Chapter 3
Terrestrial Ecosystems in the Precambrian

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3.1   Introduction

Three lines of evidence for ancient life on land are paleosols, microfossils, and MISS (microbially induced sedimentary structures). The three appear in the rock record as old as 3.5–2.7 Ga (Hallbauer and van Warmelo 1974; Rye and Holland 2000; Wacey et al. 2008; Johnson et al. 2010; Noffke et al. 2013). The primitive terrestrial biosphere is plausibly as old as the oldest marine biosphere (Schopf 1983; Schopf and Klein 1992), with both aquatic and terrestrial realms teeming with life 3.5 Ga ago. This result is unsurprising considering the modern ecological capabilities of microbes, which are comparable with fossil analogs and have profound effects in sediments and soils (Belnap and Lange 2001; Navarro-González et al. 2003; Beraldi-Campesi and Garcia-Pichel 2011; Retallack 2012). Functional aquatic and terrestrial ecosystems may be even older than 3.5 Ga, but undisputable biosignatures before that time are ambiguous (Nutman et al. 2010; Papineau et al. 2010), and the timing when life first populated the Earth is still unresolved.

Direct evidence for ancient terrestrial communities comes from the fossil record, and the several limitations of such should be noted. First, it has many temporal gaps, which makes it difficult to appreciate the continuous history of life on Earth (Sadler 1981). Second, not all the rocks have been studied in detail, especially biosignatures preserved in it, so that recent and future discoveries may yield more tangible information about the history of the biosphere (Dutkiewicz et al. 2006). Third, microbial biological diversity of the Precambrian fossil record is

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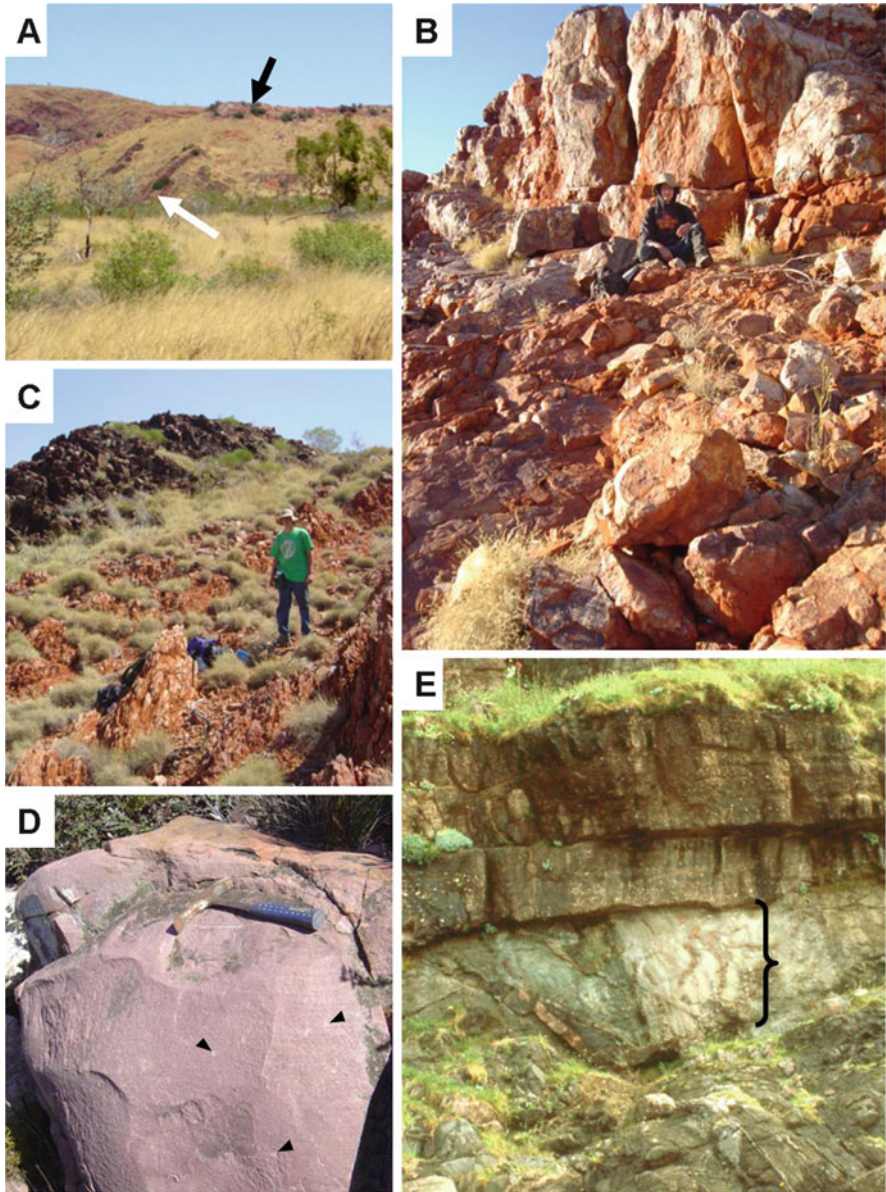Geological Sciences, University of Oregon, Eugene, OR 97403-1272, USA

