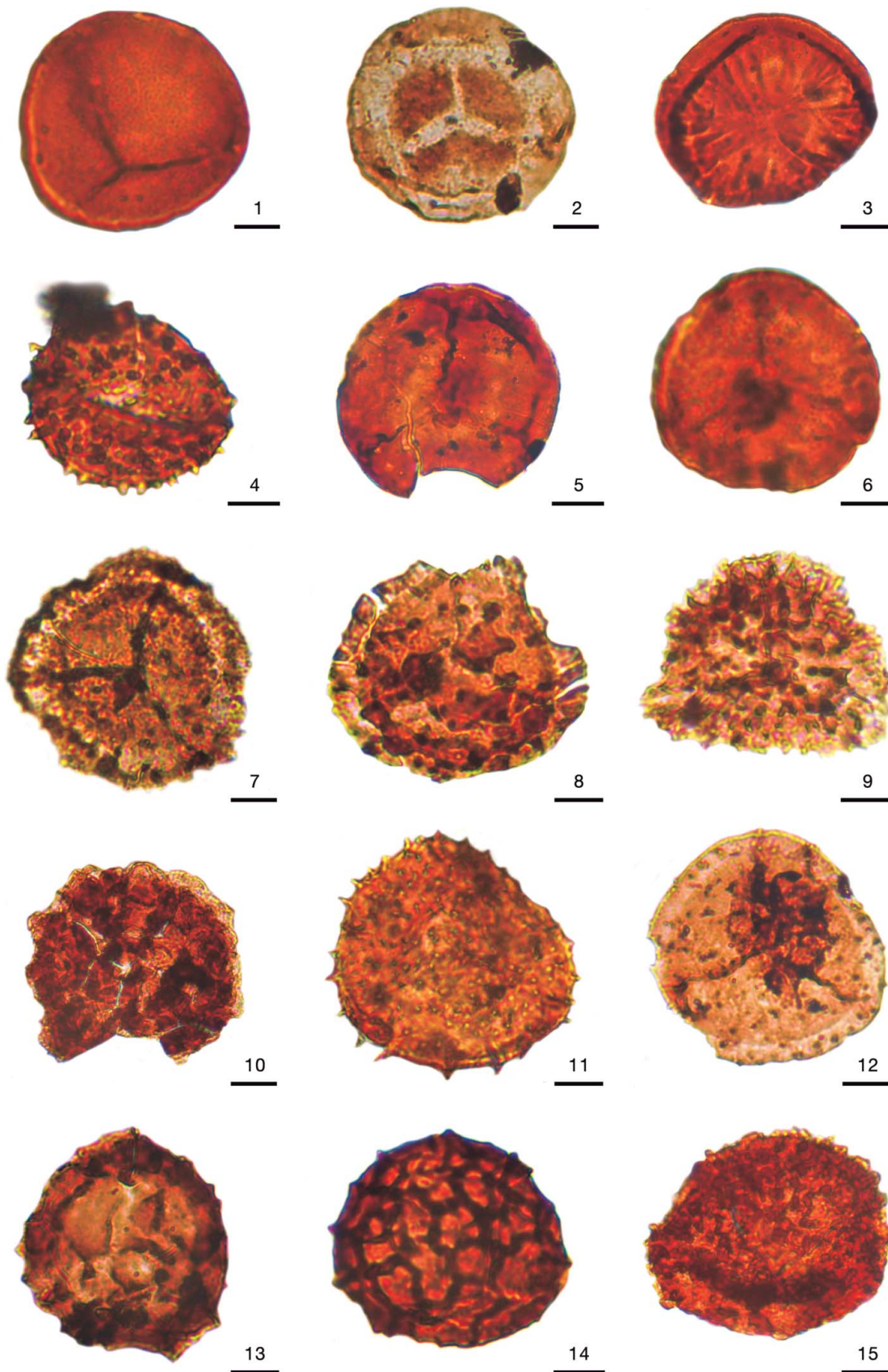
*Veryhachium polyaster* Staplin 1961



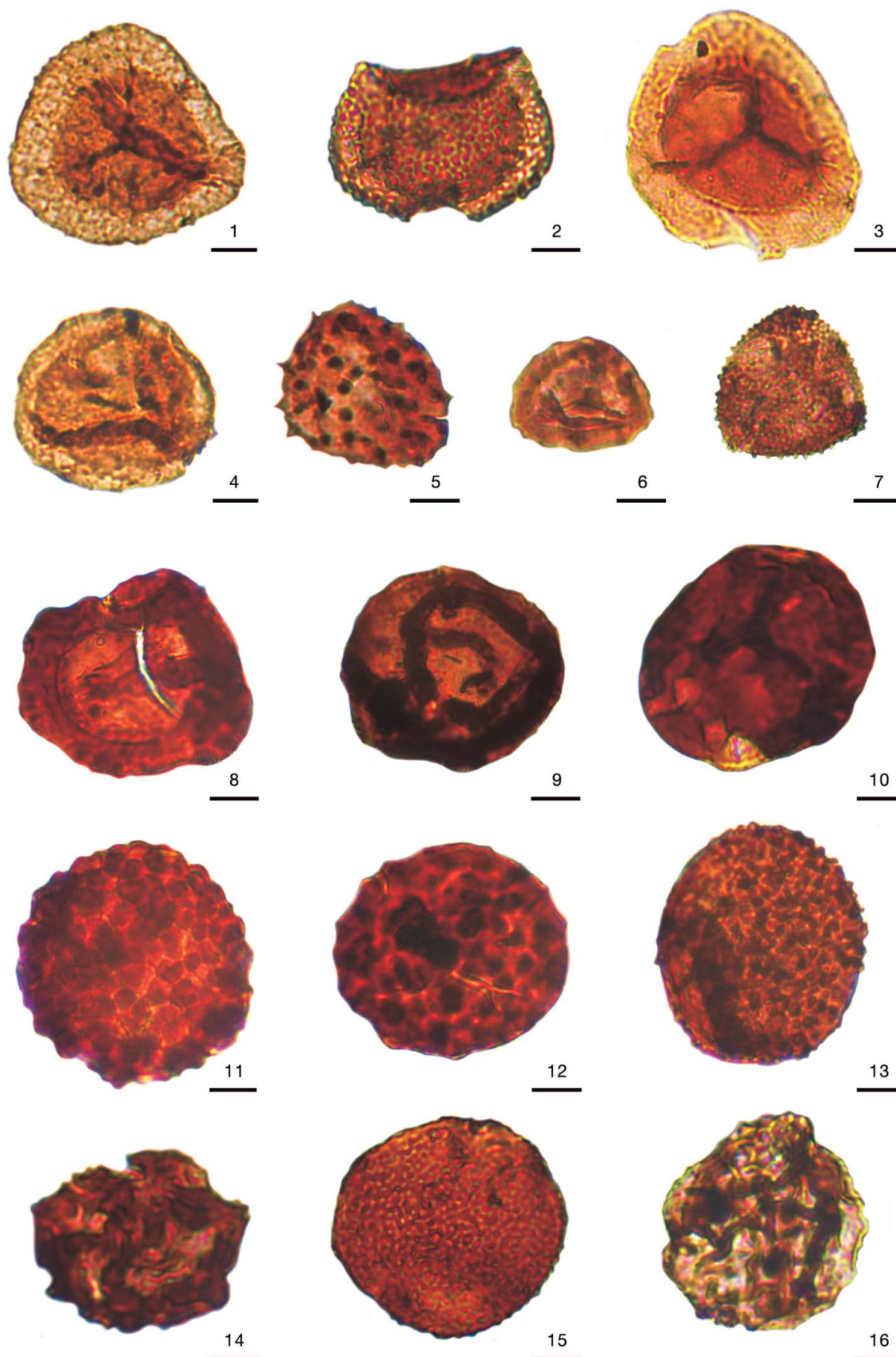
**Plate 1.** Phytoplankton species from Three Forks and Sappington formations at Logan Gulch. 1. *Ammonidium loriferum-garrasinoi*, CICYTTP-PI 253(1) EF S47/0. 2. *Gorgonisphaeridium ohioense*, CICYTTP-PI 257(1) EF W49/1. 3. *Gorgonisphaeridium absitum*, CICYTTP-PI 1110(HF2-HCI) EF B46/4. 4. *Gorgonisphaeridium winslowiae*, CICYTTP-PI 1102(HF2-2) EF V38/4. 5. *Gorgonisphaeridium plerispinosum*, CICYTTP-PI 253(2) EF P30/2. 6. *Gorgonisphaeridium evexispinosum*, CICYTTP-PI 1102(HF2-2) EF O57/0. 7. *Maranhites stockmansii*, CICYTTP-PI 1102(HF2-1) EF V58/3. 8. *Cymatiosphaera chelina*, CICYTTP-PI 1102(HF2-2) EF O48/2. 9. *Cymatiosphaera* sp., CICYTTP-PI 1102(HF2-2) EF O33/2. 10. *Stellinium micropolygonale*, CICYTTP-PI 1102(HF2-1) EF R52/2. 11. *Michrystidium coronatum*, CICYTTP-PI 253(2) EF O61/3. 12. *Maranhites moesii*, CICYTTP-PI 1102(HF2-1) EF Q52/4. 13. *Unellium piriforme*, CICYTTP-PI 253(1) EF O46/1. 14. *Unellium lunatum*, CICYTTP-PI 254(1) EF E50/1. 15. *Dictyotidium fairfieldense*, CICYTTP-PI 1110(HF2-HCI) EF G57/0. Scale bars: 1–4, 7, 9, 10, 12, 14 = 15 μm; 5, 6, 8, 11, 13, 15 = 10 μm.



