close modern inorganic analogue with corrugations has not yet been found.

Inasmuch as I favor the biologic interpretation of the corrugated structures, I propose a formal taxonomic designation for reference purposes:

Rhysonetron, new genus

Diagnosis: Discrete long, curved, cylindrical, and tapering rods or spindles of fine-to-medium-grained arenaceous material, embedded in similar material; varying in cross section from circular to elliptical, and having longitudinal median marking or sculpture and oblique, lateral, crescentic corrugations; pointed ends may be free or may connect with or overlap other specimens. Length, at least five times greater than width.

Etymology: *rhysos* (Gr.), wrinkled; *netron* (Gr.), spindle

Type species: *Rhysonetron lahtii,* new species (Figs. 4 and 5)

Specific characteristics: Spindles longitudinally differentiated, and marked with well-developed alternating or symmetrically paired corrugations spaced 1 to 3 mm apart.

Etymology: Named in honor of the discoverer, Victor Lahti of Elliot Lake.

Type occurrence: Bar River Formation (Fig. 2; *7*)

Type material: Geol. Surv. Can. holotype No. 9876; paratypes No. 22626, 22627

Rhysonetron byei, new species
(Figs. 7 and 8;
Frarey and McLaren, 1963, Fig. 1)

Specific characteristics: Spindles smaller, more slender, and less well differentiated longitudinally than *R. lahtii*; faint corrugations spaced 0.8 to 1.0 mm apart.

Etymology: Named in honor of the discoverer, C. E. Bye of Sault Sainte Marie, Ontario.

Type occurrence: Crestal area of outcrops of red sandstone in woods 2.6 km northeast of main intersection in Desbarats, Ontario (46°21′40″N, 83°53′55″W). Lorrain Formation

Type material: Geological Survey of Canada holotype No. 15379; paratype No. 22628

Implications. The corrugated structures from Flack Lake are at present the most convincing evidence of the existence of advanced organisms during the Early Proterozoic. Although it is possible that they may be explained as of inorganic origin, it seems unlikely that forms of such complexity could entirely result from some inorganic process. The complex organization favors a biologic interpretation.

If they are in fact biogenic, the implications of this discovery are profound: they may establish the presence of metazoans more than 2000 million years ago, extending the range of our knowledge to at least three times the previously generally accepted age of the oldest metazoan remains. The implications would include problems relating to the definition of the base of the Paleozoic (*18*), and to theories of the geochemical evolution of the oceans and the atmosphere, especially with regard to the appearance of oxygen. Furthermore, they would emphasize that well-sorted arenaceous sediments are a suitable environment for the preservation of Precambrian biogenic structures, and are favorable prospecting grounds for Precambrian fossils.

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22 December 1966

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Mössbauer Analysis of Iron in Clay Minerals

Abstract. *Mössbauer absorption spectrography can be used to establish the presence of* Fe^{2+} *and* Fe^{3+} *in clay minerals. In the sheet structure silicates, octahedrally coordinated iron can be distinguished from tetrahedrally coordinated iron. Siderite and goethite, common contaminants of the clay minerals, can usually be detected. Goethite has a well-organized structure, though, owing to its fine grain size, it may appear to be amorphous to x-rays. The various families of clay minerals show minor differences in isomer shift and quadrupole splitting values, caused by variations in the character of the octahedral layer.*

Most of the clay minerals contain iron in their lattice and often they may contain an admixture or coating of iron oxides. It is difficult and often impossible to determine whether the iron associated with clays is in the lattice or is external. Even if this can be established, there is no means of knowing where the iron occurs in the lattice (tetrahedral or octahedral coordination). It is also usual for most clays to have iron present in both the Fe^{2+} and Fe^{3+} state. In addition, the clay minerals have three different structural arrangements—2 : 1 sheet, 1 : 1 sheet, and chain. The sheet structure types have dioctahedral (two of the three octahedral positions filled) and trioctahedral members. Thus, the octahedral iron occurs in a wide variety of environments. Further, chlorite contains two octahedral layers. In one layer the iron is coordinated largely by oxygen ions, and in the other (brucite layer), by hydroxyl ions (*1*).

We obtained Mössbauer spectra of Fe^{57} in a number of clay minerals and micas to determine the feasibility of using this technique to learn more about the character and location of the iron in clay minerals. A number of studies (*2, 3*) have indicated that the various silicate families have relatively distinct spectra, but there appears to have been no systematic study of any specific family of silicate minerals.

