Acritarchs in Lower Greenschist Facies Argillite of the Middle Proterozoic Libby Formation, Upper Belt Supergroup, Montana

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Lower greenschist facies dark gray argillite of the Middle Proterozoic Libby Formation in northwestern Montana contains a low abundance and low diversity acritarch microbiota. Four morphological types of acritarchs have been detected: 1) Small (8 to 32 μm in diameter) spherical types (leiomorphs) are represented by four categories based on size, wall configuration, and spatial arrangement of cells; 2) large (up to 150 μm maximum diameter) ovoidal, thick-walled, dense types with granular surfaces; 3) fusiform types; and 4) an odd, torus-shaped type. Most of the acritarchs average between 10 and 40 μm in diameter.

These microfossils are significant for several reasons: 1) They occur in low-grade metamorphic sedimentary rocks. Normally, it is expected that such metamorphism will destroy organic-walled microfossils. 2) The Libby microfossils augment our understanding of Middle Proterozoic microbial assemblages in western North America where only a few other localities of this age are known. 3) Taphonomic considerations aside, the low abundance and diversity of this microbiota may be a function of either its age and/or environment in which the organisms lived. 4) The Libby microbiota appears to be slightly more diverse than sphaeromorph acritarch assemblages from unmetamorphosed lower Belt Supergroup strata in the eastern Belt basin. This suggests one of two things: a)

The Libby acritarchs inhabited a more open marine environment than those from older strata in the east, or b) a paleobiologically significant amount of time separates the two biotas from the lower and upper Belt, such that by Libby time, sphaeromorph acritarchs had diversified somewhat. 5) The acritarchs resemble other Middle Proterozoic spheroidal acritarch assemblages. However, no biostratigraphically diagnostic taxa that might help clarify controversial upper Belt ages have been discovered in these samples. This might be a function of taphonomy, metamorphism, and age of the strata as well as the fact that morphologically distinctive, relatively short-lived species are not common among Middle Proterozoic acritarch taxa.

We emphasize that, although it is desirable to search for remains of Proterozoic life in unmetamorphosed rocks, less attractive lithologies should not be ignored. Weakly metamorphosed strata can provide additional sources of paleontological information.

INTRODUCTION

Acritarchs are organic-walled microfossils of uncertain taxonomic affinity that are normally studied from acid-resistant materials recovered by palynological techniques applied to fine-grained clastic rocks. Although acritarchs have been known for many years, it wasn't until the late 1940's and early 1950's that their potential for biostratigraphic analyses was fully recognized in Paleozoic rocks (e.g., Naumova, 1949). The use of acritarchs to help resolve Proterozoic stratigraphic problems was developed in the USSR, apparently first by Naumova (1951) on the acri-

tarchs from Late Proterozoic rocks in the Urals and soon followed by the works of Timofeev (e.g., 1955, 1959, 1969) from Late Proterozoic sequences in Siberia and the Baltic region. Rather intensive study of Eurasian Proterozoic acritarchs by micropaleontologists has produced a large body of data that allows for rather precise correlations of Late Proterozoic strata (Timofeev, 1959, 1969; Roblot, 1963; Volkova, 1968; Hsing [Xing] and Liu, 1973; Vidal, 1976b). The application of acritarchs for resolving Proterozoic chronostratigraphic problems has now been applied elsewhere (for example, Peat et al., 1978; Cloud and Germs, 1971; Lenk et al., 1982; Amard, 1986; Damassa and Knoll, 1986).

Unmetamorphosed, fine-grained sedimentary rocks are the rocks of choice for most acritarch research. Metasediments are usually avoided because of the thermal alteration and destruction of organic-walled microfossils during metamorphism (Vidal, 1981). Several examples of acritarchs, however, have been recovered from low-grade metamorphosed clastic rocks (for example, Vialov and Timofeev, 1959; Vidal, 1976a; Konzalova, 1981; Amard, 1983; Pflug and Reitz, 1988). Acritarchs and filamentous microfossils have also been recovered from chert nodules in shales metamorphosed to lower greenschist facies (Knoll and Ohta, 1988). High-grade metamorphism destroys the structural integrity of microfossils; upon maceration, minute kerogen or graphite particles are all that remain. In addition, research by Wang and Luo (1982) suggests that a fortuitous organization of kerogen particles can occur during some maceration processes, resulting in the formation of pseudomicrofossils. Apparently this is the origin of the alleged microfossil *Manneria* Lopuchin (Lopuchin, 1975).

