

4(b). Mats on Sandy Substrates - Diagenetic Features Related to Mat Metabolism and Decay

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Microbes can influence mineral precipitation in a mat by (a) changing the pore water chemistry in a mat via metabolic activity, such as photosynthesis, or (b) by nucleation of minerals on their cell walls and (c) within the extra cellular polysaccharides (EPS) that surround them. Given the right water chemistry, syngenetic carbonate precipitation due to photosynthesis (a) is considered a viable process (e.g., Thompson and Ferris, 1997; Pratt, 2001) and can occur in terrigenous clastic environments just as well as in carbonate dominated settings. Evidence for carbonate precipitation on cell walls (b) has been presented by a variety of authors (e.g. Krumbein, 1974; Gerdes and Krumbein, 1987; Chafetz, 1994), and EPS mediated microbial calcium carbonate precipitation (c) was for example reported by Arp et al. (2001) from biofilms in an alkaline lake. Cell wall nucleation of calcium carbonate by sulfate reducing bacteria has been examined by Van Lith et al. (2003).

Krumbein and Cohen (1977), in a study of modern cyanobacterial mats, concluded that not metabolic CO₂ uptake, but rather biogeochemical reactions during bacterial degradation of the mat were responsible for carbonate (calcite, aragonite) precipitation. The carbonates mimicked filament structures and formed onkoidal structures of 1-2 mm diameter, yet less than 1% of the modern mat volume was transformed into preservable carbonate matter (Cohen et al., 1977). Above observations on carbonate formation during decay in cyanobacterial mats were confirmed in a study by Gerdes and Krumbein (1987), that found micrite accumulation along cyanobacterial sheaths, micrite encapsulation of coccoidal cells, as well as formation of concentrically coated carbonate ooids. The latter authors also documented that with increasing burial depth and carbonate precipitation, original structures (filaments, coccoids) become increasingly unrecognizable. Thus, whereas it is difficult to determine with certainty whether a given carbonate grain in a modern mat is due to photosynthesis induced precipitation, there is good observational evidence that decay of microbial filaments, cells, and EPS fosters abundant carbonate precipitation.

Examples of ancient microbial mat deposits in siliciclastics that show evidence for early lithification through mat associated carbonate precipitation include examples from the Proterozoic of the Belt Basin (Chapter 7b), the Cambrian of Israel (Soudry and Weissbrod, 1995) and Minnesota/USA (Davis, 1968), the Carboniferous of North Africa (Bertrand-Sarfati, 1994), the Permian of New Mexico/USA (Harwood, 1990), and the Miocene of Spain (Martin et al., 1993; Braga and Martin, 2000). The common thread for all these deposits is the alternation of laminae that are enriched in carbonate minerals (usually calcite or dolomite) with laminae that consist largely of siliciclastic grains. Carbonate bearing laminae may be largely carbonate with small additions of siliciclastic grains, but may also contain in excess of 50% siliciclastics. Textural features in carbonate-rich laminae include fenestrae (Fig. 1B), calcified filaments, micritic laminae (Fig. 1A, B), peloidal and clotted micrite (Fig. 1A, C), siliciclastic grains with thin oolitic coatings, and “floating” siliciclastic grains within micrite (Fig. 1A, C, D). In most of

these examples, early lithification via carbonate precipitation allowed for the accretion and preservation of dome-shaped and laminated microbial buildups (see Chapter 7b).

In modern day mats, there is typically not enough calcite precipitation to preserve original mats structures via early lithification (Cohen et al., 1977). Thus, judging from the comparatively low number of reported sandy microbial mat deposits with carbonate enriched laminae in the rock record, conditions conducive to this type of mat preservation may only have occurred intermittently through geologic time. A possible control may have been the interplay between dissolved CO₂ and calcium concentrations in basin waters (Arp et al., 2001).

Probably at least since the Mesoproterozoic (Johnston et al., 2005), many shallow marine microbial mats likely accumulated beneath oxygen bearing seawater. Considering the interconnected pore spaces and high permeability of sandy sediments, one would expect to find minerals that require reducing conditions for precipitation (e.g. pyrite, ferroan carbonates) at some depth below the sediment-water interface. Under these conditions, lateral variability of permeability and irregular distribution of organic matter would probably preclude stratiform distribution of these “reducing” minerals.