Three useful parameters are readily determined from the Mössbauer interaction: the isomer shift, the nuclear electric quadrupole splitting, and the nuclear magnetic dipole splitting. The isomer shift, which is related to chemical bonding and changes in valence, and the quadrupole splitting, which is sensitive to structural environment, are the prime considerations in this work. The isomer shift is a function of the charge density of electrons at the nucleus (*4*). The quadrupole splitting depends on the asymmetry of the electric field at the nuclear site. The splitting for FE^{+3} is caused by the neighboring ions, but for Fe^{+2} the relevant field gradient is related primarily to the electron distribution in the iron ion (*5*). Previous studies (*2, 3*) have established the following ranges for the values of these parameters in silicate minerals: the isomer shift (measured from the center of the iron metal pattern, −0.25 mm/sec in our work) for Fe^{+3}, from 0 through +0.5 mm/sec; for Fe^{+2}, from +1.0 through +1.4 mm/sec. Quadrupole splitting for Fe^{+3} ranges from 0 through 0.5 mm/sec, and for Fe^{+2}, from 0.40 through 1.7 mm/sec. Table 1 summarizes the data obtained.

For comparison purposes, spectra were obtained for several biotites (Fig. 1). The two peaks are due to the quadrupole splitting of the octahedral Fe^{2+}. The line width at half-height of the +2.1 mm/sec peak is larger than that of the −0.30 mm/sec peak. Pollak *et al.* (*6*) considered this to be due to two slightly different octahedral sites (hydroxyl ions at opposite corners and hydroxyl ions at adjacent corners of the octahedron), both of which produce quadrupole splitting. Each doublet consists of one component at approximately −0.3 mm/sec and the other at about +2.1 mm/sec. The former lines are superposed to produce a large peak and the latter two are slightly offset to produce a broadened peak.

Our biotite spectrum has a small peak at about +0.4 mm/sec which is probably due to a minor amount of Fe^{3+} in the octahedral layer. The other half of the doublet is masked by the large peak at −0.30 mm/sec.

The powdered sample was prepared from the biotite flake and the spectrum obtained was similar to that for the flake, except for broadening of the 2.10 mm/sec peak. It may be that the flake edge area is increased enough so that a significant number of the octahedral sites occur at the edge and form a third type of site. Thus the spectrum would contain three quadrupole doublets.

Table 1. Mössbauer parameters determined in this work for clays and associated minerals; Oct., octahedral; Tet., tetrahedral.