The low-grade metamorphic argillite of the Middle Proterozoic Libby Formation (upper Belt Supergroup) contains a reasonably well-preserved acritarch microbiota that has been detected in both macerates and in thin section. These acritarchs are low in diversity and abundance; however, they and discoveries like these in other metamorphic terranes are highly significant because they have the potential to help resolve important chronostratigraphic problems that are common for many metamorphosed Proterozoic sedimentary sequences. The age of the upper Belt Supergroup is uncertain, controversial, and a case in point. Based on radiometric, paleomagnetic, and lithostratigraphic analyses, ages of 900 Ma and >1200 Ma have been interpreted for upper Belt rocks (see discussion below). The discovery of acritarchs in the Libby Formation provides an opportunity to employ paleontological methods toward resolution of this controversy. Although metamorphosed, the Libby acritarchs are sufficiently well-preserved to warrant an attempt to evaluate their biostratigraphic usefulness.

Before one can expect the Libby acritarchs to be the remedy for solving upper Belt age problems, some caveats are appropriate here. The metamorphism of the Libby, although low, might have destroyed or altered potentially useful taxa. Diagnostic, relatively short-lived acritarch taxa are rare for Middle Proterozoic strata except for those of late Middle Proterozoic (approximately 1050 to 900 Ma)

age. Most Middle Proterozoic acritarchs are simple spheroids and thus taxonomy is difficult. Middle Proterozoic acritarchs have not been studied in as great a detail as those from the Late Proterozoic, and, hence, precise biostratigraphically determined ages cannot be made with a great deal of confidence.

In addition to potential use for helping in chronostratigraphic problems, the Libby acritarchs provide additional micropaleontological information for the Middle Proterozoic, an interval of pre-Phanerozoic time which is biogeologically important because of the early radiation of eukaryotes (Vidal, 1984).

GEOLOGIC SETTING

The Belt Supergroup is a thick succession of predominantly fine-grained, weakly metamorphosed sedimentary rocks deposited over a large area in northwestern Montana, northern Idaho, northeastern Washington, and southern British Columbia and Alberta (Fig. 2). The supergroup overlies crystalline basement in the southeast, and it is unconformably overlain by the Late Proterozoic Windermere Supergroup in the northwest and by Paleozoic strata in other parts of the region. Belt sedimentation was primarily turbiditic during the early history of the basin (Cressman, 1985), and was dominated by shallow-water deposits throughout the remainder of its history (Harrison, 1972; Winston, 1986). Depositional environments in these rocks are controversial, with both marine (Price, 1964; Harrison, 1972; McMechan, 1981) and non-marine (Winston et al., 1984; Winston, 1986) interpretations proposed.

The Libby Formation is the uppermost Belt unit in the western part of the basin in the United States (Figs. 1, 2). The lower Libby consists of greenish-gray and dark gray argillite and siltite that were deposited in shallow water environments associated with stromatolites, ooids, shrinkage cracks, and symmetrical ripples that can be interpreted as either a tidal flat or periodically exposed lacustrine setting (Kidder, 1988a, b). No evaporites have been detected in the Libby, although salt casts are preserved near Missoula (Fig. 1) in the McNamara Formation, a unit that correlates to the lower Libby Formation (Kidder, 1988b). The upper Libby is characterized by hummocky cross-stratified, coarse-grained siltite and fine-grained arkosic and subarkosic wackes and arenites that do not display the shallow-water features present in the lower Libby (Kidder 1988a, b). The upper part of the Libby Formation was probably deposited below fair-weather wave base, but without diagnostic fossils or geochemical data, confident placement of the upper Libby into a marine lacustrine setting is difficult.

