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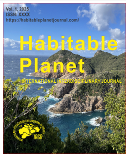

Darwinian soil evolution: A review

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ABSTRACT

Soils diversified through geological time with the evolution of life on land, from microbes in Archean soils, to fungi and amoebozoa in early Proterozoic soils, and lichens in late Proterozoic soils. Ordovician non-vascular land plants lived in Entisols and Aridisols, but trees created clayey subsurface horizons of Alfisols and Ultisols, and ferruginous horizons of Spodosols. Finally, sod grasslands created distinctive surface horizons of Mollisols.

Organisms can be adapted to particular environments, but also adapted by natural selection to other organisms in coevolutionary pairs, such as grasses and grazers, or trees and termites. Coevolution is notable for creating communities that can change the environment, such as ice ages brought on by the evolution of forests and then grasslands.

If life can be defined as a system capable of Darwinian evolution by natural selection, the same can also be said of soils. Even before the evolution of life, clay formation in soils that were like carbonaceous chondrite meteorites would have maintained a place in the sun and rain for the formation of more clay. Soils at matric potential for water were like multicellular creatures with tiny reactive centers within menisci of water between grains. They would have been a starter system for entirely organic-based life capable of natural selection with the same aim: hold the ground against erosion.

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Research Highlights

- Soils diversified through geological time with the evolution of life on land.
- Trees created Devonian Alfisols, then Carboniferous Ultisols and Spodosols.
- Sod grasslands created distinctive surface horizons of Mollisols by Miocene.
- Adaptation to other organisms rather than just environment can change environments.
- Before life, clayey soils like carbonaceous chondrites were naturally selected.

1 Introduction

Life and soil have both been difficult to define, in part because they are as familiar, yet as elusive, as love and home (Retallack, 1998). After long debate about whether life should be defined by complexity, metabolism, mortality, or reproduction, there is some consensus that life should be defined as a system capable of Darwinian evolution by natural selection (Benner, 2010; Retallack, 2019). Similar debates about definition of soil (Jabr, 2024), especially within the context of other planets such as Mars and the Moon (Broz, 2020), have devolved on whether biological weathering is essential to soil (Soil Science Society of America, 2022), or are chemical and physical weathering

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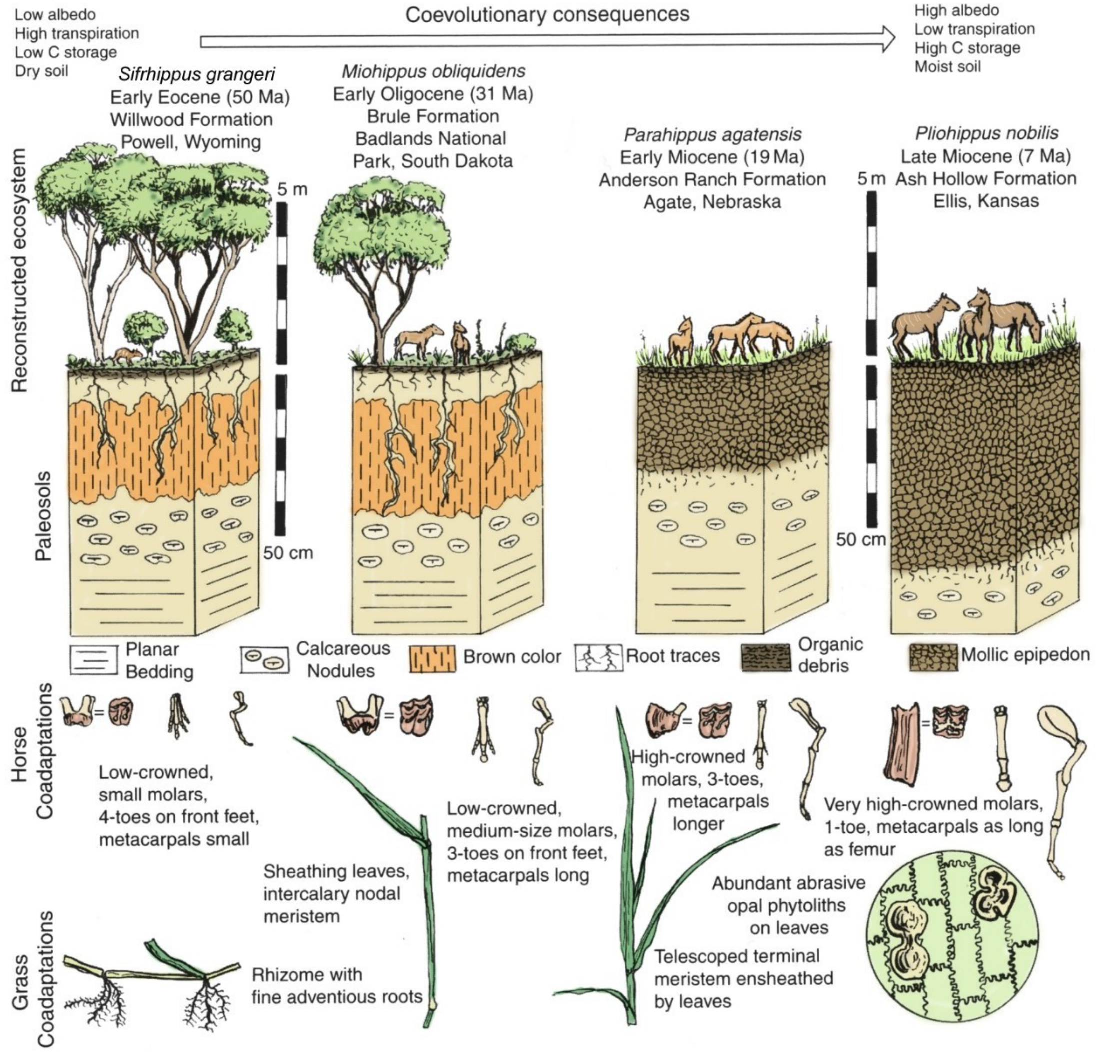