fundamentally different and more challenging to study than Phanerozoic megafossils (Schopf et al. 2007), which unduly bias orthodox interpretations about the evolution of the biosphere.

3.2 Paleosols

The oldest geological unconformity in a sedimentary sequence, which is metamorphically recrystallized in a way that its sedimentary history is almost indecipherable, is ~3.4 Ga old (Fig. 3.1a; Buick et al. 1995). This surface beneath the Strelley Pool Formation has thick (4 m) clayey paleosols (Fig. 3.1b–d), which are deeply weathered, aluminous, and metamorphosed to apophyllite facies, implying high fluid fluxes, secondary mineralization, and an overall alteration after burial (Buick et al. 1995; Hoffman 1995; Ohmoto et al. 2007; Johnson et al. 2009, 2010). Such thick clayey paleosols are very common on basement unconformities of the Archean and Proterozoic record (Table 3.1). Well-developed paleosols as the Neoproterozoic Sheigra paleosol (Fig. 3.1e) are direct records of ancient landscapes, where microbial contribution can be sought. The development of relatively thick paleosols implies stable intracratonic settings where weathering and soil formation prevail and sediment transport is minimal. What stabilized these paleosols to allow such deep chemical weathering could have been life itself, considering evidence of elemental depletions, isotopic composition of carbon, and fossilized structures (Retallack 2001).

The master reaction for weathering in soils is hydrolysis, the incongruent dissolution of feldspars, or other minerals to form clays, by dilute solutions of carbonic acid from dissolution of soil CO_2 in pore waters (Retallack 2001). Such reactions have been demonstrated in many Precambrian paleosols, but could theoretically be created abiotically. However, life has accelerated weathering processes over geological time, and biotic effects are more obvious than abiotic effects. A series of experiments by Neaman et al. (2005) showed that organic ligands are necessary for any significant depletion of phosphorus in modern soils, and such depletions have been demonstrated in paleosols as old as 2.7 Ga (Driese et al. 2011), as well as for geologically younger paleosols (Retallack and Mindszenty 1994; Retallack et al. 2013a, b).

Carbon isotopic composition of biologically produced organic matter is depleted in ^{13}C ($\delta^{13}C_{org} = -10‰$ to $-30‰$), and soil methanogens create extreme depletion ($\delta^{13}C_{org} = -30‰$ to $> -90‰$; Schidlowski 2001). Such low values have been reported from Precambrian paleosols as old as 2.7 Ga (Rye and Holland 2000), as well as for geologically younger paleosols (Retallack and Mindszenty 1994; Watanabe et al. 2000; Retallack et al. 2013a, b). Ironically, the carbon content of Precambrian paleosols is very low ($<<1$ wt %), about the same as for Phanerozoic paleosols (Tomescu et al. 2009). While low-carbon isotopic compositions reveal microbial primary producers in soils, low-carbon abundance also implies an active decomposition of organics by microbes on the land, which may imply well-

Fig. 3.1 Examples of Precambrian paleosols: (**a**) One of the oldest paleosols at the 3.42 Ga geological unconformity between the tabletop basalt (basalts dipping to left; *white arrow*) and Strelley Pool Formation (chert of ridge top; *black arrow*) 4 miles east of Strelley Pool, Pilbara region, Western Australia (Buick et al 1995): (**b**) Another section of the 3.42 Ga paleosol on Double Bar Formation basalts below basal cherty sandstones of the Strelley Pool Formation, Pilbara region, Western Australia (Van Kranendonk 2000; Altinok 2006): (**c**) Another section of the 3.42 Ga paleosol on Panorama Formation showing felsic tuffs that record ancient volcanism on Steer Ridge, Pilbara region, Western Australia (van Kranendonk 2000; Johnson et al. 2010): (**d**) A younger 1.8 Ga paleosol with elephant-skin texture (A horizon) above gypsum nodules (*arrowheads*) in Stirling Range Sandstone, Barnett Peak, Western Australia (Bengtson et al. 2007):

developed and complex communities, not only of photosynthesizers but hetero-trophs as well.