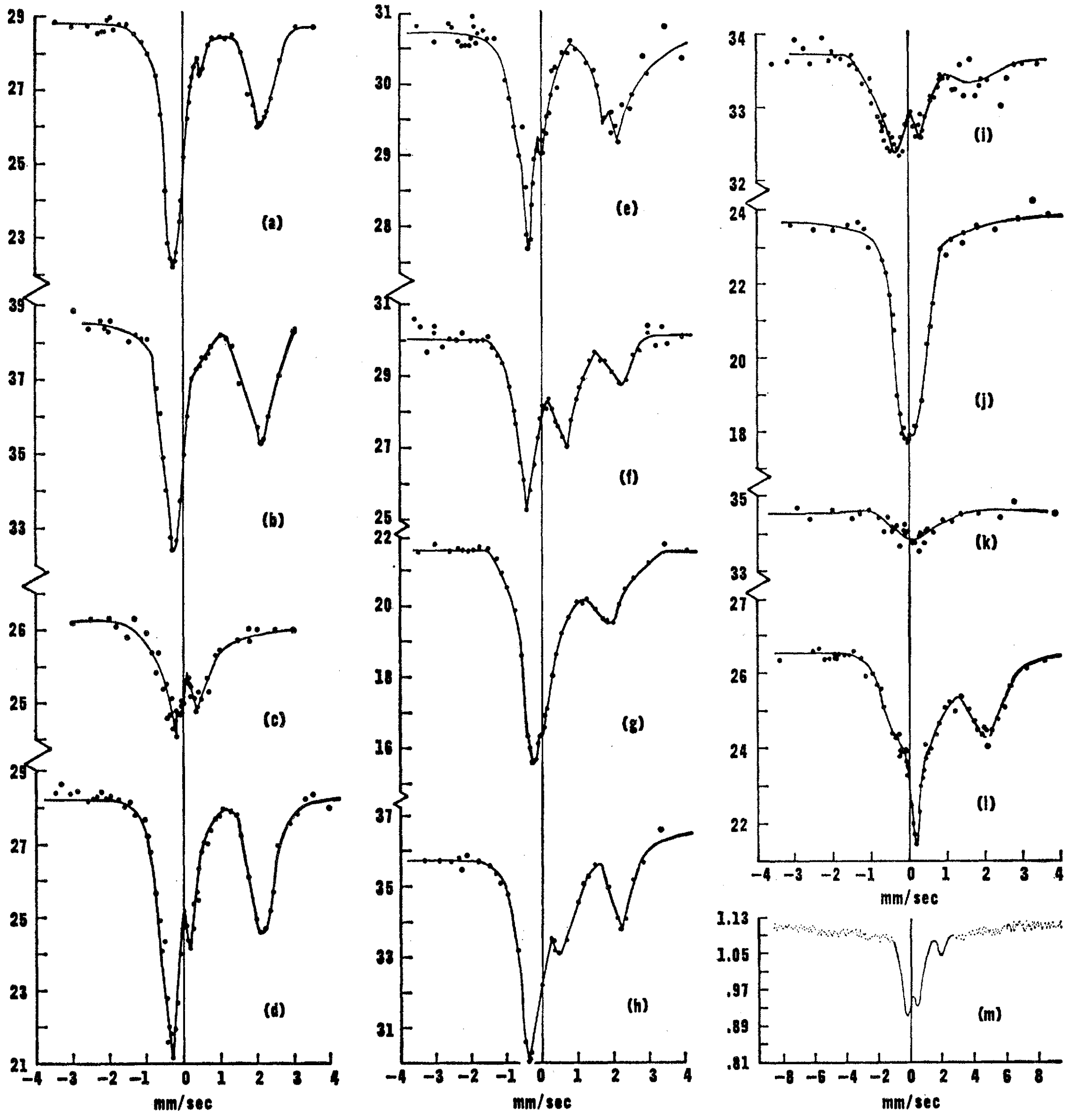
Clay	Oxidation state	Lattice position	Isomer shift*	Quadrupole splitting*
2:1 Dioctahedral				
Nontronite $(Al_{0.48}Fe^{3+}_{1.49}Mg_{0.03})(Si_{3.52}Al_{0.09}Fe^{3+}_{0.39})O_{10}(OH)_2$	Fe^{+3}	Oct.	~0.3	~0.3
Illite $(Al_{1.41}Fe^{3+}_{0.35}Fe^{2+}_{0.11}Mg_{0.19})(Si_{3.47}Al_{0.53})O_{10}(OH)_2$	Fe^{+2} Fe^{+3}	Oct. Oct.	~1.1 0.35	~1.0 0.30
Glauconite $(Al_{0.35}Fe^{3+}_{1.12}Fe^{2+}_{0.19}Mg_{0.40})(Si_{3.58}Al_{0.42})O_{10}(OH)_2$	Fe^{+2} Fe^{+3}	Oct. Oct.	~1.1 0.35	~1.1 0.32
Montmorillonite $(Al_{1.55}Fe^{3+}_{0.19}Fe^{2+}_{0.02}Mg_{0.26})(Si_{3.88}Al_{0.12})O_{10}(OH)_2$	Fe^{+3}	Oct.	0.4	
2:1 Chain				
Xylotile $(Fe^{3+}_{0.60}Fe^{2+}_{0.10}Mn^{2+}_{0.02}Mg_{1.78})(Si_{3.57}Al_{0.06}Fe^{3+}_{0.37})O_{10.7}Ca_{0.02}$	Fe^{+2} Fe^{+3}	Oct. Oct./Tet.	~1.2 ~0.2	~1.0 ~0.4
Attapulgite $(Al_{0.68}Ti_{0.03}Fe^{3+}_{0.19}Fe^{2+}_{0.01}Mg_{0.99})(Si_{3.91}Al_{0.09})O_{10.5}$	Fe^{+3}	Oct.	~0.3	~0.2
1:1 Trioctahedral				
Chamosite $(Al_{0.85}Fe^{3+}_{0.01}Fe^{2+}_{1.84}Mg_{0.22})(Si_{1.33}Al_{0.67})O_5(OH)_4$	Fe^{+2}	Oct.	~1.2	~1.25
Cronstedtite $(Fe^{3+}_{0.71}Fe^{2+}_{2.38})(Si_{1.12}Al_{0.07}Fe^{3+}_{0.41})O_5(OH)_4$	Fe^{+2} Fe^{+3} Fe^{+3}	Oct. Oct. Tet.	~1.3 ~0.4 ~0.05	~1.2 ~0.3 ~0.4
2:1 Trioctahedral				
Biotite, single crystal $K(Fe^{2+}_2Mg)(Si_4)O_{10}(OH)_2$	Fe^{+2} Fe^{+3}	Oct. Oct.	1.15	1.20 ~0.2–0.6
Biotite Powder	Fe^{+2}	Oct.	1.15	1.20
Weathered Biotite	Fe^{+3}	Oct.	0.35	0.30
Ferriannite $K\,Fe^{2+}_3(Si_3Fe^{3+})O_{10}(OH)_2$	Fe^{+2} Fe^{+3}	Oct. Tet.	1.15 ~0.05 or more	1.20 ≤0.40
Griffithite $(Al_{0.04}Fe^{3+}_{0.52}Fe^{2+}_{0.44}Mg_{1.88})(Si_{3.19}Al_{0.81})O_{10}(OH)_2$	Fe^{+2} Fe^{+3}	Oct. Oct.	~1.2 ~0.4	~1.3 ~0.6
1:1 Dioctahedral				
Iron kaolin $(Al_{1.63}Fe^{3+}_{0.29}Mg_{0.08})Si_{2.00}\,O_5(OH)_4$	Fe^{+3}		0.37	0.22
Iron Minerals				
Goethite $Fe\,O\,(OH)$	Fe^{+3}	Oct.	0.37	0.27
Siderite $Fe(CO_3)$	Fe^{+2}		1.25	0.90