Lower greenschist metamorphism is inferred for the Libby Formation near Libby, Montana based on proximity to equivalent exposures east of Clark Fork, Idaho, approximately 40 km southwest of the acritarch-bearing rocks of this study. Samples of the Libby Formation at Clark Fork were interpreted as lower greenschist facies based on the presence of 2M muscovite (Maxwell and Hower, 1967). Chlorite (x-ray analysis of argillite) has been reported from
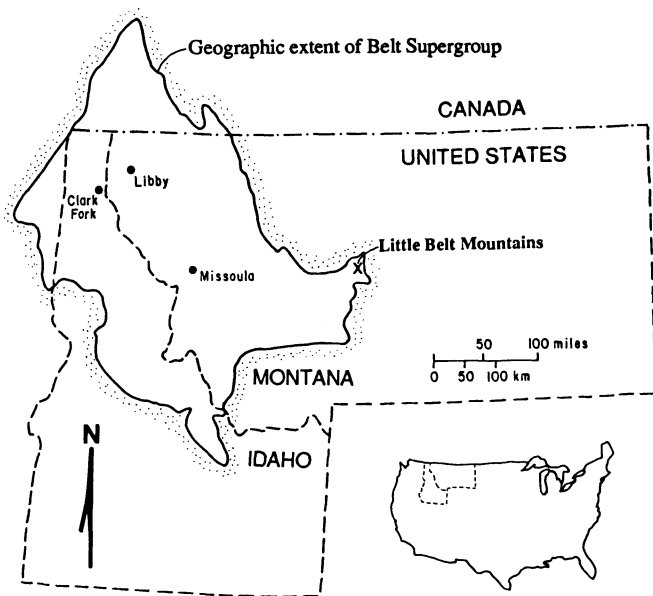
FIGURE 1—Location map showing geographic extent of Belt Super-group and localities mentioned in text.
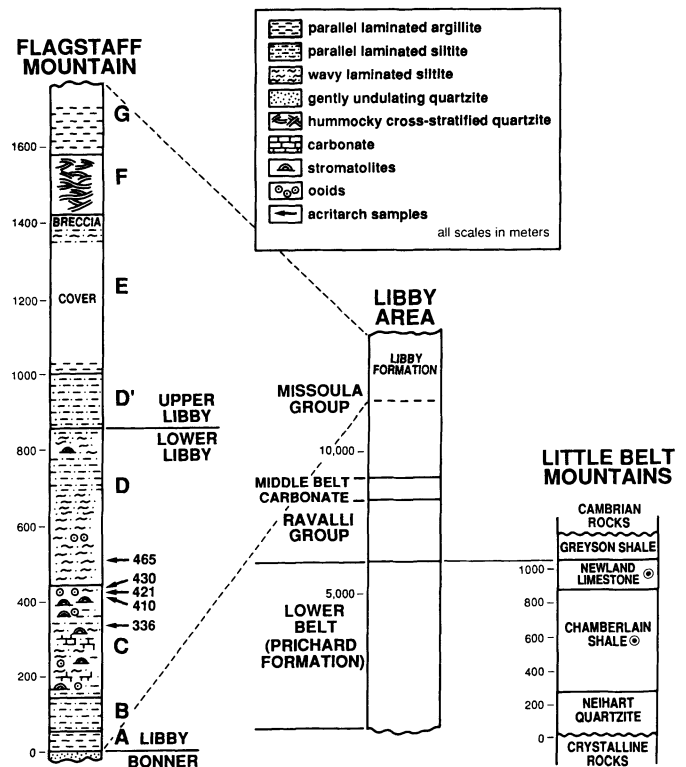

FIGURE 2—Stratigraphy of the type locality of the Libby Formation along the Kootenai River about 8 miles west of the town of Libby, Montana, and correlation to acritarch-bearing formations in the lower Belt Supergroup (solid line). Shown are the position of the Libby Formation within the belt (dashed lines), informal members within the Libby Formation after Kidder (1988a, b), and zones in which acritarchs have been recovered.

the same locality (Eslinger and Sellars, 1981). We interpret the metamorphic grade at Libby to be similar to that at Clark Fork based on the slaty nature of some beds and the general field and petrographic similarities of the Libby Formation at Libby and Clark Fork.

Older units in the Belt are also interpreted to have been metamorphosed. At Clark Fork, metamorphic grade increased down section from the Libby to biotite grade in the basal Prichard Formation (Maxwell and Hower, 1967; Eslinger and Sellars, 1981). Oxygen isotopic analyses of coexisting quartz and illite phases in the Middle Belt Carbonate and Ravalli Group in Glacier National Park (approximately 150 km east of Libby) led Eslinger and Savin (1973) to conclude that burial metamorphism elevated temperatures in these older Belt rocks to 225°–310° C.

Age

Definitive age data for the Belt Supergroup are limited. A maximum age of approximately 1700 Ma for these rocks is derived from a 1700 Ma Rb-Sr whole rock isochron age for metamorphosed crystalline basement rocks (Giletti, 1966) and a metamorphic U-Pb age of 1668 ± 32 Ma from zircons from migmatitic paragneiss beneath the Belt (Armstrong et al., 1987). Minimum ages are older than 770 Ma for rocks just beneath the Windermere Supergroup from the MacKenzie Mountains Supergroup in the Northwest Territories of Canada. These minimum ages are based on Rb-Sr dates on several types of mineral fractions separated from diabase by magnetic and density methods (Armstrong et al., 1982). Older minimum ages (827–918 Ma) have been obtained by K-Ar analyses on whole-rock sam-

ples and plagioclase separates from the basaltic greenstone assemblage in the Windermere Supergroup that unconformably overlies the Belt Supergroup and equivalent Deer Trail Group in northeastern Washington (Miller et al., 1973; Miller and Whipple, 1989).

Ages for rocks within the supergroup include the lower Belt Prichard Formation age of 1433 ± 13 Ma determined by U-Pb analysis of zircon from the Crossport C sill in Idaho (Zartman et al., 1982). Ages for stratigraphically younger Belt rocks are less reliable. Rb-Sr whole rock isochron analyses of argillite (Obradovich and Peterman, 1968, 1973; Obradovich et al., 1984) yield ages of approximately 1300 Ma (Ravalli Group), 1100 Ma (Missoula Group), and 900 Ma (Garnet Range Formation and Pilcher Quartzite). K-Ar analyses of glauconite (Obradovich and Peterman, 1968) yield ages of about 1100 Ma for rocks from the Ravalli Group through McNamara (=lower Libby) Formation. The most direct radiometric age estimate for the Libby Formation comes from a stratigraphic correlation (Harrison, 1972) to an argillite in the Garnet Range Formation (upper Libby) in the Alberton region approximately 50 km west of Missoula in which Rb-Sr whole rock

analyses have yielded a 900 Ma-old age (Obradovich et al., 1984).