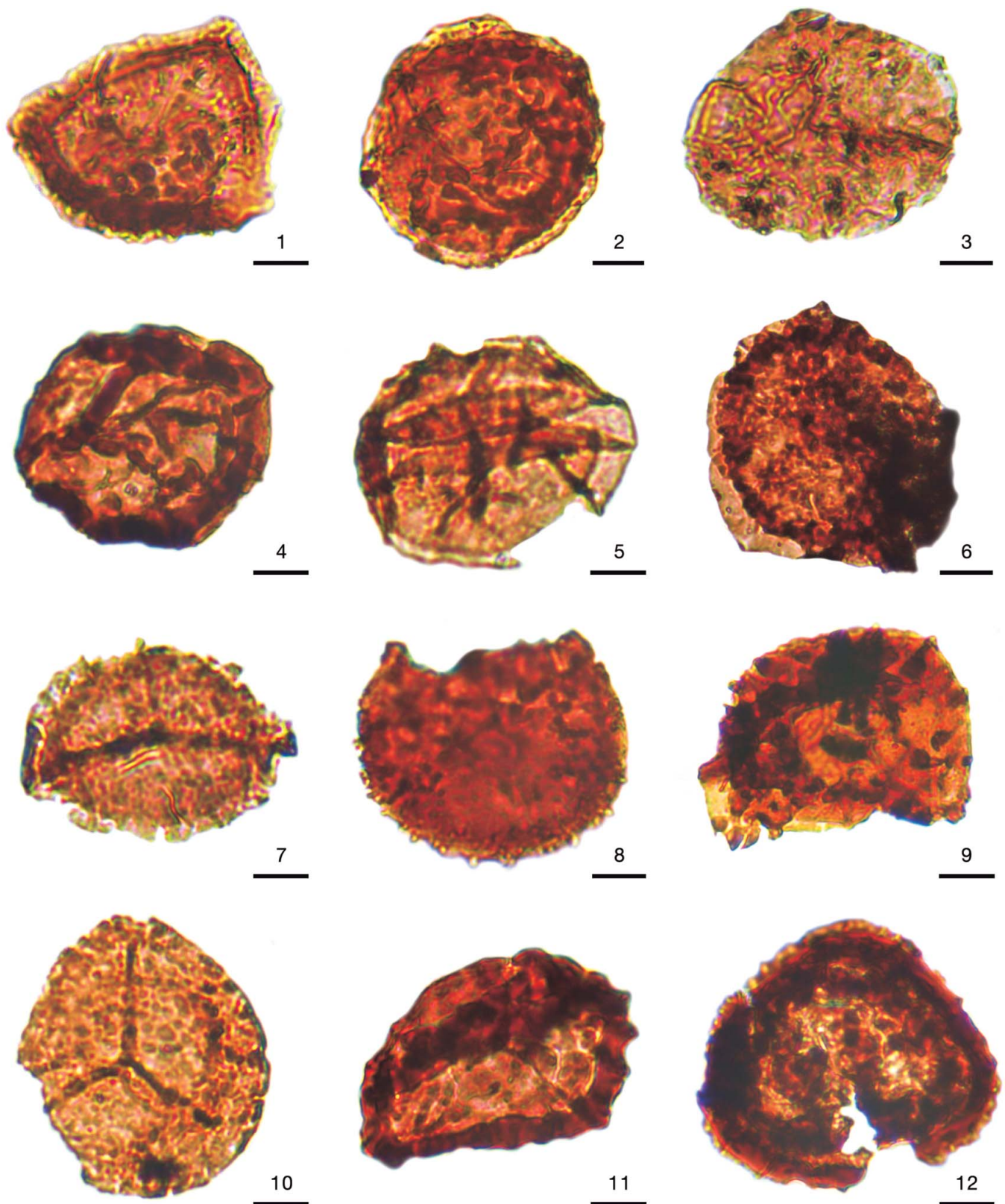
**Plate 2.** Palynomorphs from Three Forks, Sappington, Cottonwood Canyon Member (CCM) and Lodgepole formations at Logan Gulch. 1. *Tasmanites* sp., CICYTTP-PI 253 (2) EF J43/0. 2, 3–6. *Leiosphaeridia* group. 2. CICYTTP-PI 253(3) EF B66/3. 3. CICYTTP-PI 253(2) EF P52/2. 4, 6. Ornamented specimens with short, distally rounded processes. 4. CICYTTP-PI 253(2) EF M65/2. 6. CICYTTP-PI 1109(HF2-2) EF S47/4. 5. CICYTTP-PI 252(1) EF T30/4, poorly preserved, pyrite damaged. 7. *Navifusa/Deusillites* sp., CICYTTP-PI 1110(HF2-HCI) EF K57/3. 8–11. Scolecodonts. 8. CICYTTP-PI 253(3) EF M28/0. 9. CICYTTP-PI 254(1) EF G58/3. 10. CICYTTP-PI 254(1) EF K50/3. 11. CICYTTP-PI 257(3) EF J56/0. 12. *Waltzispora polita*, CICYTTP-PI 255(1-ox) EF O46/4. 13. *Leiotriletes* sp. cf. *L. sphaerotriangularis*, CICYTTP-PI 255(1-ox) EF J36/0. 14. *Punctatisporites* sp. cf. *P. glaber*, CICYTTP-PI 258(2) EF R48/0. 15. *Botryococcus braunii*, CICYTTP-PI 255(1-ox) EF V64/2. Scale bars: 1–4, 6–9, 11 = 25  $\mu$ m; 5, 10, 15 = 50  $\mu$ m; 12–14 = 15  $\mu$ m.



**Plate 3.** Selected spore species from Unit 4, Middle Sappington Formation. 1. *Punctatisporites hannibalensis*, CICYTTP-PI 1102(HF2-HCI-1) EF O28/0. 2. *Retusotriletes crassus*, CICYTTP-PI 1102(Nd-HF2) EF N55/0. 3. *Emphanisporites orbicularis*, CICYTTP-PI 1103(HF2-HCI-1) EF M37/1. 4. *Pustulatisporites dolbii*, CICYTTP-PI 1102(HF2-HCI-1) EF Q37/1. 5. *Retusotriletes* sp. cf. *R. leptocentrum*, CICYTTP-PI 1103(HF2-HCI-Nd) EF M44/2. 6. *Retusotriletes incohatus*, CICYTTP-PI 1102(HF2-2) EF T46/0. 7. *Vallatisporites vallatus*, CICYTTP-PI 1105(HF2-2) EF O31/1. 8. *Vallatisporites spendens*, CICYTTP-PI 1102(HF2-1) EF G34/0. 9. *Vallatisporites drybrookensis*, CICYTTP-PI 1102(HF2-1) EF H37/0. 10. *Convolutispora major*, CICYTTP-PI 1106(HF2-1) EF K34/0. 11, 12. *Grandispora echinata*. 11. CICYTTP-PI 1102(HF2-2) EF K48/0. 12. CICYTTP-PI 1103(HF2-HCI-Nd) EF G34/3. 13. *Dictyotriletes flavus*, CICYTTP-PI 1102(HF2-Nd-1) EF R55/0. 14. *Dictyotriletes trivialis*, CICYTTP-PI 1102(HF2-Nd-1) EF P53/0. 15. *Verrucosiporites mesogrumosus*, CICYTTP-PI 1106(HF2-2) EF H29/3. Scale bars: 1, 2, 4, 6–9, 12, 14 = 10 µm; 3, 10 = 25 µm; 5, 11, 13, 15 = 15 µm.



**Plate 4.** Selected spore species from Unit 4, Middle Sappington Formation. 1, 2. *Retispora lepidophyta*. 1. CICYTTP-PI 1102(HF2-Nd) EF N55/3. 2. CICYTTP-PI 1109(HF2-2) EF R24/3. 3. *Endosporites micromanifestus*, CICYTTP-PI 1102(HF2-2) EF W44/3. 4. *Auroraspora macra*, CICYTTP-PI 1109(HF2-HCI-Nd) EF N44/2. 5. *Anapiculatisporites* sp. cf. *A. semicuspidatus*, CICYTTP-PI 1109(HF2-HCI-Nd) EF G43/2. 6. *Tumulispora rarituberculata*, CICYTTP-PI 1109(HF2-1) EF P30/4. 7. *Pustulatisporites* sp., CICYTTP-PI 1102(HF2-HCI-1) EF J48/0. 8. *Knoxisporites literatus*, CICYTTP-PI 1110(HF-1p) EF R40/1. 9. *Knoxisporites concentricus*, CICYTTP-PI 1102(HF-2) EF T44/4. 10. *Knoxisporites* sp. cf. *K. here-datus*, CICYTTP-PI 1109(HF2-HCI-1) EF M57/3. 11, 12. *Verrucosisorites nitidus*. 11. CICYTTP-PI 1102(HF2-1) EF K56/4. 12. CICYTTP-PI 1109(HF2-2) EF S47/0. 13. *Verrucosisorites papulosus*, CICYTTP-PI 1102(HF2-Nd) EF V51/2. 14, 16. *Convolutispora vermiformis*. 14. CICYTTP-PI 1109(HF2-HCI-1) EF K57/0. 16. CICYTTP-PI 1102(HF2-1) EF S57/0. 15. *Convolutispora oppressa*, CICYTTP-PI 1109(HF2-HCI-1) EF R57/0. Scale bars: 1–7, 10–16 = 10  $\mu$ m; 8, 9 = 15  $\mu$ m.



**Plate 5.** Selected spore species from Unit 4, Middle Sappington Formation. 1. *Kraeuselisporites explanatus*, CICYTTP-PI 1102(HF2-HCI-1) EF K24/0. 2. *Diaphanospora rugosa*, CICYTTP-PI 1102(HF2-2) EF N36/0. 3, 4. *Velamispurites perinatus*. 3. CICYTTP-PI 1102(HF2-1) EF M50/3. 4. CICYTTP-PI 1110(HF2-HCI-1) EF B34/0. 5. *Diaphanospora preplexa*, CICYTTP-PI 1110 (HF2-HCI-1) EF A34/0. 6. *Diaphanospora submirabilis*, CICYTTP-PI 1109(HF2-HCI-Nd) EF J37/0. 7. *Grandispora praecipua*, CICYTTP-PI 1110(HF2-HCI-1) EF T49/3. 8. *Endoculeospora setacea*, CICYTTP-PI 1109(HF2-1) EF T34/0. 9. *Grandispora saurota*, CICYTTP-PI 1109(HF2-HCI-Nd) EF C59/0. 10. *Spelaeotrilites* sp. cf. *S. pretiosus*, CICYTTP-PI 1109(HF2-HCI-Nd) EF Q36/1. 11. *Cristatisporites* sp. cf. *C. mattheusii/menendezii*, CICYTTP-PI 1109(HF2-1) EF U34/0. 12. *Bascaudaspora submarginata*, CICYTTP-PI 1109(HF2-HCI-Nd) EF R41/0. Scale bars: 1–5, 7, 8, 10 = 10 µm; 6, 9, 11, 12 = 15 µm.

*Veryhachium trispinosum* (Eisenack) Deunff 1954 'complex'. Remarks: Although we have kept the two species separated, some workers consider *Veryhachium downei* included in rather than synonymous with the *V. trispinosum* 'complex' (see Servais et al. 2007).