Fig. 1. Paleosol-based reconstruction of habitats and vegetation over the past 50 million years in North America, with evolution of teeth and limbs in horses, and of defensive mechanisms in grasses (modified from Retallack, 2007).

sufficient, with biological weathering an option (Hartemink, 2016). This paper explores the idea that soils too can be considered systems capable of Darwinian evolution, and that soils may have been important in the origin of life.

The idea that life is essential to soils goes back to Dokuchaev's (1883) seminal monograph on Russian Chernozem (Mollisol in US taxonomy), showing that those soils were not merely chemically weathered loess, nor physically deposited eolian or alluvial beds. Chernozems are a biological product of worms and marmots underground, and antelope grazing above to create sod grasslands, within a particular climatic zone of semiarid to subhumid plains. Grasslands can be regarded as ecosystems adapted by natural selection to a particular climate and environment, but not only that. Darwin (1881) also appreciated the role of worms in soil formation, and that organisms became adapted not only to environments, but to other organisms, such as the coevolution of bees and blossoms (Darwin, 1862). Darwin corresponded with Kovalevsky (1873) about his pioneering work on the Cenozoic coevolution of grasses and grazers within grasslands. These concepts are emphasized in this review of the coevolution of both soils and their biota, because simple adaptation of organisms or soils to their environment does not change their environment. In contrast, coevolution with other organisms offers the prospect of changing the world (Retallack, 2013a).

2 Grass-grazer coevolution and global change

Major features of evolution from dawn horses the size of a domestic cat to thoroughbreds of today over the past 60 million years was first outlined by Kovalevsky (1873). Similar changes in size, teeth, and cursoriality have been found for many different kinds of mammals (Fig. 1), especially perissodactyls and artiodactyls (Janis, 1989). With increased size, herbivores evolved larger teeth that protruded further from the gum line (hypsodont), with flat grinding surfaces that were self-sharpening due to ridges of hard enamel versus soft dentine, as adaptations to the abrasiveness of grass leaves. Manouverability was less advantageous in open grassland, where improved running escape speed came from increased limb lengths by conversion of foot and hand bones (metatarsals and metacarpals respectively) to yet another major limb bone. Versatile five-fingered hands and feet were sacrificed for the simple hoof of perissodactyls, or pair of hooves of artiodactyls, best for galloping over open ground. As mammals adapted to open ground, grasses were adapting to the increasingly effective onslaught of hard hooves and high crowned teeth (Stebbins, 1981). Grasses evolved underground stems (rhizomes), where new leaves were produced from telescoped apical meristems, hidden low within sheathing leaves, so that feeding on the ends of leaves did not interfere with new leaf production. Grass leaves also evolved sheaths at their base with intercalary meristems that produce more leaf when the ends were eaten, resulting in greater growth on heavily grazed swards (McNaughton, 1979). Older parts of the leaves became encrusted with plant opal bodies (phytoliths: Fig. 1), which were abrasive like sandpaper for mammal teeth (Strömberg, 2006). More heavily grazed grasses had a higher density of these abrasive particles than less grazed grasses (Retallack, 1992). Roots of sod-forming grasses are adventitious and slender (1–2 mm), forming a dense network underground like a thick carpet to resist disruption by hooves. Bulldozer herbivores, such as elephants, cleared trees for the continued invasion of grasses (Leeuwis et al., 2018).