The most common megascopic traces in Precambrian paleosols are irregular subvertical and macroscopic tubular structures, either picked out by drab haloes in red beds (Fig. 3.2a) or preferentially cemented compared with their matrix (Fig. 3.2b, c). These features are formed through pedogenic processes and may be biotically or abiotically mediated. As these macrostructures seem to be embedded within the paleosol profile, it is unlikely that photosynthetic microbes were involved in their formation. Nevertheless, similar alteration features have been observed in modern biocrusts (Retallack 2011). Fungi can also be envisioned as potential candidates for their formation. Examples include the honey mushroom, *Armillaria mellea* (Mihail and Bruhn 2005), or lichen rhizines like those of *Toninia sedifolia* (Poelt and Baumgärtner 1964). The drab-haloed tubular structures are a fossil form called *Prasinema gracile*, best known from Cambrian paleosols (Retallack 2011). The distinctive gray-green haloes of the fine tubules, preserved by burial gleization, are also comparable with drab-haloed root traces in paleosols (Retallack 2001). These tubular structures (Fig. 3.2b) also found analogs among calcareous and siliceous rhizoconcretions in paleosols (Esteban and Klappa 1983). However, unlike root traces in Phanerozoic paleosols, tubular structures in Precambrian paleosols are much smaller and lack several orders of branching and tapering downward.

Paleosols are important pieces of the record of continental evolution because they are widely distributed on all present continents and are represented in the rock record since 3.5 Ga. Most of them contain indirect evidence of biological weathering, biomass accumulation, and potential fossils and ichnofossils (Retallack 2001). Recognizing pedogenic processes is essential for better understanding depositional paleoenvironments and distinguishing those of marine from nonmarine origin. For instance, pedogenic processes recognized in association with Ediacaran fossils, such as *Dickinsonia costata* (Retallack 2013; Fig. 3.2d). Together with the evidence cited below, the long history of paleosols and their likely biotic features attest for well-established, nonmarine, and widespread terrestrial ecosystems by the Paleoarchean.
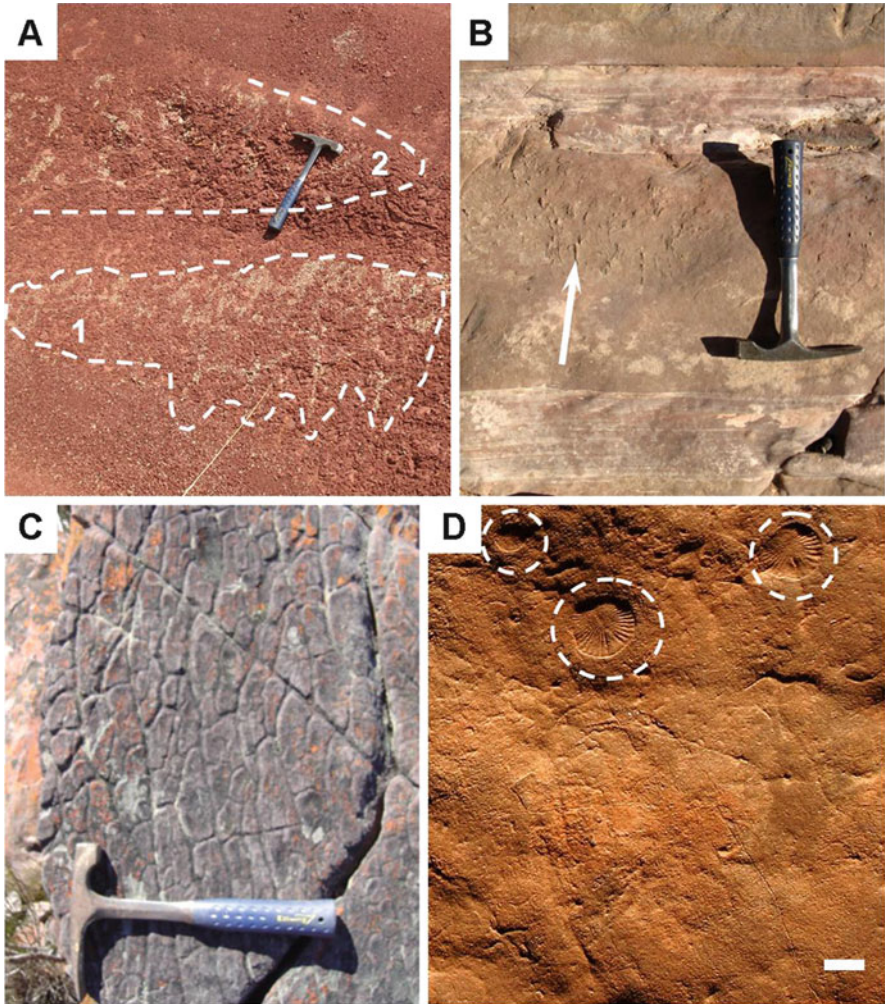
3.3   Microfossils