#### 4.2. *Las2 Assemblage – upper Trident Member*

The green shale of this unit yielded abundant amorphous organic matter, *Botryococcus braunii* Kützing 1849, and spheroidal forms with different sizes from small (50 µm) to large (150 µm). Among the latter (Plate 2, figures 1–6), we can differentiate the prasinophyte *Tasmanites* (*Pachysphaera*-type), characterised by thicker walled cysts with less frequent folds and radial pore canals (i.e. not smooth wall), from the other two morphotypes grouped under the term *Leiosphaeridia*: one exhibiting thin and smooth walls with frequent folds, and the other with ornamented vesicles possessing spines, cones or small processes (Guy-Ohlson 1988; Van Waveren 1992; Strother 1996). Pyritisation of their walls was observed (Plate 2, figure 5).

Leiosphaerids are considered a polyphyletic group related to autotrophic protists (prasinophyte *Halosphaera* type, Guy-Ohlson 1996), acritarchs (Strother 1996) or invertebrate eggs produced by arthropods such as pelagic copepods from freshwater Palaeozoic (Womack et al. 2012) to present marine environments (Van Waveren 1992). The latter affinity is also applied to large forms generally over 100 µm, variably folded and ornamented. Although the smooth vesicles cannot be restricted to the arthropod group, we attribute this affinity partially to large smooth forms and to the ornamented ones, especially found at this level and in some samples of the *Las3* (Figure 5).

An external mold of a ribbed clymenid ammonoid (Alan Titus, pers. comm.) was also found at this level (Figure 4B, CICYTTP-In 24).

#### 4.3. *Las3 Assemblage – Middle Sappington Formation, Unit 4*

A fairly well-preserved assemblage (*Las3*) from the Middle Sappington Unit 4 is recorded from seven productive samples. It is composed of diverse and abundant spores (*Auroraspora macra*, *Emphanisporites rotatus*, *Endosporites micromanifestus*, *Grandispora echinata*, *Knoxisporites concentricus*, *Lophozonotriletes* spp., *Punctatisporites hannibalensis*, *Pustulatisporites dolbii*, *Retispora lepidophyta*, *Retusotriletes crassus*, *Tumulispora rarituberculata*, *Vallatisporites drybrookensis*, *V. splendens*, *V. vallatus*, and *Verrucosisorites nitidus*) and less frequent acritarchs (*Gorgonisphaeridium absitum*, *G. ohioense*, *G. plerispinosum*, *Stellinium micropolygonale*) and prasinophytes (*Cymatiosphaera perimembrana*, *Dictyotidium fairfieldense*, *Maranhites* spp.). Among the spores, *Retispora lepidophyta* and species of *Punctatisporites* are abundant in all levels; a few species are registered at low frequency in almost all levels (*Cymatiosphaera* spp., *Gorgonisphaeridium absitum*, *Gorgonisphaeridium ohioense*, *Gorgonisphaeridium plerispinosum*, *Grandispora echinata*, *Punctatisporites hannibalensis*), while leiosphaerids (both smooth and ornamented morphotypes; see Section 4.2), are frequent in most levels and *Botryococcus braunii* is frequent in two levels (Figure 5; Plates 1–5).

#### SPORES

*Anapiculatisporites* sp. cf. *A. semicuspidatus* Playford & Melo 2012 (Plate 4, figure 5). Remarks: It differs from the original species in having sparser biform conii. SD: Tournaisian (Playford et al. 2012).

*Apiculatasporites* sp.

*Apiculiretusispora* sp. cf. *A. verrucosa* (Caro-Moniez) Streel in Becker et al. 1974. Remarks: *A. fructicosa* is considered a junior synonym (see Streel 2009). SD: Famennian–Tournaisian (Higgs 1975; Coleman & Clayton 1987; Higgs et al. 1988; McNestry 1988; Richardson & Ahmed 1988; Filipiak 2004; Ghavidel-syooki & Owens 2007; Marynowski & Filipiak 2007; Marynowski et al. 2010).

*Auroraspora macra* Sullivan 1968 (Plate 4, figure 4).

*Bascaudaspora submarginata* (Playford) Higgs et al. 1988 (Plate 5, figure 12). SD: Strunian–Mississippian (Higgs et al. 1988; McNestry 1988; Filipiak 2004; Azcuy & di Pasquo 2005, 2006; Marynowski & Filipiak 2007; Amenábar et al. 2009; Playford & Melo 2012).

*Claytonisporites rarisetosa* (Kedo) Playford & Melo 2012. Remarks: It is the senior synonym of *Dibolisporites abstrusus* (Playford) Playford 1976 in Playford & Melo 2012. SD: Strunian–Tournaisian (Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Avchimovitch 1993; Playford 1993; Playford & McGregor 1993; Marynowski & Filipiak 2007; Playford & Melo 2012).

*Convolutispora oppressa* Higgs 1975 (Plate 4, figure 15). SD: Strunian–Mississippian (Coleman & Clayton 1987; Higgs et al. 1988; McNestry 1988; Melo & Loboziak 2003; Azcuy & di Pasquo 2006; Playford & Melo 2012).

*Convolutispora vermiformis* Hughes & Playford 1961 (Plate 4, figures 14, 16). SD: Strunian–Mississippian (Playford 1964; Utting 1987a, 1987b; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; McNestry 1988; Avchimovitch 1993; Vigran et al. 1999; Melo & Loboziak 2003; Playford & Melo 2012).

*Convolutispora major* (Kedo) Turnau 1978 (Plate 3, figure 10). SD: Strunian–Tournaisian (Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Avchimovitch 1993; Turnau et al. 1994; Vigran et al. 1999; Melo & Loboziak 2003; Filipiak 2004; di Pasquo 2007; Marynowski & Filipiak 2007; Melo & Playford 2012; Playford et al. 2012).

*Cordylosporites* spp.

*Cristatisporites* sp. cf. *C. mattheusii* Higgs et al. 1988/*C. menendezii* (Menéndez & Azcuy) Playford 1978 (Plate 5, figure 11). SD: Strunian–Carboniferous (Coleman & Clayton 1987; Avchimovitch et al. 1988; Higgs et al. 1988; McNestry 1988; Amenábar et al. 2009).

*Cyclogranisporites* sp.

*Diaphanospora perplexa* Balme, Hassel emend. Evans 1970 (Plate 5, figure 5). SD: Famennian–Tournaisian (Playford 1976; McGregor & Playford 1992; Ghavidel-Syooki & Owens 2007).

*Diaphanospora rugosa* (Naumova) Byvscheva 1985 (Plate 5, figure 2). SD: late Famennian (Playford 1976; Avchimovitch et al. 1988; McNestry 1988; Melo & Loboziak, 2003; Melo & Playford 2012).

*Diaphanospora submirabilis* (Kedo) Byvscheva 1985 (Plate 5, figure 6). SD: early Tournaisian (Avchimovitch et al. 1988, 1993; Avchimovitch 1993).

*Dictyotrilites flavus* Keegan 1977 (Plate 3, figure 13). SD: Tournaisian (Higgs et al. 1988).

*Dictyotrilites trivialis* Naumova in litt., in Kedo 1963 (Plate 3, figure 14). SD: late Famennian–Tournaisian (Sandberg et al. 1972; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; McNestry 1988; Braman & Hills 1992; Amenábar et al. 2009; Melo & Playford 2012).

*Emphanisporites orbicularis* Turnau 1986 (Plate 3, figure 3). SD: Givetian–Famennian (Playford 1993; Playford & McGregor 1993).

*Emphanisporites rotatus* McGregor emend. McGregor 1973. A synonym of this taxon was illustrated as *Emphanisporites* cf. *E. robustus* by Sandberg et al. (1972, pl. 4, fig. 16).

*Endoculeospora setaceae* (Kedo) Avchimovitch, Higgs in Avchimovitch et al. 1988 (Plate 5, figure 8). SD: Strunian (Avchimovitch et al. 1988, 1993; Avchimovitch 1993; Filipiak 2004; Hartkopf-Fröder 2004; Marynowski et al. 2010, 2012).

*Endosporites micromanifestus* Hacquebard 1957 (Plate 4, figure 3). SD: late Famennian–Tournaisian (Playford 1976; Utting 1987a, 1987b; Higgs et al. 1988; McNestry 1988; McGregor & Playford 1992; Vigran et al. 1999; Melo & Playford 2012).

*Grandispora echinata* Hacquebard 1957 (Plate 3, figures 11, 12).

*Grandispora praecipua* Playford 1976 (Plate 5, figure 7). SD: Strunian.

*Grandispora saurota* (Higgs et al.) Playford & McGregor 1993 (Plate 5, figure 9). SD: Strunian–Tournaisian (Higgs et al. 1988; Playford 1993; Playford & McGregor 1993).

*Grandispora* spp.

*Knoxiosporites concentricus* (Byvsheva) Playford & McGregor 1993 (Plate 4, figure 9). SD: Strunian–mid Tournaisian (Coleman & Clayton 1987; Higgs et al. 1988; Playford 1993; Playford & McGregor 1993; Melo & Playford 2012).

*Knoxiosporites literatus* (Waltz) Playford 1964 (Plate 4, figure 8). SD: Strunian–Tournaisian (Playford 1976, 1993; Utting 1987a, 1987b; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; McNestry 1988; Richardson & Ahmed 1988; Braman & Hills 1992; McGregor & Playford 1992; Avchimovitch 1993; Playford & McGregor 1993; Melo & Loboziak 2003; Filipiak 2004; Hartkopf-Fröder 2004; Marynowski & Filipiak 2007; Melo & Playford 2012).

*Knoxiosporites* sp. cf. *K. heredatus* (Ischenko) Playford 1964 (Plate 4, figure 10). Remarks: it is the senior synonym of *K. pristinus* Sullivan 1968 (see Playford & Melo 2012). SD: late Famennian–Tournaisian (Sandberg et al. 1972; Playford 1976; Braman & Hills 1992; Avchimovitch et al. 1988, 1993; Avchimovitch 1993; Melo & Loboziak 2003; Filipiak 2004; Hartkopf-Fröder 2004; Marynowski & Filipiak 2007).

*Kraeuselisporites explanatus* (Luber) Azcuy & di Pasquo 2005 (Plate 5, figure 1). SD: Strunian–Tournaisian (Playford 1976; Coleman & Clayton 1987; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Richardson & Ahmed 1988; Avchimovitch 1993; McGregor & Playford 1992; Playford 1993; Playford & McGregor 1993; Vigran et al. 1999; Melo & Loboziak 2003; Filipiak 2004; Marynowski & Filipiak 2007; Melo & Playford 2012; Wicander & Playford 2013).

*Leiotrilites* cf. *struniensis* Moreau-Benoit 1979. SD: late Famennian (Melo & Loboziak 2003).