The presence of a microbial mat at the surface, however, makes a significant difference in this regard. Microbial mats typically define a sharp geochemical boundary. In photosynthetic cyanobacterial mats for example, the very surface layer of the mat may produce oxygen. Directly below this layer, however, the region where anaerobic bacteria degrade original mat material, the sediment tends to be strongly reducing (Bauld, 1981; Gerdes et al., 1985). In such a geochemical environment conditions are favorable for the precipitation of “anoxic” minerals, such as pyrite, siderite (Fig. 2B, D), ankerite, and ferroan dolomite, in addition to calcium carbonate. The lowering of pore water sulfate concentrations by sulfate reducing bacteria also promotes the precipitation of early diagenetic dolomite (Baker and Kastner, 1981).

The comparatively high permeability of sandstones allows for microbial metabolization of most co-deposited organic matter very early in burial history, and microbially mediated reactions between buried organic matter and migrating formation waters most likely will remove the remainder. Thus, in sandstones, carbonaceous horizons due to microbial mat colonization most likely will not make it into the rock record as such. Aside of morphological features in the sandstone host, such as domal buildups (see Chapter 7b), cementation of sandstones by above “anoxic” minerals may be the only remaining indicator of the former presence of microbial mats (Fig. 2A). Just as the bulk of calcium carbonate deposition in mats appears to be largely due to decay processes (Krumbein and Cohen, 1977; Cohen et al., 1977; Gerdes and Krumbein, 1987), these cements of “reducing” minerals are considered “mat decay mineralization”. Mat decay mineralization produces thin, well defined, and laterally extensive horizons of calcium carbonate and/or “reducing” minerals, such as ferroan carbonates or pyrite. Conversely, if a sandstone contains instead disseminated carbonate, ferroan carbonate, or pyrite cement, a mat origin is unlikely.

Photomicrographs in Fig. 2 illustrate mat decay mineralization features from the Revett Formation, Mesoproterozoic Belt Supergroup of Montana, USA (Schieber, 1998, 1999). These sandstones formed in a fluviodeltaic setting (Boyce, 1973; Bowden, 1977; Mumma et al., 1982; White et al., 1984), and thus the pore waters contained in the original deposits were either freshwater or brackish water. This implies, in comparison to

seawater, a shortage of dissolved sulfate and consequently low concentrations of sulfide in the anoxic sediments beneath mats. This is probably the underlying reason for the preponderance of ferroan carbonates in microbial mat layers of the Revett Formation. Garlick (1988) was the first to realize that the ferroan minerals in Revett sandstones were probably the diagenetic imprint of former microbial mats. Had these sandstones been deposited in a setting dominated by marine waters, cementation by pyrite might have occurred instead of ferroan carbonates.

If minerals related to mat decay were originally deposited in a mat with abundant organic matter and EPS, siliciclastic grains were probably originally surrounded by an organic matrix. Once this organic matrix is gradually replaced by decay related minerals (carbonates, ferroan carbonates, pyrite), “floating” quartz and mica grains are to be expected (Fig. 2E).

If the mat simply gets buried by more sand, stratiform layers of decay related minerals may result (Fig. 2A, B, F). If erosion of mat bound sand occurs, fragments may be reburied and then undergo reducing diagenesis because of their contained organic matter (Fig. 2C). Desiccated eroded mat fragments may produce curved features once buried in more sand. On weathering these curved features may be marked by carbonate cement and even iron stains from “reducing” cement minerals (see Chapter 3, Chapter 7b). If the mat is still soft when erosion occurs it may produce buried contorted fragments that again can be seen due to their cementation and/or iron stain (see Chapter 3, Chapter 7b).