The oldest reported biosignatures occurring in shallow marine to nonmarine strata may be simple "ambient inclusion trails" from basal sandstones of the Strelley Pool Formation in Western Australia (Lowe 1983; Wacey et al. 2008). Higher within the

Fig. 3.1 (continued) (**e**) Cross section of the 1 Ga Sheigra paleosol on Lewisian gneiss (*bottom*) and Staca paleosol on amphibolite (*bracket*) beneath the Applecross Formation (fluvial facies) of the Torridonian Supergroup, west of Sheigra, northwest Scotland (Retallack and Mindszenty 1994)

Table 3.1 Examples of Archean and Proterozoic paleosols

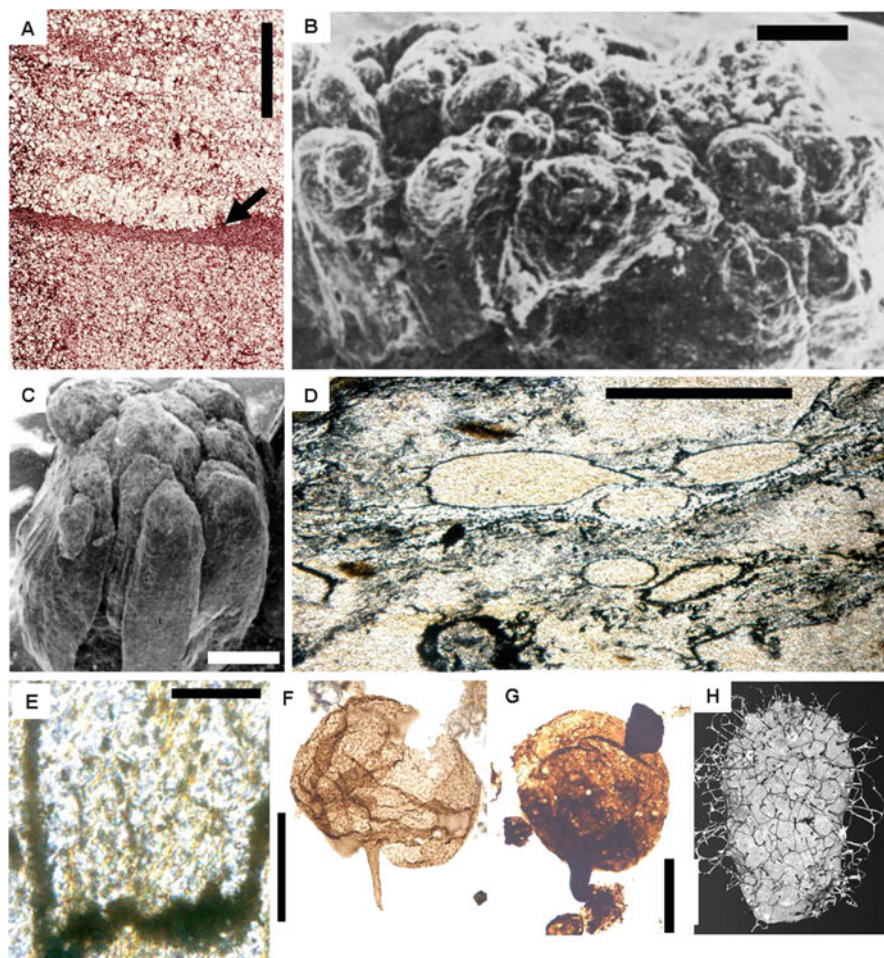| References	Province	Country	Age
Johnson et al. (2010)	Warrawoona Group	Australia	3.4 Ga
Kimberley and Grandstaff (1986)	Dominion Reef Conglomerate	South Africa	3.1–2.8 Ga
Kimberley and Grandstaff (1986)	Pongola	South Africa	3.0–2.9 Ga
Grandstaff et al. (1986)	Dominion and Pongola Supergroups	South Africa	3.0–2.8 Ga
Reimer (1986)	Kaapvaal Craton	South Africa	3.0–2.5 Ga
Hallbauer and van Warmelo (1974)	Witwatersrand and Ventersdorp Supergroups	South Africa	2.9–2.7 Ga
Macfarlane et al. (1994)	Fortescue Group Mt Roe 1 and 2	Australia	2.7–2.6 Ga
Kimberley and Grandstaff (1986)	Timiskaming Group	Canada	2.7 Ga
Rye and Holland (2000)	Mt Roe 1 and 2	Australia	2.7 Ga
Martini (1994)	Paleosol on ultramafics in the eastern Transvaal	South Africa	2.6 Ga
Watanabe et al. (2000)	Transvaal Sequence	South Africa	2.6 Ga
Pandit et al. (2008)	Tulsi Namla section	India	2.5–2.1 Ga
Aspler and Donaldson (1986)	Nonacho Basin	Canada	2.5–2.0 Ga
Banerjee (1996)	Paleosols BGC–Aravalli boundary	India	2.5–1.9 Ga
Bandopadhyay et al. (2010)	Singhbhum–Orissa craton	India	2.5 Ga
Prasad and Roscoe (1996)	Lower and upper sub-Huronian paleosols	Canada	2.4–2.2 Ga
Kimberley and Grandstaff (1986)	Black Reef Quartzite	South Africa	2.4–2.0 Ga
Gay and Grandstaff (1980)	Huronian SpGp	Canada	2.3 Ga
Farrow and Mossman (1988)	Huronian SpGp	Canada	2.3 Ga
Yang and Holland (2003)	Hekpoort paleosol Pretoria Group	South Africa	2.23 Ga
Gutzmer and Beukes (1998)	Gamagara Fm	South Africa	2.2–2.0 Ga
Jackson (1967)	Gowganda Fm	Canada	2.2 Ga
Retallack (1986b)	Paleosol near Waterval Onder	South Africa	2.2 Ga
Holland and Zbinden (1988)	FlinFlon Paleosol	Canada	1.8 Ga
Driese et al. (1995)	Lochness formation	Australia	1.8 Ga
Gall (1994)	Proterozoic Thelon paleosol	Canada	1.7 Ga
Driese and Gordon-Medaris (2008)	Baraboo Range	USA	1.7 Ga
Zbinden et al. (1988)	Sturgeon Falls paleosol	USA	1.1 Ga
Retallack and Mindszenty (1994)	paleosols from northwest Scotland	UK	1.1 Ga