*Leiozonotrilites* sp. cf. *Spelaeotrilites crustatus* Higgs 1975

*Lophotrilites* sp.

*Lophozonotrilites* sp. cf. *triangulatus* (Ischenko) Hughes & Playford 1961. SD: Strunian–Tournaisian (Playford 1976; Coleman & Clayton 1987; Higgs et al. 1988; Vigran et al. 1999).

*Lophozonotrilites* sp. cf. *L. curvatus* Naumova 1953. SD: Givetian–Tournaisian (Playford & McGregor 1993).

*Punctatisporites hannibalensis* Wicander & Playford 2013 (Plate 3, figure 1). SD: Strunian–Lower Mississippian. Remarks: this species is markedly similar to *Punctatisporites resolutus* Playford 1971 from the Strunian–Mississippian.

*Punctatisporites* spp.

*Pustulatisporites dolbii* Higgs et al. 1988 (Plate 3, figure 4). SD: Strunian–Tournaisian (Avchimovitch et al. 1988; Higgs et al. 1988; Playford 1993; Playford & McGregor 1993; Filipiak 2004; Marynowski & Filipiak 2007; Amenábar et al. 2009).

*Pustulatisporites* sp. (Plate 4, figure 7).

*Retispora lepidophyta* (Kedo) Playford 1976 (Plate 4, figures 1, 2). SD: Strunian; cosmopolitan (Sandberg et al. 1972; Playford 1976, 1993; Coleman & Clayton 1987; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Richardson & Ahmed 1988; Avchimovitch 1993; McGregor & Playford 1992; Playford & McGregor 1993; Hartkopf-Fröder & Streel 1994; Vigran et al. 1999; Melo & Loboziak 2003; Filipiak 2004; Hartkopf-Fröder 2004; Marynowski & Filipiak 2007; Marynowski et al. 2010; Melo & Playford 2012; Wicander & Playford 2013).

*Retusotrilites crassus* Clayton et al. 1980 (Plate 3, figure 2). SD: Famennian–Mississippian (Coleman & Clayton 1987; Utting 1987a, 1987b; Higgs et al. 1988; Azcuy & di Pasquo 2006; Melo & Playford 2012).

*Retusotrilites incohatus* Sullivan 1964 (Plate 3, figure 6).

*Retusotrilites* sp. cf. *R. leptocentrum* Higgs 1975 (Plate 3, figure 5). SD: Strunian–Tournaisian (Playford & McGregor 1993).

*Rugospora flexuosa* (Juschko) Byvsheva 1985. Synonymy: *Hymenozonotrilites famenensis* Kedo in Sandberg et al. (1972, pl. 4, fig. 5). SD: Famennian–Tournaisian (McGregor 1979; Coleman & Clayton 1987; Avchimovitch et al. 1988; Higgs et al. 1988; Richardson & Ahmed 1988; Vigran et al. 1999; Melo & Loboziak 2003; Filipiak 2004; Ghavidel-Syooki & Owens 2007).

*Rugospora radiata* (Juschko) Byvsheva 1985. SD: Famennian; McGregor & Playford 1992; Avchimovitch 1993; Melo et al. 1999; Melo & Loboziak 2003; Marynowski et al. 2010, 2012; Racka et al. 2010).

*Spelaeotrilites crustatus* Higgs 1975. SD: Strunian–Mississippian (Higgs et al. 1988; Avchimovitch 1993; Playford 1993; Playford & McGregor 1993; Vigran et al. 1999).

*Spelaeotrilites* sp. cf. *S. pretiosus* (Playford) Neves & Belt 1970 (Plate 5, figure 10). SD: Strunian–Mississippian (Coleman & Clayton 1987; Higgs et al. 1988; Melo & Loboziak 2003; Filipiak 2004; Melo & Playford 2012).

*Tumulispora rarituberculata* (Luber) Potonié 1966 (Plate 4, figure 6). Remarks: this species is considered to be *T. rarituberculata* following the discussion by Playford & Melo (2012). We also consider that *T. variverrucata* (Playford) Staplin & Jansonius 1964 is likely the other taxon forming a morpho-group that shows variable morphology between the two extremes (see also Marynowski & Filipiak 2007). SD: Strunian–Tournaisian (Sandberg et al. 1972; McGregor 1979; Avchimovitch et al. 1988, 1993; McGregor & Playford 1992; Avchimovitch 1993; Playford 1993; Playford & McGregor 1993; Melo & Loboziak 2003; Filipiak 2004; Marynowski & Filipiak 2007; Melo & Playford 2012).

*Vallatisporites drybrookensis* morphon in Playford & McGregor 1993 (Plate 3, figure 9). SD: Strunian–Mississippian (Higgs et al. 1988; Playford & McGregor 1993; Melo & Playford 2012).

*Vallatisporites splendens* Staplin & Jansonius 1964 (Plate 3, figure 8). SD: Strunian–Mississippian (Playford & McGregor 1993; Melo & Loboziak 2003; Melo & Playford 2012).

*Vallatisporites vallatus* Hacquebard 1957 (Plate 3, figure 7). SD: Strunian–Carboniferous (Sandberg et al. 1972; Higgs et al. 1988; Avchimovitch 1993; Playford & McGregor 1993; Melo & Loboziak 2003; Filipiak 2004; Marynowski & Filipiak 2007; Melo & Playford 2012).

*Velamispores perinatus* (Hughes & Playford) Playford 1971 (Plate 5, figures 3, 4). SD: Strunian–Tournaisian (Sandberg et al. 1972; Playford 1976; McGregor 1979; Higgs et al. 1988; Vigran et al. 1999; Amenábar et al. 2009).

*Velamispores/Rugospora* spp.

*Verrucosiporites mesogrumosus* (Kedo) Byvscheva 1985 (Plate 3, figure 15). SD: Strunian–Mississippian (Avchimovitch et al. 1988, 1993; Avchimovitch 1993; Turnau et al. 1994; Melo & Playford 2012).

*Verrucosiporites nitidus* Playford 1963 (Plate 4, figures 11, 12). SD: Famennian–Tournaisian (Playford 1976, 1993; Coleman & Clayton 1987; Avchimovitch et al. 1988, 1993; Higgs et al. 1988; Richardson & Ahmed 1988; McGregor & Playford 1992; Avchimovitch 1993; Playford & McGregor 1993; Vigran et al. 1999; Melo & Loboziak 2003; Filipiak 2004; Azcuy & di Pasquo 2005, 2006; Marynowski & Filipiak 2007; Marynowski et al. 2012; Wicander & Playford 2013).

*Verrucosiporites papulosus* Hacquebard 1957 (Plate 4, figure 13). SD: Tournaisian (Hacquebard 1957; Playford 1964; Smith 1971; Utting 1987a, 1987b).

#### ACRITARCHS/PRASINOPHYTES

*Cymatiosphaera chelina* Wicander & Loeblich 1977 (Plate 1, figure 8). Comparison: the specimens are similar to that assigned to *Dictyotidium* sp. illustrated in Marynowski et al. (2012, fig. 4d), from the LN Zone in Poland. SD: Strunian (Figure 7). Frasnian–earliest Carboniferous (Wicander & Loeblich 1977; Filipiak 2005; Marynowski & Filipiak 2007; Marynowski et al. 2010; Wicander & Playford 2013).

*Cymatiosphaera* sp. (Plate 1, figure 9). Remarks: the specimens are similar to *Dictyotidium araiomegaronium* Hashemi & Playford 1998 but muri in the latter species are slightly wider and the range of diameter is smaller (15–23 µm) than in ours (35–45 µm).

*Cymatiosphaera* spp.

*Duvernaysphaera tenuicungulata* Staplin 1961

*Navifusa/Deusilites* sp. (Plate 2, figure 7).

*Dictyotidium fairfieldense* Playford 1976 (Plate 1, figure 15). Remarks: The specimen *Dictyotidium* sp?, illustrated by Marynowski & Filipiak (2007, fig. 8.0) in the Kowala Shale (LN Zone) in Poland, is here assigned to *D. fairfieldense*. SD: Strunian.

*Gorgonisphaeridium absitum* Wicander 1974

*Gorgonisphaeridium evexispinosum* Wicander 1974

*Gorgonisphaeridium ohioense* (Winslow) Wicander 1974

*Gorgonisphaeridium plerispinosum* Wicander 1974

*Gorgonisphaeridium winslowiae* Staplin et al. 1965 (Plate 1, figure 4). Synonymy: *Gorgonisphaeridium* sp. in Sandberg et al. (1972, pl. 3, figs 24–26). SD: Strunian (Playford 1993; Playford & McGregor 1993; Filipiak 2005; Wicander & Playford 2013).

Leiosphaerids–tasmanitids.

*Maranhites stockmansii* Martin emend. Martin 1985 (Plate 1, figure 7). SD: Famennian (Filipiak 2005; Marynowski & Filipiak 2007; González 2009).

*Maranhites mosesii* (Sommer) Brito emend. González 2009 (Plate 1, figure 12). Remarks: one specimen of *Maranhites britoi* Stockmans & Willièrè 1969, which is a junior synonym in González 2009, is also recognised. SD: Givetian–Famennian (di Pasquo et al. 2009; González 2009; Marynowski et al. 2010).

*Pterospermella* sp.

*Stellinium micropolygonale* (Stockmans & Willièrè) Playford 1977

*Verhyachium downei* Stockmans & Willièrè 1962

#### 4.4. Las4 Assemblage – the Cottonwood Canyon Member of the Lodgepole Formation

This assemblage (Sample CICYTTP-PI 255, Figure 4) is dominated by black particles and *Botryococcus braunii* (Plate 2, figure 15). Among the former, several forms looked like spores (e.g. *Waltzispore polita*, Plate 2, figure 12) but they were not recognisable until some became brown to light brown after oxidation (*Leiotriletes* sp. cf. *L. sphaerotriangularis*, Plate 2, figure 13, *Lophozonotriletes* sp.).

#### SPORES

*Leiotriletes* sp. cf. *L. sphaerotriangularis* (Loose) Potonié & Kremp 1954 (Plate 2, figure 13). SD: Carboniferous (Smith & Butterworth 1967)

*Lophozonotriletes* sp.

*Waltzispore polita* (Hoffmeister, Staplin & Malloy) Smith & Butterworth 1967 (Plate 2, figure 12). SD: early Viséan–Carboniferous (Smith & Butterworth 1967; Melo & Loboziak 2003; Azcuy & di Pasquo 2005; Melo & Playford 2012; di Pasquo & Iannuzzi 2014).