* Values are millimeters per second. All isomer shifts are positive.

The spectrum obtained with a weathered biotite (Fig. 1) shows peaks at approximately −0.2 and +0.4 mm/sec, which are presumably due to octahedral Fe^{3+}. Though the biotite showed no evidence of expansion to vermiculite, the lack of any appreciable peak at +2.1 mm/sec indicates most of the iron has been oxidized. Some goethite (−0.1 mm/sec) is also present.

The spectrum (Fig. 1) of a synthetic biotite containing both octahedral and tetrahedral iron (*7*) differs from that of normal biotite in having a well-developed peak at +0.20 mm/sec. This peak is presumably due to tetrahedral Fe^{3+}. The other half of the quadrupole doublet may occur at −0.6 mm/sec. The data suggest a break in slope at −0.6 and the average slope of the left side of the peak is slightly less than that for the biotite with only octahedral Fe^{3+}. If this is true, the isomer shift value is 0.05 mm/sec and the quadrupole split is 0.40, both within the accepted range of values for Fe^{3+}. The isomer shift appears to be less than that for octahedrally coordinated Fe^{3+} but it is not possible to be certain how the quadrupole split compares. Similar differences in isomer shift for tetrahedral and octahedral Fe^{3+} have been reported for rare earth-iron garnets (*8*). The shoulder at +0.5 mm/sec is probably due to the presence of a small amount of octahedral Fe^{3+}.

A sample of chlorite containing octahedrally coordinated Fe^{2+} in both the talc and brucite sheets (*9*) was studied. X-ray analysis showed that a minor amount of siderite was present. The peaks at +0.1 and +1.8 mm/sec (Fig. 1) are likely due to the siderite. The peaks at −0.3 and +2.2 mm/sec are similar to those for biotite. The relative narrowness of the strong adsorption peak would suggest that these two layers (one containing predominantly hydroxyl ions and the other predominantly oxygen ions) are not sufficiently different to cause appreciable differences in isomer shift and quadrupole splitting values; however, there is the suggestion of shoulders on the left side of the −0.3 mm/sec peak and right side of the +2 mm/sec peak which may be due to the brucite iron. If this