Fig. 3.2 Terrestrial sedimentary biostructures: (**a**) oblique cross section of drab-haloed tubular structures (*dashed lines*) from two successive paleosols (numbered 1 and 2, respectively) in redbeds of the 1.8 Ga Elgee Siltstone, 20 km south of Wyndham, Western Australia (Schmidt and Williams 2008): (**b**) cross section of a 2.7 Ga paleosol with silicified tubular structures (*arrow*), and bound by water-laid deposits. Carraman Formation in Nobles Nob mine, 14 km east of Tennant Creek, Northern Territory, Australia (Black 1984): (**c**) plan view of a sandstone surface of a cracked paleosol (A horizon) as if it were clayey, in the 1.8 Ga Stirling Range Sandstone, Barnett Peak, Western Australia (Bengtson et al. 2007): (**d**) sandstone surface with elephant-skin texture (*Rivularites repertus*) with three individual fossils of *Dickinsonia costata* (*encircled*) in the 550 Ma Ediacara Member of the Rawnsley Quartzite, from Crisp Gorge, South Australia (Retallack 2013); scale bar = 1 cm

Formation are shallow marine stromatolites, but the basal sandstones of the Formation unconformably overlay the clayey paleosol developed on basaltic basement (Fig. 3.1b, c), which has been interpreted as estuarine and littoral talus (Allwood et al. 2006). Possible taxonomic affinities of these trails are unknown other than they are small, simple, and very likely prokaryotic (Wacey et al. 2008). Small spheroidal microfossils found in the same formation (Sugitani et al. 2013) may have lived in shallow waters, although nonmarine environments, including fluvial and playa lake environments (Hickman 2008), were also present at the same time and space. In highly oxidized paleosols, where the primary organic matter may not be preserved, casts of putative microfossils may appear as ordered particles along "filamentous" structures that cut across bedding planes (Fig. 3.3a).