The natural carpet of sod-forming grasses is a key feature (mollic epipedon) of Mollisols and Chernozems, defined as a surface horizon at least 18 cm thick, dark with at least 2.5% organic carbon, and fertile with at least 50% base saturation (Soil Survey Staff, 2020). Grazing of grasslands drives carbon and productivity underground where it feeds a diverse community of earthworms and other invertebrates (Darwin, 1881). The structure of the mollic epipedon consists of crumb peds of fertile smectite clay the size of rice grains, stabilized against slaking by organic coatings from the fine network of roots, and passed as coated fecal pellets of earthworms. It is this distinctive ped structure that is key to recognizing Mollisols in the rock record, because paleosols usually lose an order of magnitude of organic carbon compared with similar surface soils (Broz, 2020). From surveys of paleosols on all continents except Antarctica (Retallack, 2013a), the record of Mollisols, and thus sod grasslands, goes back only about 20 million years (Fig. 1), a timing confirmed by studies of fossil grass phytoliths (Strömberg et al., 2018).

The appearance of fossil Mollisols, grazing mammals, and sod-forming grass phytoliths in geological time could be taken as an evolutionary response to climatic drying and cooling (Janis, 1989; Strömberg et al., 2018). However, that does not account for the geographic expansion of grasslands at 7 million years ago, evident from mammals (Janis, 1989), phytoliths (Strömberg et al., 2018), and paleosols (Retallack, 2013a) at a time of higher atmospheric CO_2, and warmer, more humid climate (Retallack and Conde, 2020). This range extension to a different climatic belt means that grasslands, at first limited to semiarid regions some 20 million years ago judging from climatically sensitive features of their paleosols, expanded their range to subhumid regions about 7 million years ago (Retallack, 2013a). Paleoclimatic and paleogeographic expansion of the empire of grass also coincides with declining atmospheric CO_2 and cooling as determined from an isotopic paleobarometer using pedogenic carbonate, and from stomatal index of fossil leaves (Retallack and Conde, 2020). This CO_2 drawdown can be blamed on carbon sequestration and chemical weathering in newly evolved Mollisols, and cooling was also aided by high albedo and lower transpiration of grasslands compared with previous shrublands and woodlands (Fig. 1; Retallack, 2007, 2013a). Grassland Mollisol expansion into desert Aridisols also mitigated a middle Miocene greenhouse spike of 588 ppm CO_2 from massive volcanic eruptions of the Columbia River Basalt, bringing it back to levels closer to 300 ppm within less than a million years (Retallack, 2022a). Thus, grasslands did not evolve to fill in dry spots on the landscape, but coevolutionary obsession of grazers with grasses, and grasses with grazers, made grasslands a global force for cooling and drying (Retallack, 2013a).

Grasslands are the source of human agroecosystems, providing both grain and meat, so that human management is critical to their future, and may be their undoing (Stebbins, 1981). Current global domestic livestock biomass has been estimated at 630 Mt, compared with human biomass of 390 Mt, and total wild terrestrial animal biomass of only 20 Mt (Greenspoon et al., 2023). Traditional agriculture of plowing, herbicides, fungicides, artificial fertilizers and feed has destroyed both soil carbon and topsoil thickness (Ruddiman, 2005). New techniques of carbon farming and regenerative agriculture are needed to enlist farmlands in the quest for carbon sequestration (Brown, 2018).