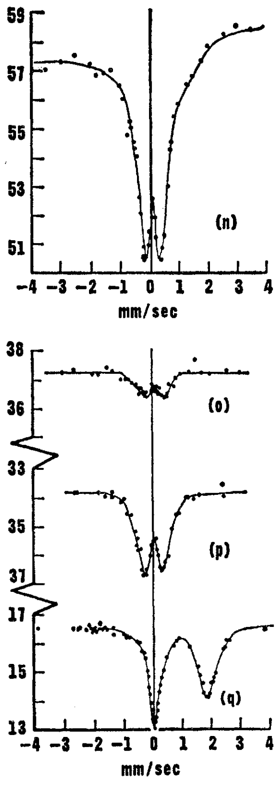


Fig. 1 (first 3 columns). Mössbauer spectra. Intensity scale is in thousands of counts per minute for all samples except glauconite. (The glauconite spectrum was made at Oak Ridge National Laboratory, by L. D. Roberts and D. O. Patterson.) *a*, Biotite flake; *b*, ground biotite; *c*, weathered biotite; *d*, ferrianite; *e*, chlorite; *f*, griffithite; *g*, cronstedite; *h*, chamosite; *i*, illite; *j*, nontronite; *k*, montmorillonite; *l*, xylotile; *m*, glauconite. Fig. 2 (4th column). Mössbauer spectra. *n*, Iron-rich kaolinite; *o*, Fe-kaolinite with FeO (OH) extracted; *p*, FeO (OH) from weathered diabase [a similar spectrum was obtained for FeO (OH) from an iron-stained kaolin]; *q*, siderite.

is so, then the splitting value for brucite Fe^{2+} is larger than for talc Fe^{2+}, which suggests that the talc octahedra are more distorted than the brucite octahedra (*10*).

The spectrum for griffithite, an iron-rich trioctahedral clay, is similar to biotite but shows the effects of a relatively high content of Fe^{3+} (7.32 percent Fe_2O_3, 7.83 percent FeO). The broadness of the peaks at +0.75 (Fe^{3+}) and +2.30 (Fe^{2+}) mm/sec is presumably due to the presence of two quadrupole doublets for each type of iron. The asymmetry of these two peaks suggests that the paired doublets are not of equal intensity and therefore the two types of iron are not shared equally between the two different octahedral sites. The splitting values for both Fe^{3+} and Fe^{2+} appear to be larger than for biotite. This indicates that Fe^{3+} is concentrated in the less symmetrical sites and Fe^{2+} is concentrated in the more symmetrical sites. The octahedral sites are presumably similar to those in biotite. (If the low velocity portion of each double peak were more equal in intensity to the high velocity portion, the resulting average peaks would have values similar to biotite. The overlapping peaks appear to be separated by 0.4 to 0.5 mm/sec.)

The chamosite (1 : 1 trioctahedral) spectrum (Fig. 1) is similar to that of the 2 : 1 trioctahedral clays. The Fe^{3+} is more abundant than indicated by the chemical analyses (Fe_2O_3, 0.2 percent; FeO, 40 percent), and x-ray analyses indicate some iron oxides are present in the sample. The FeO(OH) spectrum (Fig. 2) has a quadrupole doublet with peaks at −0.15 and +0.4 mm/sec Thus most of the peak at +0.5 mm/sec and the gentler slope of the right side of the major peak is presumably attributable to iron oxides.

Cronstedtite is 1 : 1 trioctahedral clay containing approximately 30 percent Fe_2O_3 and 42 percent FeO. Some of the Fe^{3+} is believed to be present in the tetrahedral layer. The spectrum has a major peak at −0.2 mm/sec and a smaller one at approximately +2.0 mm/sec with a distinct shoulder on the high velocity side (Fig. 1). Both peaks have slightly smaller velocity values than the other trioctahedral sheet minerals. The presence of some siderite (Fig. 2) in the sample may account for this; however, x-ray analysis indicated only trace amounts. The relatively shallow slope of the left side of the major peak plus the breaks in slope indicate the presence of several peaks. A tentative interpretation would be: siderite quadrupole doublet due to Fe^{2+} at +0.1 and +1.9 mm/sec; octahedral Fe^{2+} doublet at approximately −0.2 and +2.3 mm/sec; octahedral Fe^{3+} doublet at −0.2 and +0.5 mm/sec; tetrahedral Fe^{3+} doublet at −0.6 and +0.2 mm/sec.