#### CHLOROPHYCEA

*Botryococcus braunii* Kützing 1849 (Plate 2, figure 15)

#### 4.5. Las5 Assemblage – Lodgepole Formation False Bakken (Pole Canyon)

Sample CICYTTP-PI 258 from the basal Lodgepole Formation yielded abundant black particles and *Botryococcus braunii*, and very few recognisable spores such as *Grandispora echinata* and *Punctatisporites* sp. cf. *P. glaber* (dark brown to black colour, TAI –4/+4, Plate 2, figure 14).

#### SPORES

*Grandispora echinata* Hacquebard 1957

*Punctatisporites* sp. cf. *P. glaber* (Naumova) Playford 1962 (Plate 2, figure 14)

#### CHLOROPHYCEA

*Botryococcus braunii* Kützing 1849

### 5. Discussion

#### 5.1. Ages of the assemblages and biostratigraphic correlations

Detailed biostratigraphic correlations in the CMT are incomplete, and many discrepancies in stratigraphic correlations are related to differences in ages given by different fossil groups (Figures 3, 7). Hence, association with global biotic events such as the DCB is difficult to assess at this time. Herein, five palynomorph assemblages (Las1 to Las5) from the Trident Member (Three Forks Formation), the middle (Unit 4) Sappington and

the basal Lodgepole (CCM and false Bakken) formations are defined by the SD of the palynomorphs (Figures 5–6) at Logan Gulch. The global stratigraphic ranges of selected species support the age of these assemblages (Figure 6).

The LAs1 assemblage in Montana is characterised by the co-occurrence of *Gorgonisphaeridium ohioense* and other acritarch/prasinophytes such as *Cymatiosphaera perimembrana*, *G. absitum*, *G. plerispinosum*, *Solisphaeridium astrum*, *Stellinium micropolygonale*, *Unellium lunatum* and *U. piriforme*, and the spores *Auroraspora macra*, *Grandispora echinata* and *Retusotriletes incohatus* (Figure 5). The stratigraphic position and ranges of selected species suggest a Famennian–pre-Strunian age, correlated to the *Gorgonisphaeridium ohioense* Zone Wicander (1975), from Ohio (USA) (Figure 6). The correlation of LAs1 to the *Grandispora gracilis*–*G. famenensis* (GF) Zone Streel et al. (1987) of Western Europe and *Cornispora varicornuta* Avchimovitch et al. (1993, in Filipiak 2004) for the Russian Platform, is also possible due to the presence of *G. echinata*, *A. macra* and *R. incohatus* (see also McGregor & Playford 1992) and the acritarchs *G. ohioense* and *Unellium* spp. (Filipiak 2005). These zones were correlated to the middle Famennian *marginifera/trachytera* conodont zones for the Trident Member of the Three Forks Formation (Figure 7; Sandberg et al. 1989).

An age cannot be given for the LAs2 assemblage based on palynologic content, although the presence of an indeterminate clymeniid ammonoid indicates a Late Devonian age (Alan Titus, pers. comm.). Korn & Titus (2006) studied one of the most diverse Famennian ammonoid assemblages in the Trident Member shale in sections near Logan Gulch. They suggest that the co-occurrence of the genera *Platyclymenia*, *Pleuroclymenia* and *Carinoclymenia* is good evidence for a position within the middle Famennian *Platyclymenia annulata* Zone, from which assemblages with morphologically similar species are particularly well known from Central and Eastern Europe and North Africa.

Racka et al. (2010) recorded clymeniid ammonoids from the *Annulata* Black Shale at Kowala in the Holly Cross Mountains (Poland), and recovered from the same interval miospores and acritarchs of the *Diducites versabilis*–*Grandispora famenensis* (VF) Zone (Avchimovitch et al. 1993) and conodonts of the *trachytera* Zone (Figure 7).

The underlying LAs1 assemblage shares several palynomorphs (*Gorgonisphaeridium ohioense*, *Ammonidium loriferum*, *Unellium* spp., *Michrystidium* spp., *Veryhachium* spp., *Auroraspora macra*, *Grandispora echinata*) with this locality, and the similarity to *Platyclymenia* sp., as illustrated by Racka et al. (2010; fig. 13.1–4), to our external mold (Figure 4B) supports the attribution of LAs1 and LAs2 to the *Annulata* Event (Figure 7).

Even though the upper Unit 1 (shale) of the Lower Sappington Formation was not palynologically productive at Logan Gulch, it was dated to the Lower *expansa* Zone (ex Upper *Polygnathus styriacus* zone) by Sandberg et al. (1972) as shown in Figure 3. Hence, a gap during the *postera* Zone occurs at the top of the Trident Member (Figure 7).

The LAs3 assemblage (Unit 4) comprises 55 spore species and 17 phytoplankton species (Figure 5). The co-occurrence of the diagnostic spores *R. lepidophyta* and *V. nitidus* allowed the assignment of a latest Famennian (Strunian) LN Zone age (Figures 6, 7). Strunian species present in this assemblage that

reinforce this age are *Dictyotidium fairfieldense*, *Gorgonisphaeridium winslowiae*, *Endoculeospora setacea* and *Grandispora praecipua*. Other Strunian–Early Mississippian species (e.g. *Punctatisporites hannibalensis*, *Pustulatisporites dolbii*, *Retusotriletes crassus*, *Tumulispora rarituberculata*, *Vallatisporites drybrookensis*, *V. splendens* and *V. vallatus*), or those whose last occurrence (LO) is at the top of the Strunian, include *Rugospora radiata*, *Diaphanospora rugosa*, *Gorgonisphaeridium absitum*, *G. ohioense* and *G. plerispinosum*. Rare echinoderms in Unit 4 channels have been compared to forms found in the Louisiana Limestone with *Gnathodus kockeli* (Sprinkle & Gutschick 1967).

The most current DCB biozonations are derived mostly from intensive biostratigraphic studies of continuous settings in Western (see Higgs et al. 1988; Streel 2009) and Eastern Europe (see Kaiser et al. 2015). The transition from the later Famennian to the Carboniferous is defined by three miospore Interval Zones (LL, LE, LN), and the early Tournaisian *Vallatisporites vallatus*–*Retusotriletes incohatus* (VI) Zone which covers six conodont zones (from Upper *expansa* to *duplicata/sulcata*).

The first occurrence of *Retispora lepidophyta* is considered to be the basal Strunian Stage, which correlates to the Upper *expansa* conodont Zone. The appearance of *Verrucosisporites nitidus* defines the base of the LN Zone and coincides with the main extinction Hangenberg Event in Germany (Higgs & Streel 1994; Streel 2009). *Vallatisporites vallatus* first appears in Poland in the upper LN Zone (Marynowski & Filipiak 2007).

In Russia, *Tumulispora malevkensis* is used to separate an uppermost Devonian *V. pusillites*–*T. malevkensis* (PM) Subzone (Avchimovitch 1993; Avchimovitch et al. 1993). However, in Brazil, the Rle and LVa zones are recognised between two gaps (Figure 7) due to the presence of diamictite units at the DCB (e.g. Loboziak et al. 1993; Melo & Loboziak 2003). The LVa Zone is defined on the first appearance of *Vallatisporites vallatus* (Melo & Loboziak 2003). Streel et al. (2013) and Streel (2015) emphasised that the combined Rle and LVa zones may represent lateral facies variants of the LE/LN zones time interval, which has implications for the dating of glacial pulses in South America.

In Montana, Sandberg et al. (1972) defined the *Siphonodella praesulcata* Zone in units 2, 3 and 5 of the Sappington Formation at (Peak 9559) near Hardscrabble and Sacajawea peaks in the Bridger Mountains, the same location where they described a palynomorph assemblage with *R. lepidophyta* in the upper part of Unit 4 (Figures 1, 3). This assemblage is composed of 34 spore species (18 illustrated spore species and 16 unillustrated) and three acritarch species (one illustrated and two not illustrated; see Appendix 1).

Several species are shared with our LAs3 assemblage at Logan Gulch (*Dictyotriletes trivialis*, *Emphanisporites rotatus*, *Grandispora echinata*, *Knoxisporites heredatus*, *Retusotriletes incohatus*, *Tumulispora rarituberculata*, *Velamisporites perinatus* and *Vallatisporites vallatus*), although *Verrucosisporites nitidus* was not recorded by Sandberg et al. (1972). In agreement with Streel et al. (2013), the latter species shows intra- and interspecific variations that prevent a positive identification (Turnau et al. 1994). Furthermore, it is rare in the LN Zone of LAs3 (Figures 5–7), indicating that it could be facies sensitive (Becker et al. 2016).

Most of the LAs3 species were also recognised and illustrated in correlative assemblages of the LN Zone (see list of

species above, and Figure 7) in the Bedford Shale in Kentucky (Coleman & Clayton 1987), the Saverton Shale in Pike County, Illinois, (Wicander & Playford 2013), the Bakken Formation sub-surface, and equivalent units in Canada (McGregor & Playford 1992; Playford & McGregor 1993), as well as north-east Greenland (Vigran et al. 1999). These occurrences reinforce the age and correlation of Unit 4 of the Sappington Formation (Figure 7).

In Poland, similar DCB assemblages were documented by Filipiak (2004, 2005) from a trench and three boreholes in the Holy Cross Mountains (HCM). In general, acritarch assemblages I (Famennian) and II (Tournaisian) are similar in composition to our LAs1 (Trident Member) and LAs3 (Sappington Unit 4) assemblages; some taxonomic differences may reflect palaeo-environmental changes. Dominance of *Gorgonisphaeridium* spp. is observed in the HCM Famennian assemblage whereas *Michrystidium* and *Unellium* predominate in the Tournaisian assemblage.

Concerning the spores, Filipiak (2004), Filipiak & Racki (2010) and Marynowski et al. (2012) recorded from the HCM miospore assemblages of the *Retispora lepidophyta*–*Verrucosporites nitidus* (LN) Zone. Several additional species are illustrated in those works (e.g. *Bascaudaspora submarginata*, *Grandispora echinata*, *Knoxisporites literatus*, *Kraeuselisporites explanatus*, *Pustulatisporites dolbii*, *Rugospora flexuosa*, *Tumulispora rarituberculata*, *Vallatisporites* spp.) and support their correlations with our LAs3 correlation to the last black shale of the Hangenberg Event (Figure 7).

Intermediate morphological stages of of *Vallatisporites* morphotypes in the LAs3 assemblage were shown in the LN Zone by Playford & McGregor (1993) and Filipiak & Racki (2010). The latter authors interpreted the presence of abnormal morphologies of some species in the LN Zone associated with volcanic ash interbeds, as well as charcoal debris and polycyclic aromatic biomarkers, as possibly reflecting the mutagenic effect of regional acidification due to explosive volcanism and forest wildfire.