3 Forest-arthropod coevolution and global change

A seminal event in the history of soils was evolution of trees (defined as at least 10 m tall) during the Silurian to Devonian: there were no trees before then (Fig. 2). Early Paleozoic life on land consisted of herbaceous plants and lichens on weakly developed soils (Entisols) and desert soils (Aridisols: Retallack, 2015, 2022b). Tree stature was achieved independently in several lineages of plants, by

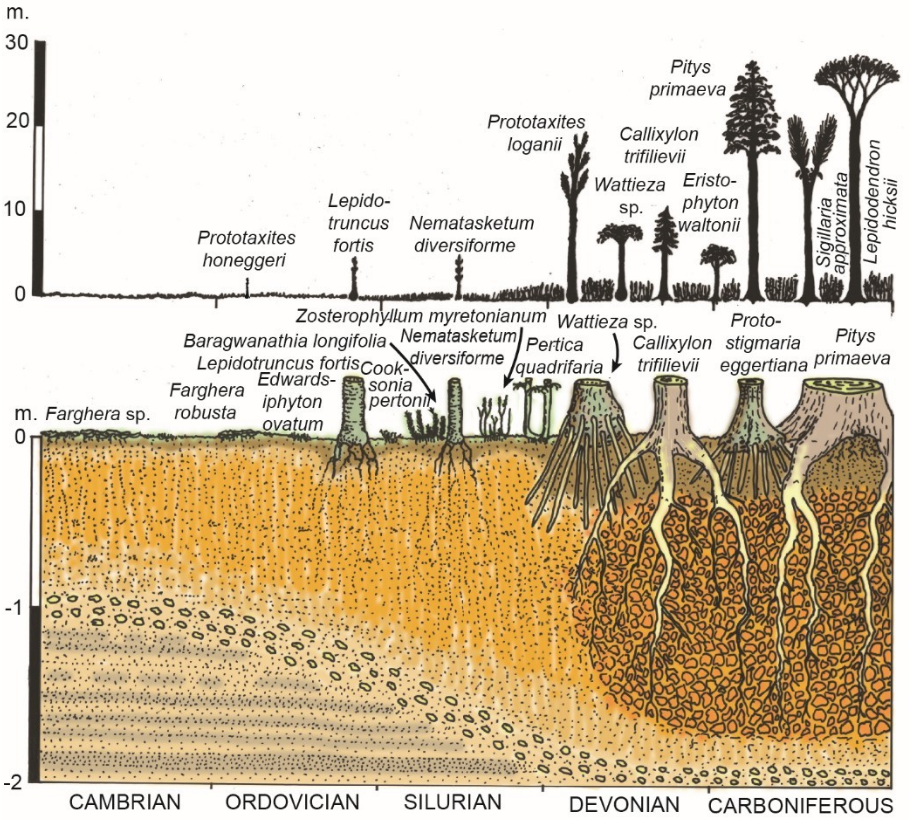

Fig. 2. Early Paleozoic plant and soil coevolution from 539 to 299 million years ago, showing maximum depth of calcic (Bk) horizon (rounded light yellow nodules), of mycelial bioturbation (dotted vertical lines) and rhizoids and roots (solid lines). Cambrian to Silurian soils have simple profiles thet would be classified as Aridisols (A-Bk or A-Bw-Bk) homogenized by cracking, rhizines and mycelia, but Devonian and later soils have thicker profiles of the first Alfisols (A-Bt-Bk), with a slickensided hackly (blocky, argillic, or Bt) horizon created by tapering woody roots. Stumps and reconstructions are based on actual fossils (modified from Retallack, 2022b).

evolution of a cambium producing centripetal secondary xylem and centrifugal phloem and bark from precursor protoxylem strands of early vascular land plants in progymnosperms such as *Archaeopteris*, and by bundling of multiple vascular strands in cladoxyls such as *Wattieza* (Stein et al., 2020). Also growing to tree size with growth rings were late Silurian and early Devonian lichens, such as *Prototaxites* (Retallack, 2022b). The selective advantages of large size included competition between plants, because larger plants could shade out lower plants (Boyce et al., 2017). Taller plants could also wage chemical warfare on soil animals and shorter plants with allelopathic phenolic compounds leached from leaves (John and Sarada, 2012). Also selective for trees was avoiding herbivory by millipedes and other non-flying terrestrial arthropods, because indigestible lignin of woody branches created long climbs to feed on nutritious leafy shoots and sporangia (Retallack, 2001). Mesozoic forests were also regulated by coevolution of bees and blossoms and coevolution of termites and trees (Kricher, 2011).