The 2 : 1 dioctahedral clays, illite and glauconite, have the expected quadrupole doublet for octahedral Fe^{3+} and a doublet of lesser intensity from the small amount of octahedral Fe^{2+} (Fig. 1). The glauconite sample contains some iron oxides which contribute to the Fe^{3+} quadrupole doublet. The isomer shift and quadrupole splitting values for both Fe^{3+} and Fe^{2+} tend to be smaller for the dioctahedral clays than the trioctahedral. In part this may be due to the tendency for the octahedral cations in dioctahedral clays to fill the two like sites. This would decrease the range of asymmetry and perhaps produce only a single quadrupole doublet for each ion, rather than two as suggested for the trioctahedral clays.

Montmorillonite is an extremely fine-grained, poorly crystallized clay mineral. The sample analyzed contained approximately 4 percent iron oxides. A spectrum of the less than 0.2 μ fraction (Fig. 1) indicates that the octahedral Fe^{3+} produces a broad peak with an apex at approximately +0.2 mm/sec. This peak is appreciably broader than any other found in this study. The octahedral sites are so variable in their makeup that a series of overlapping quadrupole doublets are presumably produced.

The spectrum for nontronite (31 percent Fe_2O_3, 0.4 percent FeO) has one relatively broad peak with an apex at +0.1 mm/sec. There has been some question whether all of the Fe^{3+} was in the octahedral layer or shared between the octahedral and tetrahedral layers (*11*). In the absence of iron oxides (not detected by x-ray) the quadrupole doublet for octahedral Fe^{3+} should be resolved. The lack of resolution suggests that a doublet for tetrahedral Fe^{3+} may be present. Isomer shift and quadrupole splitting values cannot be determined but must be similar to those for the other dioctahedral clay minerals.

The sample of xylotile, an iron-rich (17 percent Fe_2O_3, 1.5 percent FeO) sepiolite (chain structure), contained a considerable amount of siderite which is shown by the Mössbauer spectrum (Fig. 1). The relatively sharp peak at +0.10 mm/sec and much of the broad peak at about +2.0 mm/sec are due to siderite. A portion of the peak at +2.0 mm/sec is presumably due to the presence of some octahedral Fe^{2+} in the clay lattice, causing a shift of the peak maximum to the right of +1.9 mm/sec. Unfortunately, it was not possible to remove the siderite without destroying the xylotile. The broad peak with an apex near zero velocity appears to be similar to that for nontronite and indicates that the isomer shift and quadrupole splitting values of this chain-structure clay are similar to those for the sheet structure clays. The shoulder at −0.40 mm/sec is probably due to the octahedral Fe^{3+}.

The other chain-structure clay, attapulgite (3.4 percent Fe_2O_3, 0.2 percent FeO), has a low iron content, but a spectrum was obtained. No impurities were detected by x-ray analysis and all the iron is presumed to be in the octahedral layer. The spectrum has a broad peak near zero velocity which is presumably due to unresolved Fe^{3+} quadrupole doublets with low splitting values.

Samples of goethite FeO(OH) were obtained from a weathered diabase and an iron-stained kaolin. The x-ray patterns showed only broad, weak peaks at 4.18 Å. As goethite has a layer lattice in which the Fe^{3+} is in octahedral coordination, it might be expected that the spectra would be similar to those of the sheet silicates. The spectra for both samples were similar (Fig. 2), showing a simple quadrupole doublet with a splitting value of 0.27 and an isomer shift of +0.37 mm/sec. The splitting value is generally less than that for the octahedral Fe^{3+} in the clay minerals and the shift value is similar. The doublet peaks are sharp, indicating that the octahedral environment is relatively uniform. The weakness of the x-ray reflections are presumably due to fine grain size rather than poor structural order. The similarity of the goethite spectra to those of the sheet silicates with octahedral Fe^{3+} will make it extremely difficult to separate the two effects.

A sample of iron-rich kaolin gave a simple spectrum (Fig. 2) identical to that for goethite. Kunze and Bradley (*12*) reported 8.25 percent Fe_2O_3 in the sample but thought that most of the iron was present in the octahedral layer as Fe^{2+}. The Mössbauer spectrum indicated that no appreciable Fe^{2+} is present.

When the sample was treated to dissolve "free" iron (*13*), nearly all the iron was removed. The spectrum of the product was similar to that of the original sample, but showed considerably less absorption than the original sample ($<0.2\ \mu$ size fraction). X-ray analysis indicated kaolin was still present. Either the iron-rich kaolin is a mixture of kaolin and very fine geothite or it is an Fe^{3+} kaolin which has a Mössbauer spectrum identical to goethite (which might be expected) and from which the octahedral iron can be removed by a method designed to remove "free" iron oxides.