Thucomyces lichenoides is a striking fossil (Fig. 3.3b, c) forming palisades atop green paleosols in the carbon leader of the Central Rand basin of South Africa, and associated with filamentous (hyphae-like) microfossils (Hallbauer and van Warmelo 1974; Hallbauer et al. 1977; Mossman et al. 2008). *Thucomyces* is now known to be as old as 2.8 Ga (Schaefer et al. 2010) and has been compared with lichens because it has clavate structures above the hyphae-like filamentous mesh and with the endocyanotic living glomeromycotan fungus *Geosiphon* (Schüßler and Kluge 2000). However, there are complex radial and vertical partitions within *Thucomyces* (Hallbauer and van Warmelo 1974; Hallbauer et al. 1977) unlike the central chamber of *Geosiphon,* where the latter hosts symbiotic *Nostoc* filaments. These folded structures are comparable with those in columnar biofilms of hot springs today (Hall-Stoodley et al. 2004). The enigmatic microfossil *Diskagma buttoni* from a 2.2 Ga paleosol in the Hekpoort Formation of South Africa (Fig. 3.3d, e) has also been compared with the living fungus *Geosiphon* given its ellipsoidal central chamber (Retallack et al. 2013b). Suggestive evidence of Archean and Paleoproterozoic filamentous Actinobacteria and fungi may explain the carbon-lean composition of many Precambrian paleosols and may imply the existence of mature biocrusts.

Microfossils recorded from the 2.7 Ga Mount Roe paleosol near Whim Creek, Western Australia, are associated with extremely light-carbon isotopic values suggestive of methanogens that now live in swamp-like environments (Rye and Holland 2000). In this case, the presence of organic matter and microfossils in the same suite of paleosols speaks for a wide distribution of microbial life in widely varied terrestrial environments.

Other Precambrian microfossils have been reported for nonmarine sedimentary paleoenvironments, including deposits of lakes of the ~1.1-Ga-old Torridonian Group in Scotland (Cloud and Germs 1971; Strother et al. 2011), and alluvial settings (Beraldi-Campesi et al. 2014) and paleokarst fill (McConnell 1974; Horodyski and Knauth 1994) of the ~1.2-Ga-old Apache Group in southwestern USA. These microfossils display coccoid and filamentous shapes and likely include both prokaryotic and eukaryotic components. From the sedimentary setting of these microfossils, they were more likely washed in from the surrounding land, rather than introduced from the sea. Aquatic terrestrial environments, such as lakes, also supported large organisms such as the 1.5-Ga-old *Horodyskia moniliformis*, which

Fig. 3.3 Terrestrial microfossils: (**a**) petrographic thin section of oxidized filaments running across bedding planes (*arrow*) and more abundant in the lower paleosol than the overlying cross-bedded sandstone of the 1.8 Ga Stirling Range Sandstone on Barnett Peak, Western Australia (Bengtson et al. 2007); scale bar = 10 mm: (**b, c**) *Thucomyces lichenoides* from 2.8 Ga (Schaefer et al. 2010) Carbon Leader of the Central Rand Group near Carletonville, South Africa (Hallbauer and van Warmelo 1974; Hallbauer et al. 1977); scale bars = 200 μm. (**d, e**) petrographic thin sections of *Diskagma buttoni* to reveal a thallus group (**d**) and details of the apical cup (**e**) from the 2.2 Ga Waterval Onder paleosol of the upper Hekpoort Formation, 2 km east of Waterval Onder, South Africa (Retallack et al. 2013a, b); scale bars = 500 μm and 50 μm, respectively; (**f**) acritarch with hyphal stalk (*Ceratosphaeridium mirabile*) from the 570 Ma Wilari Dolomite Member, Tanana Formation, Observatory Hill No. 1 well, northern South Australia (Grey 2005); scale bar = 50 μm: (**g**) *Germinosphaera* sp. indet. from 590 Ma ABC Range Quartzite, SCYW 1a bore, South Australia (Grey 2005); scale bar = 20 μm: (**h**) *Tappania* sp. from 820 Ma Wynniatt Formation, on Victoria Island, Nunavut (Butterfield 2005); scale bar = 100 μm. Images (**a**) and (**b**) are courtesy of Kathleen Grey, (**f**) of Nick Butterfield and (**g, h**) of Dieter Hallbauer, all reproduced with permission

has been interpreted as prokaryotic colonies (Knoll et al. 2006), agglutinated foraminifers (Dong et al. 2008), brown algae (Grey and Williams 1990), sponge remains (Hofmann 2001), hydrozoan or bryozoan colonies (Fedonkin and Yochelson 2002), or metazoan fecal strings (Yang and Zheng 1985), but it could also be of fungal affinity (Retallack et al. 2013b). In any case, its presence in paleolake shores (Retallack et al. 2013b) indicates an ample distribution of nonmarine life in the Precambrian.