Tree roots had significant effects on soil formation, creating four new soil orders (Fig. 3): Alfisols, Ultisols, Spodosols, and Histosols (Soil Survey Staff, 2020). Strong downward taper of woody roots promoted illuviation of clay to levels diagnostic of subsurface argillic (Bt) horizons, characteristic of Alfisols and Ultisols. Base-rich Alfisols appeared during the Devonian, 386 million years ago (Retallack and Huang, 2011), but base-poor Ultisols only by Carboniferous 320 million years ago, reflecting delayed adaptations to low nutrient substrates (Gill and Yemane, 1996). Spodosols are nutrient-poor and highly acidic sandy profiles that also are difficult for plants, but supported forests by Carboniferous, 335 million years ago (Vanstone, 1991). Histosols are peaty soils at least 40 cm thick, preserved in the fossil record as coal seams by the latest Devonian, 360 million years ago (Gillespie et al., 1981). Early Devonian silicified peats such as the 410-million-year-old Rhynie Chert (Garwood et al., 2020) did not have required thickness of Histosols (Soil Survey Staff, 2020). By Carboniferous time, 300 million years ago, fossil

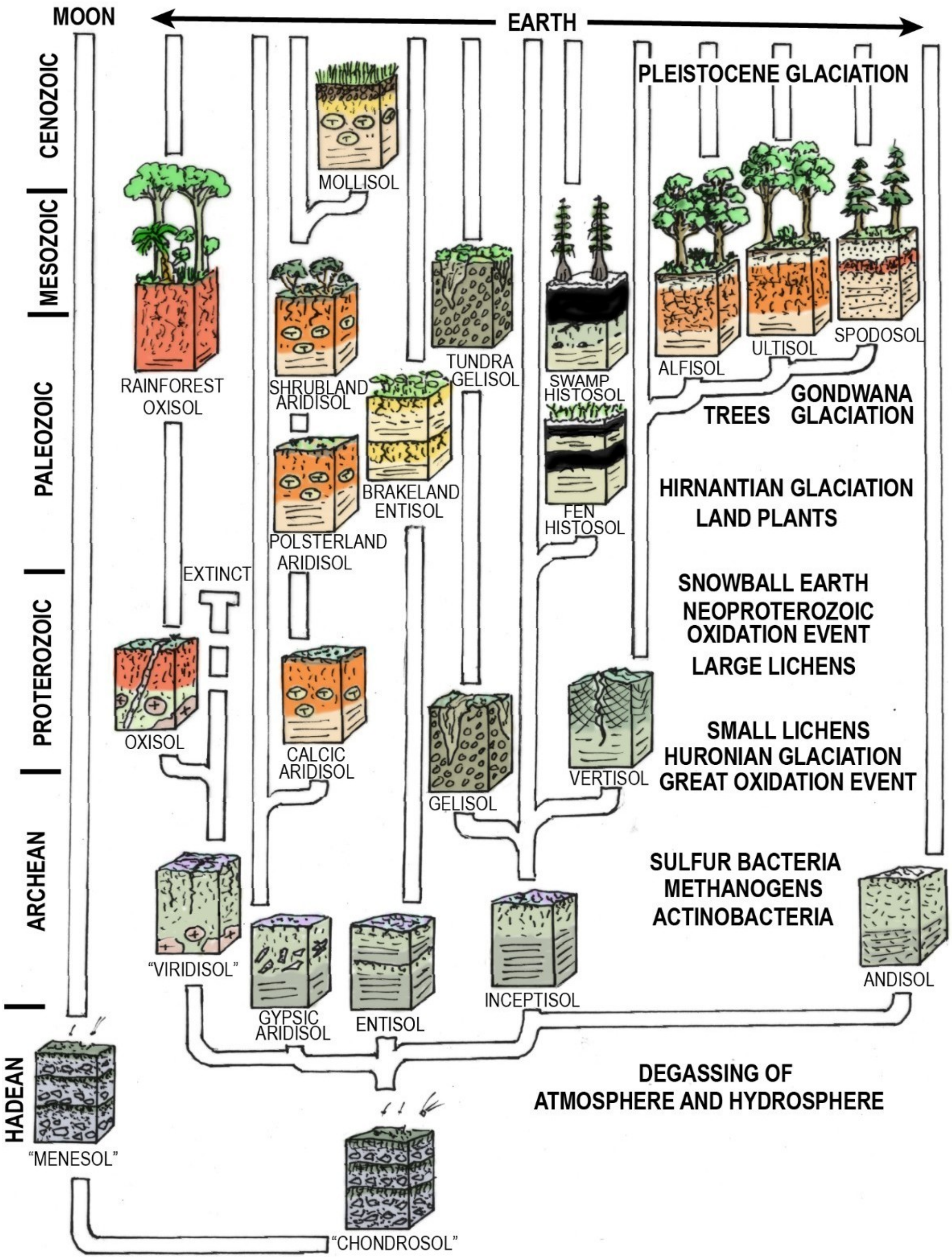

Fig. 3. Evolution of soil orders of the US Taxonomy over the past 4500 million years, as revealed by the paleosols (Retallack, 2019). In quotation marks are unofficial new soil orders for Lunar soils ("Menesols"), profiles found in carbonaceous chondrite meteorite ("Chondrosols") and deep well drained paleosols that remained unoxidized ("Viridisols").

Histosols were widespread, and included calcareous peats with coal balls, which contain fossil plants permineralized to show cellular details (Retallack, 2021a). Oxisols are extremely depleted in nutrients (Ca, Mg, Na, and K), and such deeply weathered soils go back to Archean some 3316 million years ago (Retallack and Schmitz, 2023), but by mid-Carboniferous 305 million years ago, they supported tropical rain forest (Retallack and Germán-Heins, 1994). Forests first evolved in Alfisols of semiarid to subhumid regions, but later the empire of trees expanded their range to nutrient-poor substrates of Ultisols, Spodosols, Histosols and Oxisols.