Paleomagnetic studies suggest that the entire Belt Supergroup may be older than 1200 Ma. Elston and Bressler (1980) and Elston (1984) presented a sequence of magnetic reversals and a polar wandering curve from red-bed facies in the Belt Supergroup. Although the magnetic reversals may provide time lines for correlation within the Belt basin, the ages assigned to the time lines depend primarily on correlation of the Pilcher Quartzite normal-polarity pole to a statistically identical normal-polarity pole for the Bass Limestone of the basal Grand Canyon Supergroup. The Missoula Group has been correlated to the Pioneer Shale (Apache Group, central Arizona) and the Sibley Group (Lake Superior region, Ontario), based on reversed-polarity intervals and similar pole positions (Elston, 1984). Paleomagnetic and radiometric age determinations are in relatively close agreement for the lower Belt, suggesting approximate 1450 Ma ages, but results diverge widely in the younger Missoula Group. This approach provides potential for correlation in these rocks, but more complete coverage of sequences and a better understanding of magnetic carriers are needed before Proterozoic paleomagnetic correlation can be accepted as reliable.

Although stromatolites are common in many Belt units (Rezak, 1957), the biostratigraphic resolution provided by these organosedimentary structures for resolving the age problems encountered in the Belt has not been attained. Some of the apparently more useful stromatolite taxa in Proterozoic biostratigraphic correlations, such as *Conophyton* and *Baicalia,* occur in the Belt Supergroup. White (1970) identified *Baicalia* from the upper part of the Lower Belt Altyn Formation suggesting a 1350 Ma age for the formation, although he later revised his age estimate to 1450 Ma (White, 1984). These taxa, in particular *Conophyton,* are long ranging in the Proterozoic (Bertrand-Sarfati and Walter, 1981). *Baicalia–Conophyton* cycles are found in the Upper Helena (Siyeh) Formation (Horodyski, 1976, 1983), and Serebryakov (1976) indicates that these cycles are most typical of the Middle Riphean (1350–1050 Ma ago). However, Horodyski (1983) feels that the *Baicalia–Conophyton* cycles in the Belt should not be used for chronostratigraphic purposes; the stromatolite morphology appears to be strongly influenced by environmental factors. He refrains from using any of the Belt stromatolites for biostratigraphic purposes (see Horodyski, 1989b). Stromatolites do occur in the Libby Formation, but they are flat to mound-shaped, laminated structures with poorly preserved microstructure. Such generalized stromatolite morphologies are not found to be very useful or diagnostic for biostratigraphy when compared to columnar branching forms with well-preserved microstructure (e.g., Bertrand-Sarfati, 1972). Thus the Libby stromatolites provide no new data for a more precise age determination for the upper Belt.

Horodyski (1989a) reviewed the microbial fossil record in Belt carbonate (Snowslip Formation), chert (Altyn Formation, Helena (Siyeh) Formation), and mudstone (Appekunny Formation, Chamberlain Shale). Although Horo-

dyski (1989a) did not attempt to biostratigraphically analyze these microfossils, he noted (Horodyski, 1980) that the mudstone microfossils, in particular the acritarchs, are potentially valuable for intercontinental biostratigraphic correlations. Aside from the filaments, which have little or no biostratigraphic value, the sphaeromorphs hold the greatest promise. However, only *Kildinosphaera* cf. *lophostriata* (Jankauskas) Vidal (1983) apparently is an Upper Riphean taxon (Vidal and Siedlecka, 1983); however, kildinellid-type morphs are long-ranging, known from Lower Riphean to Vendian strata (Hofmann, 1988). Although potentially useful microfossils occur in Belt rocks, they have not yet contributed to the chronostratigraphic understanding of the Belt Supergroup. Possible biogenic microstructures were reported from the Libby Formation and other Belt units by Pflug (see Horodyski, 1989a) at Clark Fork, but the biogenicity of these forms remains dubious (Horodyski, 1989a).

THE LIBBY ACRITARCHS

In the Libby, acritarchs were recovered only from members C and D in the lower Libby (Fig. 2). Member C is characterized by dark gray argillite and greenish gray siltite interbedded with stromatolitic and oolitic carbonate beds. Member D is lithologically similar to member C, except that it lacks carbonate, and the siltite beds are often wavy laminated. The insoluble residue of macerated samples consists primarily of disseminated organic material with rare microfossils. Color of both the organic matter and most of the microfossils is predominantly very dark brown, almost black, reflecting the thermal alteration due to the low grade metamorphism (Staplin, 1977). In thin section, the acritarchs and organic matter show similar colors. Rarely were microfossils observed in maceration that were greenish- to brownish-gray in color. The grayish coloration might be due to the metamorphism (Konzalova, 1974, 1981). Most acritarchs found in both macerates and thin section are solitary sphaeromorphs with rare pluricellular aggregates and nonspheroidal morphs. No filamentous morphs have been discovered. It is possible that filamentous microfossils, if originally present, were more susceptible to destruction by metamorphism.