Chapter 4b: Figures and Captions

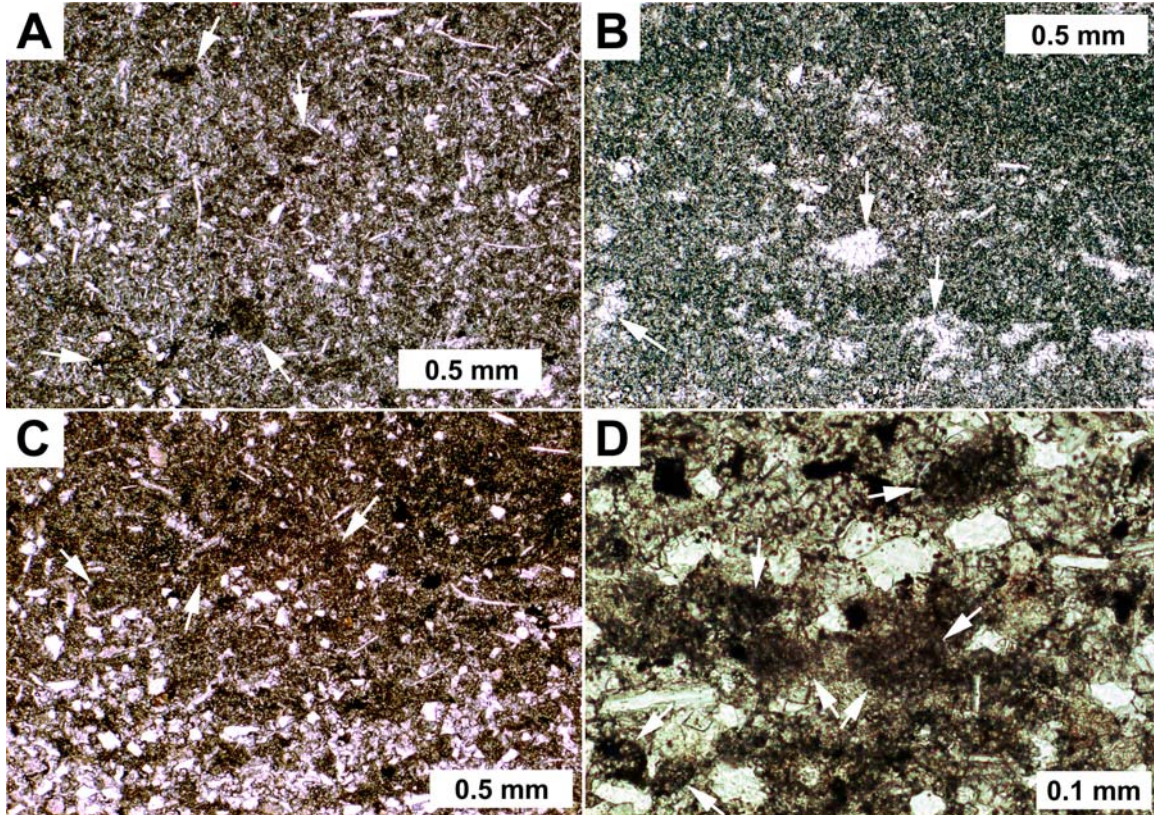


Figure 1: Photomicrographs of calcite and/or dolomite rich laminae in sandy microbial mat deposits from the Revett Formation, Belt Supergroup, Middle Proterozoic of Montana. (A) fine crystalline dolomitic lamina (originally micrite?) with scattered “floating” grains of quartz and mica. Note relict peloidal texture (arrows point to peloids). (B) dolomitic lamina with fenestrae (arrows) that are filled with dolomite spar. (C) dolomite-rich lamina with peloids in a (originally micritic?) fine crystalline dolomite matrix. Contains a higher proportion of terrigenous clastic grains. (D) peloidal texture in dolomite-rich lamina of sandy microbial mat deposit. Arrows point to clearly visible peloids.