Paleozoic forests did not fill particular climatic or geographic niches, but changed both climates and landscapes. Pedogenic carbonate and stomatal index of fossil plants are evidence of declining CO_2 in the atmosphere during the Carboniferous (Retallack and Conde, 2020), and there are widespread tillites and other evidence of glaciation in Gondwana countries such as South America, South Africa, India, Australia, and Antarctica (Montañez et al., 2016). Forests re-engineered and cooled the planet by consuming CO_2 from the atmosphere as living biomass, biomass buried as peat, and as carbonic acid consumed in the weathering of feldspar to clay (Retallack, 1997, 2022a). These effects were eventually undone by greenhouse crises from large igneous eruptions (Retallack and Conde, 2020), and the persistent work of fungi, termites, and tree-eating dinosaurs, in what has been called the Proserpina Principle (Retallack, 2019). Forested soils about a meter thick stabilized hillslopes to create characteristically wider ridges than before (Dietrich and Perron, 2006), and stabilized levees and floodplains to create meandering rather than braided streams (Davies and Gibling, 2010).

4 Soil microbiome coevolution and global change

The fossil record of soils during the Precambrian is evidence for major events such as the Great Oxidation event of 2400 million years ago, and extensive glaciations of 2300–2200 and 720–635 million years ago (Retallack et al., 2015, 2021a). Atmospheric oxidation is clear from a change from drab to red colors of paleosols, as well-drained paleosols with deep cracks and corestones became dominated by ferric rather than ferrous iron (Retallack and Schmitz, 2023). Amounts of atmospheric O_2 and CO_2 can be calculated from paleosols using mass-balance geochemical models of oxygen and carbon consumption (Retallack et al., 2021b). Glaciations are clear from paleosols with periglacial deformations, such as sand wedges and ice wedges (Retallack et al., 2015).