Forty-five samples of fine-grained clastic rocks (argillite and siltite) from Flagstaff Mountain, part of the type locality of the Libby Formation (Fig. 2), were processed for acritarchs. This locality was selected for intensive study because of near completeness of the section, excellent exposures, relatively fresh outcrops, low-grade metamorphism did not appear to substantially alter the rock, and preliminary study revealed the presence of acritarchs. The samples were processed as follows: 1) 30–50 mg rock samples were scrubbed and soaked in a solution of potassium dichromate and reagent grade sulfuric acid to remove organic contaminants from sample surfaces and cracks. 2) Samples were then crushed with a mortar and pestle to particles with maximum diameters in the granule size range (about 4 mm). 3) Some powdered sample from the crushed material was squirted with 10% HCl to check for a car-
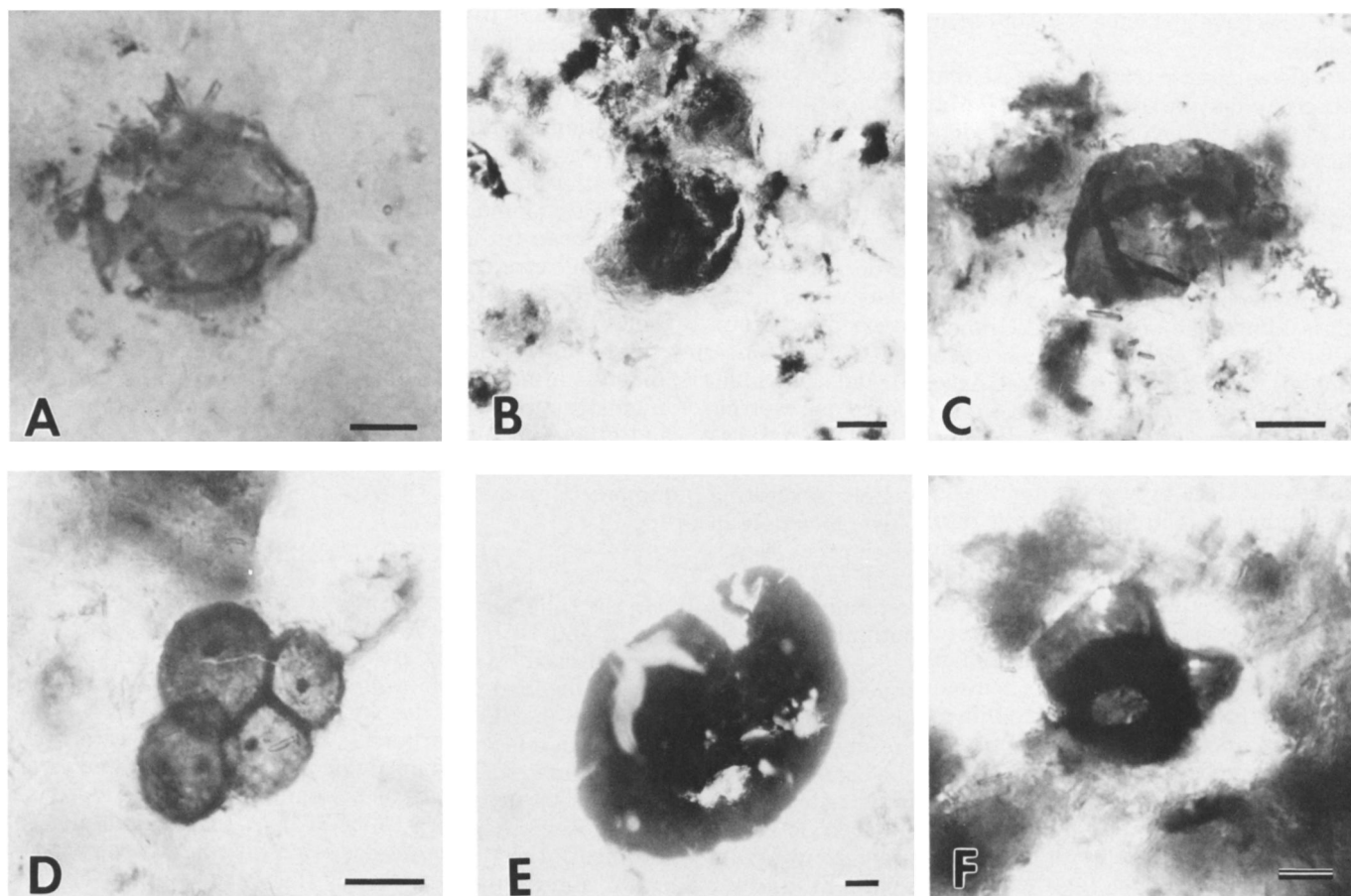
FIGURE 3—Photographs of the Libby microbiota. All photomicrographs from macerated material; bar scale = 10 μm. Locations of figure specimens are given in coordinates in millimeters from a reference "×" scribed into the slide ("×" is to the viewer's lower left). **A.** Leiosphaerid A (slide FC110) P-1; 21.5 × 12.2); **B.** Leiosphaerid B (slide FC101 P-2; 12.3 × 8.7); **C.** Leiosphaerid C (slide FC101 P-2; 33.4 × 17.0); **D.** Leiosphaerid D (slide FC101 P-2; 39.3 × 5.5); **E.** Non-leiosphaerid morph (slide FC16 P-3; 17.5 × 11.0); **F.** Torus-shaped morph (slide FC101 P-2; 29.1 × 3.1).