In south-west Spain, González et al. (2005a, 2005b) described Upper Devonian palynostratigraphic successions in the LN Zone of the Iberian Pyrite Belt, and documented abundant spores and organic-walled phytoplankton. Species in common with our LAs3 assemblage (e.g. *Grandispora echinata*, *Knoxisporites literatus*, *Kraeuselisporites explanatus*, *Pustulatisporites dolbii*, *Retispora lepidophyta*, *Retusotriletes incohatus*, *R. crassus*, *Rugospora flexuosa*, *Tumulispora rarituberculata*, *Vallatisporites* spp. and *Verrucosporites nitidus*) support their correlation (Figure 7).

Their assemblage also includes the presence of the prasino-phyte *Maranhites stockmansii* and exceptionally well-preserved, morphologically variable *Maranhites* species, which are widely distributed, long-ranging and recurrent taxa that are confined to the Southern Hemisphere during the mid–Late Devonian (Wood 1984). In the Late Famennian, their lower abundance persisted into the Mississippian (González 2009).

In agreement with Wood (1984) and González (2009), *M. stockmansii* has been found mainly from the Upper Devonian of Spain, Belgium, Turkey, Brazil and Poland (up to the lower Tournaisian, after Filipiak 2005). Tournaisian occurrences of *M. mosesii* have been considered to be genuinely *in situ* in Brazil

(Quadros 1982), Algeria (Lanzoni & Magloire 1969; Attar et al. 1980) and North America (Wood 1984).

Hence, optimal conditions for the proliferation of such species with other faunal and floral events occurred during the Frasnian and Famennian in southern Gondwana. A progressive narrowing of the ocean separating Laurussia and Gondwana favoured their migration northwards up to the southern margin of Laurussia (see Wood 1984; González 2009; Scotese 2014). The presence of *Maranhites stockmansii* and *M. mosesii* species in our LAs3 assemblage (Figure 5), even with their infrequent occurrences, supports this interpretation. The acritarch *Umbellaspheeridium saharicum*, which occurs frequently in the Southern Hemisphere and eastern North America with *Retispora lepidophyta* assemblages bearing abundant *Maranhites* species, is not present in this region, confirming its endemism (Wood 1984; Vavrdová & Isaacson 1999).

The final and global extinction of the plant producing *R. lepidophyta* in the upper part of the *kockeli* Zone defines the base of the VI Zone, between the Lower and Upper Stockum Limestones of the type area (Higgs & Streel 1984, 1994; Higgs et al. 1993; Kaiser et al. 2015; Streel 2015). The appearance of *Spelaotriletes balteatus* or *Bascaudaspora mischkinensis* is used to define eastern European basal Carboniferous *Tumulispora malevkensis*–*Spelaotriletes balteatus* (MB) and PMi subzones (Avchimovitch 1993; Byvsheva & Umnova 1993). The Lower Alum Shale Event coincides with the base of the *Kraeuselisporites hibernicus*–*Umbonatisporites distinctus* (HD) Zone (Figure 7), defined by the entries of *Kraeuselisporites hibernicus* and *Umbonatisporites distinctus* (Higgs & Streel 1984).

The LAs4 assemblage in the Cottonwood Canyon Member (CCM) of the Lodgepole Formation (Figures 2–7) is no older than Kinderhookian (mid–late Tournaisian), based on biostratigraphically useful species such as *Waltzispora polita* and *Leiotriletes* sp. cf. *L. sphaerotriangulus*. Their occurrence correlates to the BP–PC Zones of Higgs et al. (1988) that correlate to the conodont *crenulata-isosticha* Zones (Figure 7). The first appearance of *Waltzispora polita* is earliest Visean (cf. di Pasquo & Iannuzzi 2014), so the age of the LAs4 needs further study. The LAs5 assemblage from the False Bakken shales above the Scallion interval of the Lodgepole Formation (Figures 3–7) yields very few recognisable spores. *Grandispora echinata* and *Punctatisporites* sp. cf. *P. glaber*, found in lower assemblages (Figure 5), are not age-diagnostic. Therefore, the ages given to the Lodgepole Formation (CCM) and Unit 4 of the Sappington Formation indicate a relative placement of the DCB at the top of latter (di Pasquo et al. 2012; Warren et al. 2014), or somewhere within Unit 5 (Isaacson et al. 2014).

Depending on published interpretations and comparison to the Middle Bakken Formation (Gaswirth & Marra 2015), most or all of the silty Sappington Formation is generally assigned to the *praesulcata* Zone, although MacQueen & Sandberg (1970) originally placed the DCB as suggested above. Thus, the CCM where a concentrated, phosphatic lag deposit bearing conodonts of Tournaisian age occurs (see Sandberg 1979) is time-correlative with the Upper Bakken Formation (Williston Basin), and lower Banff Formation in Canada (Figure 1), thus representing a change to an Early Mississippian transgression following a major hiatus especially in western Montana (Figures 2, 3, 7).

Cratonic siltstones (i.e. uppermost Devonian Sappington Formation units 2–5) were thought to be time-equivalent to the

more offshore lower black shales of the Exshaw Formation, which are accompanied by latest Devonian ash beds (cf. Savoy 1992; Smith & Bustin 2000; Richards et al. 2002). The overlying silty member of the Exshaw is thought to be Early Mississippian based on combined conodont data and chronostratigraphic correlations (Johnston et al. 2010; Hartel et al. 2014). This interesting relationship suggests lateral time transgression of facies, something that has been difficult to demonstrate in the Sappington and Bakken formations but could possibly be resolved by further analysis of Unit 5 of the Sappington Formation and the Middle Bakken, respectively.

Middle Sappington units 3 and 5 are (so far as is presently known) barren of palynomorphs, although shale intervals are known. These units are dated by conodonts found at the margins of the CMT (Sandberg et al. 1972) (Figure 3). Current diagnoses of conodont species *Siphonodella praesulcata* and *S. sulcata* are not clear, and they overlap. As demonstrated by Kaiser & Corradini (2011), only a few specimens fit in a single definition, whereas the great majority of specimens show intermediate features between them. Therefore, several morphotypes were established between the two taxa *S. praesulcata* and *S. sulcata*, with an evolutionary trend towards increasing abundance of the curved morphotypes in younger populations. For this reason, it is necessary to find 'praesulcata-like' specimens associated with Devonian faunas and microfloras of Strunian age globally, as well as the first occurrence of 'sulcata-like' morphotypes to precisely delineate the DCB in Montana. Continuing studies of DCB units in Montana will help to locate it more accurately (di Pasquo et al. 2016; Rice et al. 2016a).

### 5.2. Palynofacies, palaeoecology and palaeoenvironmental assessment of the Trident Member

The palynological results at the major group level (Figure 5, inset) help to interpret changes in sea level and/or conditions of the depocentre during deposition of the late Famennian-age Trident Member at Logan Gulch. Palaeoecological affinities of marine phytoplankton and other species recorded in the LAs1–LAs2 were discussed by Streel (1999) based on the study of the late Famennian of the Ardenne–Rhine region, Europe. He established that the sphaeromorph abundance corresponds to a (back barrier) restricted marine environment; the abundance of the thick-spined acritarch *Gorgonisphaeridium* corresponds to intermediate marine conditions; and more diversified acritarchs characterise more open marine environments.

CICYTTP-PI 253 (Figure 5) reveals a dominance of acritarchs, especially *Gorgonisphaeridium* species, whereas leiosphaerids are a minor component of the assemblage. In the overlying sample, CICYTTP-PI 254, the abundance of leiosphaerids (and other sphaeromorphs) increases slightly. Normal marine conditions prevailed throughout this interval up to another increase in leiosphaerids which is found in CICYTTP-PI 257. This indicates a relative change to more restricted marine conditions.

A high abundance of leiosphaerids–tasmanitids, other forms attributed to arthropod eggs (see Section 4.2), *Botryococcus*, rare spores, and the presence of pyrite within them, and on their walls, as well as the absence of acritarchs characterises the LAs2 assemblage (Figure 5). The presence of clymenid

ammonoids (Figure 4B) and pyritised organic matter in the same shale is a good indicator of reducing and anoxic conditions in a low-energy (offshore) marine environment. This assemblage is likely related to the *Annulata* Event (ribbed ammonoid *Platyclymenia annulata*) (Figure 7), which is considered a record of global anoxia, paired with an interglacial sea-level rise (see Racka et al. 2010). However, we suggest that specialised species (e.g. leiosphaerids, copepods, ammonoids; see Korn 2004) would colonise this type of environment via opportunistic blooming. Such a scenario has been observed in the rhythmic deep-shelf successions of the HCM geochemically and palaeontologically characterised by Racka et al. (2010). All ammonoid taxa found in the Bozeman area (Figure 1) display significant irregularities in the spacing of the septa, regardless of their systematic assignment (Korn & Titus 2006). Septal crowding occurs not only as a phenomenon of terminal growth, where it is particularly well known, but also in intermediate stages, followed by phases with wider septal interspaces. Such patterns indicate stressed conditions during the lifetime of the ammonoids in the CMT (Korn & Titus 2006).

The Trident Member shale is interbedded with fossiliferous limestone nodules and thicker packstone beds that represent a change to oxic conditions in a shallow marine environment. Erosion between the upper Trident Limestone cap and Unit 1 of the Sappington Formation represents the Lower to Upper *post-era* zones (Figures 2, 3, 7). This hiatus is thought to be present throughout North America (Sandberg et al. 1989). Kaiser et al. (2015) related shales we think are equivalent to Sappington Unit 1A/B to global oceanic anoxia associated with the Dasberg biotic event. This event took place during eustatic transgression and was not caused by a glacial advance with regression (Hartenfels & Becker 2009).

### 5.3. DCB stratigraphic correlation, extinction events and glaciation proxies

Events of considerable interest in past climates include glaciations and their effects on extinction events linked to sedimentary successions. Not only did the Late Devonian glacial advances and retreats periodically lower global sea levels, but these events were accompanied by basin starvation, erosion and subaerial exposure, siliciclastic progradation, and anoxia in shallow epicontinental seas (Isaacson et al. 2008). The Late Devonian glaciation is a relatively new geologic paradigm that has not been widely accepted even after the well-documented discovery of glacial pavements, dropstones and diamictites across three basins in Brazil (Caputo 1985; Isaacson et al. 1999, 2008; Caputo et al. 2008).