Another line of evidence for Precambrian life on land is the suggestion of Pirozynski (1976) and Butterfield (2005) that there is a Precambrian record of fungi among the enigmatic microfossil palynomorphs known as acritarchs (Grey 2005; Moczydłowska et al. 2011). Ediacaran (~580 Ma) acritarchs such as *Ceratosphaeridium* (Fig. 3.3f) and *Germinosphaera* (Fig. 3.3g), and Cryogenian-Mesoproterozoic (~850 Ma) fossils such as *Tappania* (Fig. 3.3h), are similar to Glomeromycotan chlamydospores and vesicles. Many of these fossils have attached aseptate hyphal structures, characteristic of Mucoromycotina and Glomeromycota fungi. These fungi today mainly represent saprobiotic and mycorrhizal growth forms, respectively, but it is unclear whether they avoided lake or marine habitats in the distant past. These fossils have not been considered anything but marine until recently, and reevaluation of the sedimentary facies and communities of these fossils is needed.

3.4 Sedimentary Biostructures

Although microbial mats developing on coastal carbonate facies (e.g., Demicco and Hardie 1995) can be included in the definition of "biological soil crust" (Belnap and Lange 2001), most modern biocrusts develop on sandy, subaerial, siliciclastic substrates. Many sedimentary deposits of this type, from the Archean onward, display sedimentary biostructures of presumed biotic origin and developed subaerially (Hupe 1952; Lannerbro 1954; Voigt 1972; Eriksson et al. 2000; Prave 2002; Noffke et al. 2013; Simpson et al. 2013; Beraldi-Campesi et al. 2014). These diverse sedimentary biostructures can also be found in modern terrestrial systems, such as supratidal, fluvial, and alluvial settings, and thus indicate that the mechanisms that operate today in forming those structures are similar to those operating billions of years ago. In this regard, it must be said that the main biotic components of this type of modern biostructures are cyanobacteria and particularly filamentous taxa. These bacteria are the primary producers that allow other organisms, such as heterotrophic bacteria and fungi (among others), to become part of the biocrust community and are also essential for the cohesive properties of terrestrial MISS (Garcia-Pichel and Wojciechowski 2009; Beraldi-Campesi and Garcia-Pichel 2011; Retallack 2012).

A striking indication of MISS from dry land was first indicated by Prave (2002), who pointed out that some Precambrian sandstone surfaces show a system of cracks morphologically similar to desiccation cracks in clay stones (Fig. 3.2c).

Comparable features are common in modern biocrusts (Retallack 2012), and comparable gypsic paleosols can be found under paleosols with such cracked surface sandy horizons (Fig. 3.1d). For all these examples, it is clear that clays do not play a dominant role in the formation of sedimentary biostructures (or MISS; Noffke 2010), as the cohesive behavior of a microbial cover does, in this case a biocrust-like community (Beraldi-Campesi and Garcia-Pichel 2011).

Both filamentous and EPS-producing taxa are necessary to form sedimentary biostructures. Filaments can build tight networks among clastic particles and support the biocrust framework, while EPS acts as a gluing agent that keeps particles together, especially if it is well hydrated (Noffke 2010). Aquatic and terrestrial and recent and ancient microbial communities can develop a variety of microscopic and macroscopic features that can be recognized in thin sections and in the field, as listed in Table 3.2.

Table 3.2 Examples of sedimentary biostructures (MISS) and features found in microbial mats and earths from siliciclastic environments

Feature	Observations	References
Cohesive behavior	Ductile rheological behavior of microbial mat layers before and after burial	Fagerstrom (1967), Schieber (1999), Eriksson et al. (2000) and Beraldi-Campesi and Garcia-Pichel (2011)
Collapsed gas domes	Desiccated gas domes that form hemispherical depressions	Beraldi-Campesi et al. (2014)
Desiccation roll-ups	Organo-mineral layers upturned and rolled by desiccation	Beraldi-Campesi and Garcia-Pichel (2011)
Erosional pockets	Formed after fragments of microbial earths or mats have been removed from site	Noffke (2010)
Fairy concentric rings	Formed by small-scale plume degassing or dewatering	Gerdes et al. (1993)
Gas domes and blisters	Formed by upcoming gas pressure underneath sealed microbial mats	Bose and Chafetz (2009) and Beraldi-Campesi et al. (2014)
Gravity or flow-mediated soft deformation	Creases, wrinkles, crinkles, crumples, kinks and furrows, of mm to cm scales	Beraldi-Campesi et al. (2014)
Irregular desiccation crack margins	Unlike pure clay cracks, these have irregular margins	Gerdes (2003)
Laminated leveling structure	"Leveling" of the microtopography due to growth and stabilization of the microbial mat	Gerdes et al. (2000)
Mat chips	Eroded mat fragments transported by flows	Fagerstrom (1967), Pfluger and Gresse (1996) and Schieber 2007
Mat-layer bound small grains and heavy minerals	Accumulation of heavy minerals and micas on bedding planes	Noffke (2009)