bonate reaction. If there was no reaction, steps 4 and 5 were skipped. 4) Carbonate-bearing samples were soaked in 10% HCl for several days to two weeks depending on the duration of the reaction. 5) Samples from step 4 were rinsed by multiple centrifuge treatment. 6) Carbonate-free samples were soaked in concentrated HF for several days. These samples were periodically gently stirred and supplied with fresh HF until most or all fragments were dissolved. 7) The insoluble residue from HF-treated material was rinsed by placing samples in sealed bags constructed of dialysis tubing and left in running de-ionized water overnight. 8) Strew mounts of the residue were prepared with a mixture of Kumar ® resin and xylene. This produces semi-permanent slide preparations.

In addition to the palynological procedures employed for the microfossil study, thick (ca. 50 μm) thin sections of the clastic rocks were made from samples cut parallel to bedding.

Description of Acritarchs

The acritarchs thus far detected in Libby Formation argillite are mostly thin-walled and spheroidal to irregular in shape. No filamentous morphs have been detected. Preservation is moderate at best, and the number of microfossils found in thin section was under ten per thin section and in strew mount preparations no more than 40 per slide. The size and morphology of acritarchs found were quite variable. Diagenesis and low-grade metamorphism have probably altered the original shape and features of the microfossils such that the assignment of the Libby morphs to previously described acritarch taxa is difficult and of doubtful validity. No new taxa are described, no taxa are formally identified, and only broad comparisons are made with taxa described in the literature.

The microfossil assemblage is dominated by thin-walled, unornamented sphaeromorphs, 7–75 μm in diameter (al-

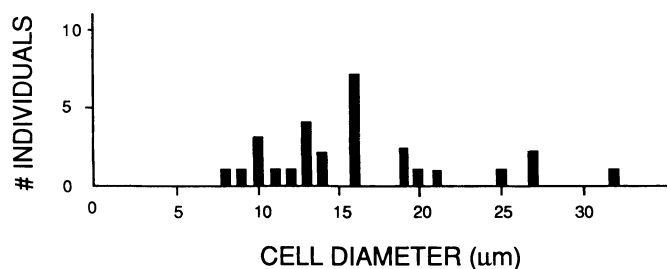FIGURE 4—Plot showing size distribution of microfossils grouped as leiosphaerid A.

FIGURE 5—Plot of length vs. width of fusiform acritarchs in the Libby Formation.

though a few specimens achieve sizes up to 150 μm), with moderately smooth to granular (shagrinate) surface textures. Such acritarchs can be classified as leiosphaerids (Lindgren, 1981; Jankauskas and Mikhailova, 1986). Fusiform microfossils and two morphologically distinctive morphs, one a large 54 to 150 μm diameter, ovoidal, dense microfossil and the other a torus-shaped microfossil, have also been detected. No recognizable acritarchs smaller than 7 μm have been found. Neither have any large (>150 μm) sphaeromorphs nor acritarchs with processes or other surface ornamentation been found.

The Leiosphaerids

Within the Libby Formation, four morphologically distinctive leiosphaerid forms are known (Fig. 3A–D):

1) Leiosphaerid A, with diameter ranging from 8 to 32 μm ($\bar{x}$ = 17.5 μm; Fig. 4), is characterized by a narrow, well-defined rim that incompletely surrounds the thin, moderately smooth, greenish-gray to brownish-gray vesicle with thin, wispy folds and equatorially located, slit-shaped cracks (Fig. 3A). The origin of the rim is unclear; it usually does not completely envelop the acritarch. In some cases it appears to be an an artifact of viewing the microfossils in a plane, where folding of the wall parallels the outer boundary. The rim does not appear to be the remains of extra-vesicle organic material. The coloration of the rim is not significantly different from the rest of the microfossil. These morphs are the most abundant category of leiosphaerids found in the Libby (n = 26).