Palaeoclimate shifts from warm and dry to cold and wet between Frasnian and Famennian time (Cecil et al. 2004; Elrick et al. 2009; Joachimski et al. 2009) on low-latitude sequences coeval with the glaciation(s) are evidenced by hiatuses/lacunae, black shales with elevated Total Organic Carbon (TOC) values, evaporites, phreatic zone breccias, and siliciclastic deposition (Isaacson et al. 1999, 2008, and references therein). The global, first-order highstand of sea level occurred in the latest Frasnian when climates were generally warm and equable, but during second-order Famennian regression, there was alpine and continental ice in the South Polar region.

The onset of a global cooling trend, near the end of the Devonian, marked the transition from greenhouse to icehouse conditions. End-Devonian global oceanic anoxic events (OAE) occurred at this time, with a major extinction (Hangenberg Event) punctuated by the disappearance of typical Devonian groups (e.g. stromatoporoids, trilobites, fish and palmatolepid conodonts) and the subsequent radiation of new groups in the basal Carboniferous (e.g. *Gattendorfia* ammonoids; see Buggisch 1991; Joachimski & Buggisch 1993; McGhee 2013).

Therefore, the transition from the Devonian to the Carboniferous witnessed a profound change in the biosphere and related sea-level changes. A minor eustatic sea-level fall (a lower biostratigraphic crisis interval) occurred in the uppermost Famennian (Kaiser et al. 2015). This interval was followed rather abruptly by pantropically widespread black shale deposition (HBS and equivalents). The transgressive and hypoxic/anoxic phase was related to a global carbonate crisis and perturbation of the global carbon cycle as demonstrated by a distinctive positive carbon isotope excursion, probably as a consequence of climate/salinity-driven oceanic overturns and outer-shelf eutrophication (Kaiser et al. 2015). Globally, it is the main extinction level for marine biota (e.g. ammonoids, trilobites, conodonts, stromatoporoids, corals, some sharks, deepwater ostracodes, placoderms, chitinozoans and acritarchs) and terrestrial early tetrapods (Caplan & Bustin 1999; McGhee 2013).

Extinction patterns were similar in widely separate basins of the western and eastern Prototethys, whereas a contemporaneous marine macrofauna record from high latitudes is missing altogether. Proposed causal factors (e.g. bolide impacts, tectonism, oceanic overturn, glacio/eustatic fluctuations, volcanism) that triggered this severe biotic crisis spanned several hundred thousand years for the complete event interval (e.g. Caputo 1985; Becker 1996; Caplan & Bustin 1999; Isaacson et al. 1999, 2008; Streel 1999; Streel et al. 2000; Kaiser et al. 2009, 2011, 2015; Becker et al. 2016).

Parallel to the phytoplankton extinction, there was expansion of a dense terrestrial vegetation during the Late Devonian, with the development of the first leafy forests. These would have caused the progressive retention of carbon, nitrogen and phosphorus in terrestrial environments. These elements constitute important nutrients in marine ecosystems, and their decrease would have prevented the good development of the phytoplankton and consequently generated a great planktonic crisis including for chitinozoans (Tappan 1986; Mullins & Servais 2008). Riegel (2008) discussed several causes of the 'Phytoplankton Blackout', which is related to the DCB crisis. Recovery of this group was very slow during the Late Palaeozoic until the Triassic, when the appearance of dinoflagellates and, somewhat later, of coccolithophorids and diatoms occurred. Schwark & Empt (2006) analysed sterane biomarkers to study algal evolution and extinction events. They found that the gradual radiation of algae may have been triggered by frequent mass extinctions at the latest Devonian, when a massive decline of acritarchs occurred.

The coeval rise in the  $C_{28}/C_{29}$ -sterane ratio occurring at the DCB indicates a fundamental change in algal composition comparable to those that followed the K/T boundary with the appearance of haptophyte algae or diatoms. During the Carboniferous the common Devonian phytoplankton cysts or phycocoma would have declined, and the appearance of a non-

encysting algal group (Strother 1996; Mullins & Servais 2008) coincided with the global augmentation of numerous filamentous Siphonales and the rise of Dasycladales. So, a time with low planktonic diversity and scarce primary productivity, called a 'phytoplanktonic blackout', occurred globally, in which only a few opportunistic acritarchs survived. Most dominant species of the acritarch genera *Veryhachium* and *Micrhystridium* were recorded in the Carboniferous and Permian along with various prasinophytes (e.g. *Hemiruptia*, *Leiosphaeridia*, *Tasmanites*, *Cymatiosphaera*, *Dictyotidium*) and a few chlorophytes (e.g. *Botryococcus*, *Tetraporina*, see Mullins & Servais 2008).

In this context, the DCB in Montana falls within a thin stratigraphic succession of Sappington Formation black shale, silty limestone and sandstone that includes combined erosional and transgressive surfaces, and surfaces that are not resolved. Following the Dasberg Event (Unit 1A/B; Figure 7) a glacial advance at the base of the *Retispora lepidophyta*-*Knoxisporites literatus* (LL) zone is the beginning of the Hangenberg Event (Figure 2). Although Unit 1 shales (A and B) do not occur at Logan Gulch, Unit 1C and 1D, Unit 2 and Unit 3 represent: (i) hiatus (SB between Unit 1B and 1C); (ii) transgression (Unit 1C and Unit 2); and (iii) slow, sustained base-level drop or filling of accommodation space (Unit 3, i.e. aggradation or a highstand system tract; cf. Phelps 2015).

Thus, part of the Sappington succession correlates to the Hangenberg Event and glacial advance. Based on sedimentology and fossiliferous content, Unit 2 is a transgressive unit/event documented by widely deposited brachiopod, sponge-oncoidal limestone beds that represents photic zone productivity followed by lower distal, calcareous shoreface conditions (gradational contact with Unit 3). Note that most authors (see Angulo & Buatois 2012; Cole et al. 2015) working with core at the same stratigraphic interval in the Bakken Formation suggest that these silty basal Middle Member limestones, like the basal Sappington Siltstone Member limestones (Unit 2), are associated with regression (Figures 2, 7).

Unit 3 is abruptly overlain by Unit 4, yielding a diverse assemblage with abundant spores and less frequent or variable proportions of acritarchs/prasinophytes, with an abundant intermixing of land-derived elements (spores, *Botryococcus*) and other sphaeromorphs (Figure 5). Variable pyritisation was also recognised, depending on the level. The spore assemblage points to the LN Zone that was associated with a glacial advance in a regressive phase (Walliser 1996; Caputo et al. 2008). This interval in the CMT was deposited in a restricted bay in a shallow marine (tidal?) environment that was freshwater-influenced under more humid climatic conditions. The presence of pyrite in the walls of palynomorphs in some levels of Unit 4 parallels the record of pyrite ooids in Devonian black shales of the Bedford Shale of the western Appalachian basin (Schieber & Riciputi 2004), interpreted as a consequence of intermittent sea-level drop in shallow-water conditions. Our interpretation neglects some proposals of sequence-stratigraphic interpretations of the Sappington Formation, which rely on Unit 4 to be an offshore marine deposit (see discussion).

Marine influence in Unit 4 is recognised at Antelope Creek by the presence of an inlet channel containing marine macrofossils including *in situ* blastoid life assemblages (Sprinkle & Gutschick

1967). Freshwater conchostracans (similar to those found in Unit 1C below) and marine brachiopods are found in this unique tidal channel of the otherwise relatively unfossiliferous Unit 4. Apparently these *kocklei* Zone-age echinoderms were mixed together with small organisms, possibly in a muddy, *Bifungites*-dominated brackish environment (Sprinkle & Gutschick 1967), where a *Retispora lepidophyta* assemblage was preserved (Rice et al. 2016b).

Although Unit 5 occurs broadly across Montana, it is not clear whether it is early and/or late *praesulcata* Zone or even Mississippian. It thins to very high-energy facies near established palaeohighs to the east, and facies-stacking, as seen at Logan Gulch, is different in sections farther west. The gap between those units is also supported by our palynological study herein showing that the CCM is not older than Tournaisian in age (Figure 7). This unit likely represents an offshore shelf environment, as evidenced by a basal transgressive surface, few poorly preserved miospores, black shale with conodonts and bioturbated siltstone with scarce bioclasts. It is correlatable to the basal Canadian Banff Formation (see Hartel et al. 2014), more so than to the upper Bakken of the Williston Basin. This is in agreement with the presence of *S. crenulata*–*isosticha* zones in the Lodgepole Formation (Sandberg & Klapper 1967; MacQueen & Sandberg 1970; Sandberg et al. 1972; Sandberg 1979).

Using geochemistry, Cole et al. (2015) found that the Devonian–Lower Mississippian boundary interval is missing in Montana and documented the largest positive  $\delta^{13}\text{C}_{\text{carb}}$  isotope excursion associated with strata of the Lower Mississippian *crenulata* Zone. This was interpreted as the result of a draw-down of atmospheric  $\text{CO}_2$ , and Tournaisian glaciation.

In the lowermost Mississippian, a series of at least three glaciations (events) occurred prior to transgression and deposition of early Tournaisian HD Zone black shales (Lower Alum Shale) and may correlate with the *sandbergi/crenulata* event (Figure 7). Most of these events are likely unrecorded in Montana, and are probably masked by tectonic effects, although Unit 5 may be involved, pending its regional ages.

The lower Tournaisian hiatus (Figure 7) appears to coincide with coeval glacial recycling events in the Bolivian subsurface and outcrop (Isaacson & Díaz Martínez 1995; di Pasquo & Azcuy 1997; di Pasquo 2007; di Pasquo et al. 2015) and Brazil (Playford et al. 2012).

Kidder & Worsley (2010) proposed that an important precondition for icehouse development is major continental collisional orogeny. The silicate weathering that characterises such major mountain-building events stimulates global cooling. End-Devonian tectonic inversions are well known, especially around the equatorial belt (Clayton et al. 2002; Racki et al. 2002). However, Devonian glaciations likely developed and waned rapidly, relative to the slow process of orogenesis. Rapid glaciation onset was generated by oceans reaching (nutrient?) threshold conditions, although other factors remain complex (Kidder & Worsley 2010). Consequently, evidence for the effects of the glacioeustatic sea level changes associated with the Hangenberg events are partly documented in DCB Unit 4 and Unit 5 proxy beds of the Sappington Formation at Logan Gulch in Montana. Further regional study of integrated conodonts and palynomorphs will combine to help delineate ages of Sappington units and

constrain extinction events as low-latitude proxies along the DCB in western Montana.