A synoptic impression of Precambrian soil microbiomes comes from paleosol salts (Retallack, 2022c), such as the sulfates, barite ($BaSO_4$), kieserite ($MgSO_4$) and gypsum ($CaSO_4.2H_2O$), and the carbonates, nahcohlite ($NaHCO_3$), dolomite ($(Ca_{0.5},Mg_{0.5})CO_3$) and calcite ($CaCO_3$). In modern deserts with gypsic horizons the microbiome is prokaryotic, including cyanobacteria, sulfur bacteria, and actinobacteria, whereas soils with calcic horizons of calcite and dolomite have many eukaryotes, including fungi, lichens, algae, and amoebozoa (Neilson et al., 2017; Araya et al., 2020). Nahcolite-bearing soils and deposits have halophilic prokaryotes (Balci and Demirel, 2016). Calcite precipitation is catalyzed by extracellular polysaccharides of glucose and mannose from cyanobacteria under oxidizing conditions, whereas dolomite is catalyzed by galactose and rhamnose produced mainly by methanogens, fermenting bacteria and sulfur-reducing bacteria under dysaerobic conditions (Zhang et al., 2012). Archean paleosols (more than 2500 million years old) have only sulfates and nahcolite, whereas dolomite appears in paleosols by 2403 million years and calcite by 1460 million years (Retallack, 2023a). The oldest known paleosols 3700 million years old in Greenland probably had kieserite sand crystals now replaced by stilpnomelane (Retallack and Noffke, 2019), and paleosols the same age on Mars had sand crystals of bassanite ($2CaSO_4{\cdot}H_2O$), a low hydration form of gypsum (Retallack, 2014). Dolomite and calcite were rare in paleosols until the Tonian 758 million years ago, but then became locally common in areas separated from sulfate soils during Cryogenian and Ediacaran periods (Retallack et al., 2021a; Retallack, 2023b). There is also information from depth within paleosol profiles of gypsic and calcic horizons, which is proportional to secondary productivity of respired CO_2 in soils (Breecker and Retallack, 2014). Both gypsic and calcic horizons were deeper within paleosols after the Great Oxidation event of 2450 million years ago, and then became deeper again from Tonian 758 million years ago (Retallack, 2022c). These are indications of increased biological productivity on land through time, with additional spikes during the Hirnantian, Gondwana, and Pleistocene Glaciations.

Further information comes from stable isotopes of carbon and oxygen in pedogenic carbonate. The difference between marine and pedogenic carbonate oxygen isotopic composition has been called the Dole effect, and used as a proxy for terrestrial productivity (Bender et al., 1994). Both oxygen and carbon isotopic compositions are fractionated on land by the photosynthetic enzyme rubisco, which selects for light isotopologues of CO_2. Oxygen, but not carbon, is also fractionated by transpiration, that became especially significant after evolution of stomates in land plants (Retallack and Bindeman, 2024). Isotopic composition of carbonate in paleosols and paleokarst does not diverge much from compositions of marine carbonate until the Tonian (758 million years ago), again confirming a rise in terrestrial productivity leading into Neoproterozoic Oxidation Event and Snowball Earth. These data also show spikes in divergence between marine and terrestrial stable isotopic composition of pedogenic carbon during the Hirnantian, Gondwana and Pleistocene Glaciation (Retallack and Bindeman, 2024).

Many Precambrian paleosols have no salts, presumably because they formed in climates too humid for salts to resist dissolution. Many of these paleosols have been studied in detail for geochemical differentiation, and turn out to be surprisingly deeply weathered (Rye and Holland, 1998). From a taxonomic perspective, these profiles are puzzling because they show evidence of good drainage from soil creep of quartz veins and spheroidal weathering at depth, yet remained green-gray and unoxidized. The informal soil order “Viridisol” has been proposed for such profiles (Retallack, 2013b). Another puzzle is that they show light organic carbon isotopic compositions indicative of photosynthetic origin, yet very low total carbon contents. This is the same as in Phanerozoic paleosols (Broz, 2020), and is evidence that their microbiome included decomposers. Especially well known profiles have been studied to assess both strain (loss of volume during weathering) and of components (Ca, Mg, Na, K, P) relative to a stable constituent (usually Ti) assumed to have remained from parent rock or sediment lower within the profile. Additional information about time for formation of the profiles from chronofunctions (Retallack, 2019) allows calculation of rates of weathering for paleosols of various geological ages. These calculations reveal rates of weathering increasing by 5 orders of magnitude since 3700 million years ago (Retallack, 2022d). Rates of nutrient base weathering increased on average 4 orders of magnitude during the Archean followed by a plateau of average values after the Great Oxidation Event and another order of magnitude in some paleosols of the Phanerozoic (Fig. 4A). Rates of phosphorus weathering, largely from the relatively insoluble apatite, increased to peaks at both the Great Oxidation Event and Huronian glaciation (2460–2060 million years ago), and then again later at the Neoproterozic oxidation event and Snowball Earth glaciation (720–635 Million years ago: Fig. 4B). This biotic enhancement of weathering would have promoted oxidation and glaciation by carbon burial in biomass and carbonic acid consumption, despite smaller land areas in the Archean than in the Proterozoic (Fig. 4C: Cawood and Hawkesworth, 2019).