2) Leiosphaerid B is a round, dense, dark brown, somewhat smooth-surfaced, unfolded vesicle with a featureless wall (Fig. 3B). Only four examples have been found, and diameters range from 19 to 30 μm. One microfossil has a peripherally located crack that is either a "median" split or a diagenetic feature.

3) Leiosphaerid C is represented by one brown colored fragment (approximately one fourth of a vesicle), with a maximum dimension of 27 μm that extrapolates to a diameter of approximately 108 μm. Distinctive, intersecting folds characterize this morph (Fig. 3C). Cell walls are thin and surface texture is smooth to slightly granular.

4) Leiosphaerid D is a cluster of four cells, averaging 9 μm in diameter with thin walls, granular surface textures,
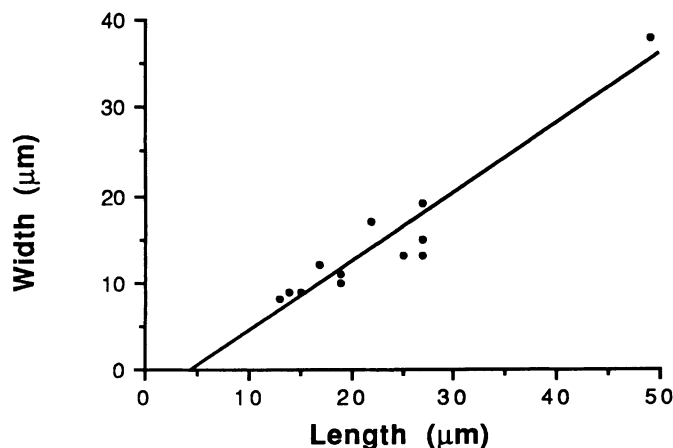
and each with a small, micrometer-diameter dense internal inclusion (Fig. 3D). The inclusions might represent the degraded remains of internal cellular material. We place no biological or taxonomic significance on the inclusions. This cluster is tentatively grouped with the leiosphaerids based on considerations presented in Lindgren (1982) on the postulated vegetative growth of leiosphaerids. The individual cells of the Libby cluster are about 9 μm in diameter, very near the lower size limit of all Libby leiosphaerids.

A number of fusiform fossils add to the assemblage. The two best preserved fusiform morphs are dark brown, have smooth walls, and are characterized by abundant, irregularly oriented and irregularly distributed folds crudely sub-perpendicular to cell margins. These morphs range in length from 22–27 μm and in width from 17–19 μm. Ten additional fusiform morphs that lack surface folds display colors of greenish-gray and light, medium, and dark brown. Length of these forms ranges from 13–49 μm, and width ranges from 8–38 μm. Length plots as a straight line against width with a good fit (Fig. 5; R = 0.90) suggesting a taxonomic relationship among all of the fusiform morphs. Due to the difficulty in obtaining high quality photomicrographs of the fusiform type, the morph is not illustrated.

Another non-leiosphaerid morph is represented by three large (75–150 μm length of long axis), ovoidal, dense, dark brown, thick-walled, vesicles with granular surface textures. These microfossils are commonly cracked, with two types of cracks: small (a few micrometers in length) superficial cracks, and large cracks, a few tens of micrometers in length, that deeply incise the spheroid and occasionally bifurcate (Fig. 3E).

Of the remaining non-leiosphaerid morphs, one, a torus-shaped form (n = 1), is the most curious (Fig. 3F). It consists of a larger 28 μm diameter, brown, dense, thick-walled central spheroid with two smaller 11 and 20 μm sized, thinner-walled, hemispheres attached. A possible interpretation for this form is it is the deformed, degraded remains of a leiosphaerid triad.

DISCUSSION

The Libby microbiota is substantially different from the shale and limestone microbiotas of the Chamberlain Shale and Newland Limestone described by Horodyski (1980, 1989a). The Libby differs from the lower formations in lacking a) filamentous morphs, b) prominently folded and wrinkled forms assigned to *Kildinosphaera* cf. *lophostriata* (Jankauskas) Vidal, and c) polygonally segmented or bumpy surface sphaeromorphs referred to *Satka* cf. *colonialica* Jankauskas (Horodyski, 1989a). No morphs appear to be common between the Libby and those reported by Horodyski from older units. A general comparison of the sphaeromorphs in the Libby Formation with those described by Horodyski (1980) suggests that the Libby may have a slightly more diverse assemblage than rocks of the lower Belt Supergroup, even though the lower Belt rocks studied by Horodyski (1980) are unmetamorphosed (Maxwell and Hower, 1967).