## 6. Conclusions

The Devonian–Carboniferous upper Three Forks, Sappington and lower Lodgepole formations at the Logan Gulch Devonian reference section near Three Forks in western Montana (Figure 1) were deposited in the cratonic Central Montana Trough (CMT). Thirteen of 29 samples processed for palynomorphs yielded 87 species of spores and phytoplankton, and some scolecodonts. Of the entire palynoflora, 27 species are prasinophytes and acritarchs (marine origin), 59 species are trilete spores, and there is one chlorophycean – *Botryococcus braunii* (fresh/brackish origin). Seven other samples from the lower Lodgepole Formation yielded abundant phytoclasts and/or amorphous organic matter without palynomorphs (di Pasquo et al. 2012; Warren et al. 2014; Warren 2015).

Five assemblages were recognised based on the SD of species (Figures 4–6) recovered from the Trident Member of the upper Three Forks Formation (LAs1 and LAs2 assemblages), Unit 4 of the Middle Sappington Formation (LAs3 assemblage), and the Cottonwood Canyon Member (LAs4 assemblage) and False Bakken (LAs5 assemblage) of the Lodgepole Formation. Their ages (Figure 7) and palaeoenvironmental interpretations are supported by diagnostic and complementary species (Figures 5, 6).

Green shales of the Trident Member (17 m) at Logan Gulch yielded a low-diversity palynomorph assemblage (LAs1), with variable preservations of mostly cosmopolitan, long-ranging phytoplankton species (Frasnian to Strunian) such as *Gorgonisphaeridium ohioense*, *G. absitum*, *G. plerispinosum*, *Unellium piri-forme*, some scolecodonts and few spores (*Auroraspora macra*, *Grandispora echinata* and *Retusotriletes incohatus*). A likely middle Famennian–pre-Strunian age is supported and correlated to the VCo Zone of western Europe and its equivalents (Figure 7). Palaeoecological affinities of marine species of *Gorgonisphaeridium* and other acritarchs (pyritised) characterise open marine conditions.

The fossiliferous limestone nodules and thicker packstone beds of the upper Trident Member represent a change to oxic conditions in a shallow marine environment. LAs2 assemblage taxa obtained from an intercalated green shale are dominated by leiosphaerid–tasmanitids and copepod eggs, less frequent *Botryococcus* and scarce spores. The low diversity of pyritised palynomorphs and a clymenid ammonoid suggest anoxic conditions in a low-energy shelf marine environment. Ammonoid taxa found in the Bozeman area (Figure 1) display significant irregularities in the spacing of the septa, which indicate stressed conditions during their lifetime. Such a stressed environment would lead to opportunistic blooming of specialised species (e.g. leiosphaerids, copepods). The change from normal marine to restricted conditions is possibly related to the *Annulata* Event, which is considered a record of global anoxia, paired with glacioeustatic variations in sea level (Figure 7).

At Logan Gulch, prospective shales of basal Unit 1 of the Lower Sappington Formation are anomalously thin (<30 cm) and palynologically barren. Erosion between the upper Trident Limestone cap and Unit 1 of the Sappington Formation

represents the Lower to Upper *postera* zones (Figures 2, 3, 7). This hiatus is thought to be present throughout North America (Sandberg et al. 1989).

The uniquely burrowed but unfossiliferous Unit 4 shale occurs within the 'Middle Sappington Siltstone' (Figure 2), where it overlies distal Unit 3 shoreface deposits, and underlies shoreface siltstone and sandstone of Unit 5. Unit 4 yielded a *Retispora lepidophyta* assemblage. Age-diagnostic taxa (LAs3 assemblage) include spores (e.g. *Grandispora echinata*, *G. praecipua*, *Endoculeospora setacea*, *Retispora lepidophyta*, *Verucosporites nitidus* and *Vallatisporites* spp.) and less frequent phytoplankton species (e.g. *Dictyotidium fairfieldense*, *Gorgonisphaeridium winslowiae*), indicating the Strunian LN zone.

The LAs3 assemblage shares many species with Unit 4 at Hardscrabble in Montana, and the Middle Bakken Member in southern Saskatchewan (Figure 1). This supports their correlation with several latest Famennian assemblages in USA, Greenland, Brazil, western and eastern Europe, and Australia (Figure 7). Although units 4 and 5 and associated surfaces have been interpreted differently, some parts of these lithotopes are likely correlatives to the Hangenberg Black Shales.

Palaeoenvironmental and palynofacies data in Unit 4 support a brackish shallow-water setting, calling into question some of the proposed sequence-stratigraphic interpretations of the Sappington Formation as an offshore marine deposit (see Discussion).

The DCB occurs above Unit 4 and remains poorly defined at Logan Gulch. An unconformity occurs above Unit 5 (palynologically barren) and below the base of the Mississippian CCM Lodgepole Formation (Figures 2, 7). The CCM is comprised of a basal lag with ca. 1 m of black shales overlain by burrowed siltstones, and another very sharp contact beneath glauconitic, crinoidal packstones of the Lodgepole Formation 'Scallion Member'. The CCM black shale (LAs4 assemblage) yielded scarce, badly preserved, dark brown to black trilete spores and abundant *Botryococcus braunii*. A mid-late Tournaisian age is assigned, based on biostratigraphically useful species such as *Waltzisporea polita*, *Leiotriletes* sp. cf. *L. sphaerotriangulus*, *Punctatisporites* sp. cf. *P. glaber* and *Botryococcus*. A revision of the age of LAs4 is required since the global range of *Waltzisporea polita* is earliest Visean (or ?latest Tournaisian), and conodonts of *crenulata-isosticha* Zones (Figures 2–3) found in this unit suggest a mid-late Tournaisian (Figure 7). The dark brown shales (LAs5 assemblage) above the basal Scallion limestone of the locally deposited 'False Bakken' yielded poorly preserved spores not diagnostic for age.

The palynological record at Logan Gulch and previous conodont studies in Montana support hiatuses in the late Famennian and earliest Tournaisian near the DCB. Therefore, these cratonic strata of the CMT show changes in base level, and anoxic to disoxic events which, especially in the Sappington Formation, allow for correlation with other DCB formations worldwide. The location of the Hangenberg Event correlates with the Middle Sappington Formation (units 2–5), with the base of the LN zone located in Unit 4. The anoxia phase corresponds to global events with analogues to several latest Famennian assemblages of the HBS event in USA, Greenland, Brazil, western and eastern Europe, and Australia (Figure 7). Significant changes in ice-sheet

dynamics associated with either the onset of glaciation or potentially intermittent deglaciation during the DCB interval can be used to place quantitative constraints on the evolutionary paths of glaciation and low-latitude proxies during this interval.

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## Appendix 1

Complete list of species recorded by Sandberg et al. (1972) in Unit 4, Sappington Formation, at Hardscrabble (Peak 9559), Bridger Mountain. Those authors presented an illustrated list that we update here with current names, whilst the IDs and relevant illustrations given by the former authors are included in parentheses.

### Spores

- Archaeoperisaccus* sp. (pl. 4, fig. 12)
- Cyrtospora cristifera* (Luber) emend. Van der Zwan (*Lophozonotriletes cristifer*, pl. 4, fig. 9)
- Emphanisporites rotatus* McGregor emend. McGregor (*Emphanisporites* cf. *robustus*, pl. 4, fig. 16)
- Grandispora echinata* Hacquebard (pl. 3, figs 27–32)
- Hystricosporites* sp. (pl. 4, fig. 13. Remarks: it is similar to *Hystricosporites porrectus* (Balme & Hassell) Allen, and it was not recorded at Logan Gulch)
- Knoxisporites* sp. cf. *heredatus* (Ischenko) Playford (*K. pristinus* Sullivan, pl. 3, figs 22–23)
- Raistrickia macrura* (Luber) Dolby & Neves (pl. 4, fig. 18)
- Retispora lepidophyta* (Kedo) Playford (*Hymenozonotriletes lepidophytus*, pl. 3, figs 1–5, pl. 4, figs 1–4)

*Retusotriletes incohatus* Sullivan (pl. 3, figs 15–21)

*Rugospora radiata* (Juschko) Byvscheva (*Hymenozonotriletes famenensis* pl. 4, fig. 5)

*Spinozonotriletes conspicuus* Playford (pl. 4, fig. 17)

*Tumulispora rarituberculata* (Luber) Potonié (*Lophozonotriletes* cf. *malevkensis*, pl. 3, figs 11–12, pl. 4, fig. 7.

*Vallatisporites drybrookensis* (morphon) Playford & McGregor (*V. pusillites*, pl. 3, figs 6–8)

*Vallatisporites splendens* Staplin & Jansonius (*V. pusillites*, pl. 3, figs 9–10)

*Vallatisporites vallatus* Hacquebard (pl. 4, fig. 8)

*Vallatisporites verrucosus* (V. cf. *verrucosus*, pl. 3, figs 13–14)

*Velamisporites perinatus* (Hughes & Playford) Playford

*Verrucosiporites* sp. (*Pustulatisporites gibberosus* Hacquebard, pl. 4, fig. 15. Remarks: this specimen shows sparse warts with rounded profile so it is considered to belong in the *Verrucosiporites* probably attributable to *V. mesogrumosus* (Kedo) Byvscheva which is recorded in the LAs3 at Logan Gulch)

### Acritarchs

*Gorgonisphaeridium winslowiae* Staplin et al. (*Gorgonisphaeridium* sp., pl. 3, figs 24–26)

Sandberg et al. (1972) also included a list of species not figured so their identifications cannot be confirmed, but valid (current) names are provided and their original names are given in parentheses.

### Spores

- Cristatisporites echinatus* Playford
- Baculatisporites fusticulus* Sullivan
- Dictyotriletes trivialis* (Naumova) Kedo
- ?*Grandispora spiculifera* Playford (*Geminispora* cf. *svalbardiae* Naumova)
- Grandispora famenensis* (Naumova) Streeel (*Acanthotriletes famenensis* Naumova and *Hymenozonotriletes famenensis* Naumova)
- Retusotriletes punctatus* Chibrikova
- Spinozonotriletes* cf. *S. tenuispinus* Hacquebard
- Grandispora senticosa* (Ishchenko) Byvscheva (*Spinozonotriletes* cf. *S. uncatus* Hacquebard)
- Tumulispora ordinaria* Staplin & Jansonius
- Ancyrospora* sp.
- Cymbosporites* sp.
- Convolutispora* sp.
- Corbulispora* sp.
- Dibolisporites* sp.
- cf. *Leiozonotriletes* sp.

### Acritarchs

- Baltisphaeridium* sp.
- Veryhachium* sp.