(continued)

Table 3.2 (continued)

Feature	Observations	References
Microsequences	mm- to cm-thick sedimentary layers, normally graded, often with organic matter vestiges	Noffke (2010)
Oriented grains	Elongated grains that have been oriented parallel to the bedding plane through microbial baffling	Noffke et al. (1997)
Petees	Ridged structures caused by dewatering along weak planes	Reineck et al. (1990)
Rolled up mats	Organo-mineral layers upturned and rolled by wind or water currents	Simonson and Carney (1999), Schieber (2004) and Eriksson et al. (2007)
Sand/sandstone cracks	Also called earth cracks, are formed by desiccation of microbial mats on sandy surfaces	Picard and High (1973), Sarkar et al. (2008) and Beraldi-Campesi et al (2014)
Sand folds	Organo-mineral layers folded by wind or water flow	Bouougri and Porada (2012)
Sand ridges	Similar to Petee structures but also caused by expansion of biocrust boundaries	Beraldi-Campesi et al. (2014)
Sinoidal structures	Preserved sinuous shapes on ripple marks stabilized by microbial mats	Noffke (2010)
Sponge pore fabrics and vesicular horizons	Gas-produced pores formed within or underneath microbial mats and earths	Noffke (2010) and Beraldi-Campesi et al. (2014)
Tufts, knobs and pinnacles	Protuberances, mm to cm tall, derived from microbial growth and behavior in microbial mats	Gerdes et al. (2000), Noffke et al. (2001), Rosentreter et al. (2007) and Noffke et al. (2008)
Wrinkle marks	Include a wide variety of rugose surface structures due to mechanical and behavioral traits	Hagadorn and Bottjer (1997), Hagadorn et al. (1999) and references therein, Gerdes et al. (2000), Porada and Bouougri (2007) and Shepard and Sumner (2010)

3.5 Conclusions

A great variety of terrestrial environments are known in the Precambrian, including soils, peats, ponds, lakes, streams, deserts, and dune fields (Rye and Holland 1998; Eriksson et al. 2004; Retallack et al. 2013a, b). Thus, environmental settings that could potentially become "terrestrial ecosystems" were already in place. Despite sustained uncertainties about these environments being populated or not (Shear 1991; Behrensmeyer et al. 1992; DiMichele and Hook 1992; Gray and Shear 1992; Gray and Boucot 1994; Bambach 1999; Blackwell 2000; Corcoran and Mueller 2004; Nesbitt and Young 2004; Gensel 2008), it is clear, through many lines of

evidence, that terrestrial ecosystems were indeed present, full of life, and functional since the Archean (Beraldi-Campesi 2013). The main lines of evidence are (a) paleosols with redoximorphic and concretionary features that are difficult to explain without the presence of microbial life, (b) microfossils found in rock deposits that represent "terrestrial" environments, and (c) sedimentary biostructures that are nonaquatic and microbially mediated. According to this evidence, the studied rock record shows that the period around the Mesoarchean to the Neoarchean (~3.2–2.5 Ga) was especially important for the development of terrestrial ecosystems, not because they could not exist before but because more plausible evidence is found in strata of that age. This "terrestrial" evolution was perhaps linked to supercontinent growth (Santosh 2010) and the associated emergence and diversification of potential new habitats and potential new biota.

The general similarity of some ancient fossils with fungi (e.g., Hallbauer and van Warmelo 1974; Hallbauer et al. 1977; Mossman et al. 2008; Rye and Holland 2000; Retallack et al. 2013a, b) may be of great importance if these turn out to be phylogenetically related. Besides pushing back their antiquity, the existence of fungi in Proterozoic rocks would speak for a greater similarity between ancient and modern biocrusts. Also, the functioning of the ancient terrestrial ecosystems should be reexamined through new insights about the ecological role of fungi in biocrusts and their potential ability to distribute nutrients (or make them available) throughout extensive areas (Collins et al. 2008). It is necessary to continue studying the ancient record of exposed continental masses in order to provide a better appreciation of the impact of life on the development of Earth's biosphere, especially that involving links between continents and oceans; however, by now, it should be well established in science that Precambrian landscapes were not barren.

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