The differences in composition and diversity between these upper Belt and lower Belt microbiotas could be either a function of depositional environment or age. With regard to significant differences in age, this depends on the age of the Missoula Group. As we have discussed, these age estimates range from about 900 Ma (which would still make the Libby microbiota 100 to 150 m.y. younger than the Newland and Chamberlain) to >1200 Ma (which would make the Libby somewhat younger than the lower microfossiliferous formations). With regard to differences in depositional environment, the Newland Limestone and Chamberlain Shale were deposited in the Helena embayment, an extension of the Belt basin that may have been a more restricted environment than the area in which the Libby Formation was deposited (Schieber, 1986; Zieg, 1986). The presence of filaments in the Newland and Chamberlain suggests a shallow, nearshore source for these presumably benthic microfossils. The Libby Formation was deposited in the central and western Belt basin, which may or may not have been connected with an open ocean. There is no evidence that the stromatolites formed a barrier resulting in a restricted environment like a lagoon. The lack of filaments suggests a) no shallow, nearshore sources contributed to microfossiliferous sediment, and/or b) in the Libby, filaments were more susceptible to destruction by metamorphism.

Although the microbiota lacks taxonomically distinctive, shorter-lived, higher resolution acritarch taxa, the biota (based primarily on the lack of diagnostic Late Proterozoic taxa) does suggest a Middle Proterozoic age and this is in agreement with all other chronostratigraphic methods. Leiosphaerid-dominated assemblages are common in Late Proterozoic (900 to 570 Ma) clastic rocks; however, these assemblages also contain other, distinctive sphaeromorphic microfossils, such as *Chuaria circularis* Walcott, *Trachysphaeridium laminaritum* (Timofeev) Vidal, *Tasmanites rifejicus* Jankauskas (Vidal and Knoll, 1983), indicating the younger Proterozoic age. It is clear that Late Proterozoic acritarchs in open marine settings are quite diverse (Vidal and Knoll, 1983), but the general

diversity level of Middle Proterozoic acritarchs is poorly known, but appears to be much lower (Jankauskas, 1982).

The diversity of the Libby spheroidal acritarchs appears to be somewhat higher than older Belt non-filamentous acritarch microbiotas (although direct comparison of specimens is needed to verify this). However, the Libby diversity is low when compared to other Proterozoic microbiotas. This low diversity is a function of 1) age of the Libby, 2) a restricted environment, and/or 3) preservational problems. The low diversity of the Libby microbiota is comparable with diversity in stressed sedimentary environments in younger Proterozoic rocks (Knoll, 1984). Paleoecological studies of Late Proterozoic acritarchs have shown that low diversity microbiotas are common to restricted sedimentary environments, whereas open marine settings usually contain a higher diversity (Vidal and Knoll, 1983; Knoll, 1984). Paleozoic acritarch paleoecological studies show similar patterns (Jacobson, 1979; Smith and Saunders, 1970; Staplin, 1961). However, the low diversity of the Libby Formation could also be a function of age if diversity was generally low in the Middle Proterozoic. Late Proterozoic acritarch diversity can be strongly influenced by depositional environment. Biostratigraphic success in the Proterozoic to date has been achieved with *marine* rocks. Therefore, even if the acritarch data seemed biostratigraphically more definitive, they would still be suspect because of the uncertainty of the depositional environment of the Belt Supergroup discussed earlier.

CONCLUSIONS

Although Middle Proterozoic acritarch microbiotas are poorly known relative to those of the Late Proterozoic, much can still be learned by the study of such microbiotas. The specific conclusions of this study are as follows:

1. Low-grade metamorphic mudrocks can sometimes preserve acritarchs sufficiently well for paleoenvironmental analysis and perhaps biostratigraphic correlation.

2. The Libby microbiota is a low diversity, low abundance assemblage of dominantly spheroidal acritarchs. Surface sculpture varies from psilate to slightly folded. Acritarch diameter ranges from 7 to 150 μm, with most morphs having a diameter of 10 to 40 μm.

3. These acritarchs offer no clarification of the controversial age of the upper Belt Supergroup. This low diversity microbial assemblage is superficially similar to other Middle Proterozoic acritarch biotas (e.g., Jankauskas, 1979; Chen et al., 1980) but appears to lack distinctive, taxonomically identifiable morphs that might help with the question of an older or younger Middle Proterozoic age.

4. The cause of the limited diversity in the Libby microbiota could be a result of a general low diversity in Middle Proterozoic acritarchs, an effect of a restricted depositional environment that stressed the microbiota, or selective destruction of forms resulting from the metamorphism. If non-filament diversity is indeed higher in the Libby when compared to acritarchs in older, less metamorphosed Belt rocks, then age and/or depositional environment influenced diversity. If age (i.e., evolutionary

diversification) is the main influence, Beltian acritarchs diversified only minimally in the 200–500 million years that separate the top and bottom of the supergroup.

5. Middle Proterozoic clastic rocks, both metamorphosed and unmetamorphosed, should be investigated more thoroughly for acritarchs. Sediments of such age were deposited during a significant interval of geological time when eukaryotes evolved and were diversifying (Vidal, 1984). Further knowledge of the Middle Proterozoic fossil record is critical for our understanding of pre-Phanerozoic biogeology.

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