The spectrum of siderite (FeCO$_3$) is a quadrupole doublet typical of the Fe^{2+} ion (Fig. 2). This spectrum exhibits peak broadening as a result of (i) nonuniformity of motion from the constant velocity drive, and (ii) absorption in a sample of appreciable thickness (*14*). The area of the absorption peak at +1.90 mm/sec equals that of the peak at +0.10 mm/sec, but the former is considerably more affected by non-uniformity in velocity. Both peaks are somewhat broadened because this sample was overly thick. A reanalysis of a thinner sample of siderite showed narrower peaks, but the position of the peaks is unchanged.

It should be noted that all of our peaks at about +2.0 mm/sec are somewhat broadened because of non-uniform velocity. For those clay samples of high iron content there is probably slight broadening because of absorption effects, but in no case did the clay samples have a thickness of iron as great as that of the siderite sample.

In summary, Mössbauer absorption spectrography should prove to be of considerable value for studying the sheet-structure silicates. The oxidation state of iron can easily be determined. It appears to be possible to distinguish between octahedrally and tetrahedrally coordinated iron. Minor variations in the spectra of the various minerals should prove to be significant and afford information on variations in the character of the octahedral layer.

Mössbauer absorption can be detected for iron present in amounts as low as 1 percent Fe$_2$O$_3$, but a higher concentration is required if useful spectra are to be obtained. Siderite as an impurity can usually be identified, but goethite will cause trouble. Conversely, goethite can be detected more reliably with Mössbauer equipment than with x-ray.

By increasing the count to achieve better statistics, and by minimizing the line-broadening effects noted above, it should be possible to analyze the fine structure of the complex peaks near zero velocity to supply much more information than the present preliminary study suggests.

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Alga-Like Fossils from the Early Precambrian of South Africa

Abstract. *Micropaleontological studies of carbonaceous chert from the Fig Tree Series of South Africa (> 3.1 × 10⁹ years old) revealed the presence of spheroidal microfossils, here designated* Archaeosphaeroides barbertonensis, *interpreted as probably representing the remnants of unicellular alga-like organisms. The presumed photosynthetic nature of these primitive microorganisms seems corroborated by organic geochemical and carbon isotopic studies of the Fig Tree organic matter, and is consistent with the geologically and mineralogically indicated Early Precambrian environment. These alga-like spheroids, together with a bacterium-like organism previously described from the Fig Tree chert, are the oldest fossil organisms now known.*

Laminated stromatolitic structures, comparable in gross morphology to the biohermal deposits of modern blue-green and red algae, have long been known from sediments of Precambrian age. Although generally devoid of cellularly preserved microfossils, these structures have been regarded as presumptive evidence of early algal activity. Firm evidence for this supposition comes from the recent investigations of siliceous stromatolites of Middle Precambrian age from the Gunflint Iron Formation (1.9×10^9 years old) along the northern shore of Lake Superior in Ontario, Canada. Various microorganisms, some morphologically similar to modern blue-green algae, are structurally preserved in the organic lamellae of these black chert stromatolites (*1, 2*). Cellular microfossils are similarly preserved in laminated, primary black cherts associated with stromatolites of the genus *Collenia* in the Late Precambrian Bitter Springs Formation (about 1.0×10^9 years old) of central Australia (*3*). Morphologically, several of these billion-year-old organisms are referable to modern families of blue-green and green algae. These two microfossil assemblages and the many deposits of calcareous stromatolites and carbonaceous sediments of comparable age establish that diverse types of primitive algae were present relatively early in geologic time.

The oldest known stromatolites are apparently those described and figured by A. M. Macgregor (*4*) from Early Precambrian limestones near Turk Mine, 53 km north-northeast of Bulawayo, Rhodesia. Columnar, dentate, and domical structures have been noted in the deposit; the domical forms have been compared with *Collenia* (*4*). Comparison of the stable isotopes C^{12} and C^{13} in reduced organic carbon, present as finely bedded graphitic laminations, and in oxidized inorganic carbon, present in the carbonate matrix, demonstrated an isotopic fractionation probably of photosynthetic origin (*5*). Microfossils have not been reported in the calcareous Bulawayan stromatolites, but the gross morphology, carbon isotopic composition, and presence of organic matter are consistent with an algal