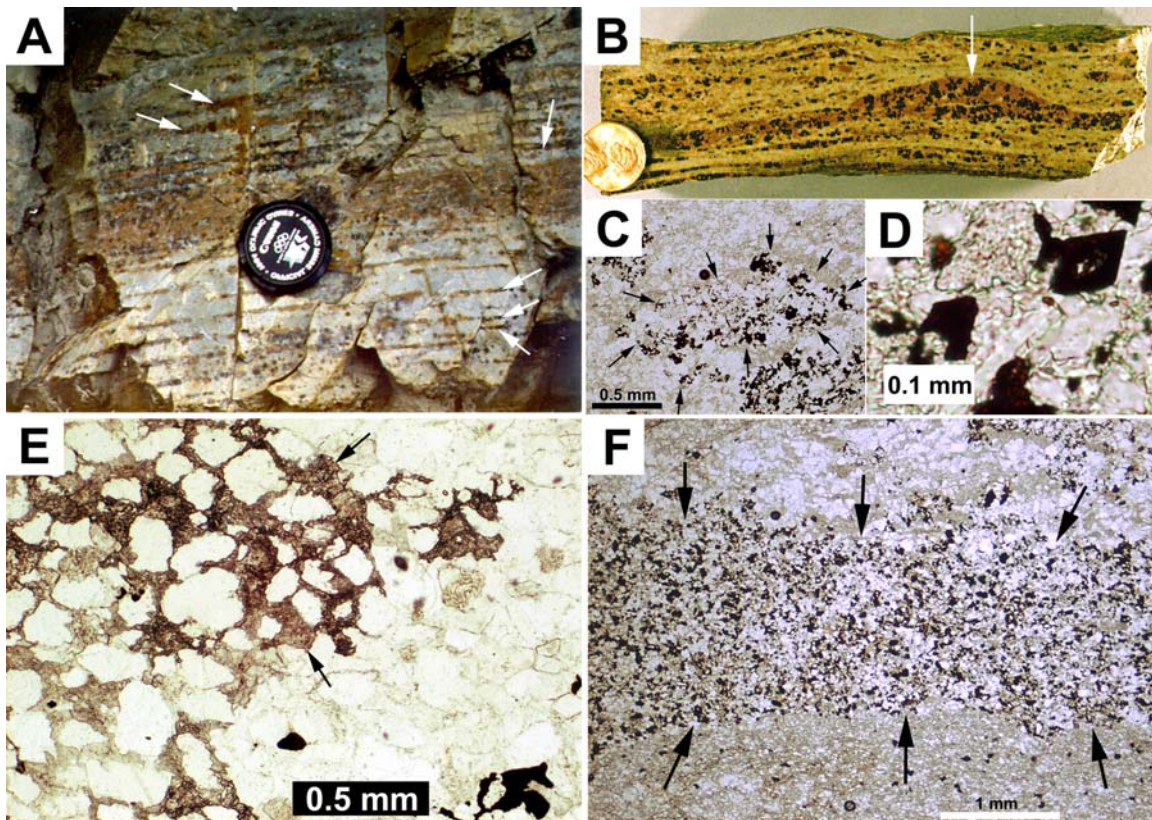


Figure 2: Iron carbonate cements in microbial mat deposits from the Revett Quartzite, Belt Supergroup, Middle Proterozoic of Montana. (A) outcrop photo that shows bedding parallel brownish horizons (arrows) that are interpreted as the location of former microbial mats. Lens cap is 45 mm in diameter. (B) Hand specimen of Revett Formation sandstone with iron stained (limonitic) laminae. The arrow points to a layer with asymmetric domal thickening, possibly an incipient stromatolite. The very dark patches are oxidized patches of siderite cement. The brownish-orange colored laminae mark the presence of ankerite and ferroan dolomite. (C) Oval patch of iron-stained sandstone (marked with arrows) within a matrix of non-stained sandstone. Detailed comparison of the matrix material to the detrital components of the patch shows that the matrix is finer grained and has a substantially larger content of clays and shale clasts. This observation suggests that the patch outlined with arrows was an eroded fragment of a prior sediment. Because sand itself is not cohesive, the sand grains might have been held together by microbial filaments during transport. (D) oxidized siderite rhombs in sandstone. (E) sandstone with patchy siderite cement (dark-brownish, upper left, arrows). Sandstone at lower right is quartz cemented. Note floating grains in siderite patch. (F) Photomicrograph of an iron stained lamina as seen in B. The black arrows point to the margins of the lamina, and the scattered opaque spots are largely iron oxides (variably hydrated) that resulted from oxidation of iron-bearing interstitial carbonates. Note that the under and overlying layers contain comparatively small amounts of iron oxides.

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