Direct evidence of Precambrian soil microbiomes comes from microfossils permineralized by silica within paleosols and from microbial trace fossils. Large lichens with permineralized thallus structure showing both photobiont and haustorial hyphae are known from the Ediacaran paleosols some 635 million years old (Retallack, 2022e), and plausible lichen fragments have been found in Tonian paleosols some 758 million years old (Retallack, 2021b, 2023b). Fungi are known from paleosols 2100 million years old (Retallack et al., 2013; Retallack, 2021c) and slime molds in paleosols 1900 million years old (Retallack and Mao, 2019). Silica-permineralized Archean paleosols 3016 million years old contain a variety of microfossils representing purple sulfur bacteria, methanogens, and actinobacteria, as inferred from unique carbon isotopic composition of individual fossil cells (Retallack et al., 2016). Framboids in the same paleosol are evidence of sulfur-reducing bacteria, completing the sulfur cycle. The accumulation of sulfate sand crystals in these paleosols may have been oxidized biologically by the activity of photosynthetic purple sulfur bacteria. The presence of such oxidized sulfate, created as a waste product of anaerobic photosynthesis, would otherwise be a puzzle in a soil with so little free oxygen, as indicated by its iron redox chemistry (Retallack et al., 2016). Hematite in waterlogged Archean paleosols is similar to hematite in banded iron formations, and also was not a direct oxidation product, but the metabolic waste of photolytic iron-oxidizing bacteria (Retallack, 2018). Precambrian soils had complex microbiomes of producers and consumers that coevolved for improved persistence and nutrient acquisition through time.

From this perspective, the Great Oxidation Event and Huronian Ice age can be seen as a transition from Archean prokaryotic communities of anaerobic photosynthetic sulfur bacteria, methanogens, and actinobacteria, to Proterozoic eukaryotic communities of cyanobacteria, amoebozoa, and fungi. The Neoproterozoic Oxidation Event and Snowball Earth was yet another step to multicellular life on land in lichenlike organisms, and a necessary prelude to supply nutrients for the Cambrian Explosion of life on land (Retallack, 2011) and at sea (Smith and Harper, 2013).

5 Soils and the origin of life

Charles Darwin envisaged the origin of life in a warm little pond by the seaside (Peretó et al., 2009), but others prefer origins in deep sea black smokers (Martin et al., 2008) and continental hot springs (Djokic et al., 2017). There is a better case to be made for soil as the cradle of life because of two fundamental problems with alternative sites (Retallack, 2019). First is the nearly impossible odds of making anything as complex as even an enzyme by chance. Second is Le Chatelier’s Principle of chemical equilibrium, which posits that molecules formed in extended aqueous medium, even as small as a soup bowl, are as likely to break down as persist, within an equilibrium balance of creation and destruction. The key to convert this long chance to necessity, is natural selection of complex molecules (Monod, 1971). Formation and persistence of biological polymers require wetting and drying cycles so that monomers can assemble by dehydration, and persist by folding (Matange et al., 2025). A precursor for complex biological molecules may have been “clay life” of Cairns-Smith (1971), preferentially accumulating during wet and dry cycles of soil weathering. Different kinds of clays such as smectites have different physical properties (like a phenotype) based on internal crystal lattices (like a genotype). Soil formation naturally selects for clays that can swell to bridge gaps between feldspar grains when wet, and then crack to reveal more of the source mineral to form more clay (Retallack, 2016). Clays that dissolve in water or form inert rinds are maladaptive because they do not confer resistance to erosion by wind or water. Formation of complex organic compounds by abiotic Urey-Miller synthesis (Boyd, 2012) further stabilizes the soil so it can persist within the

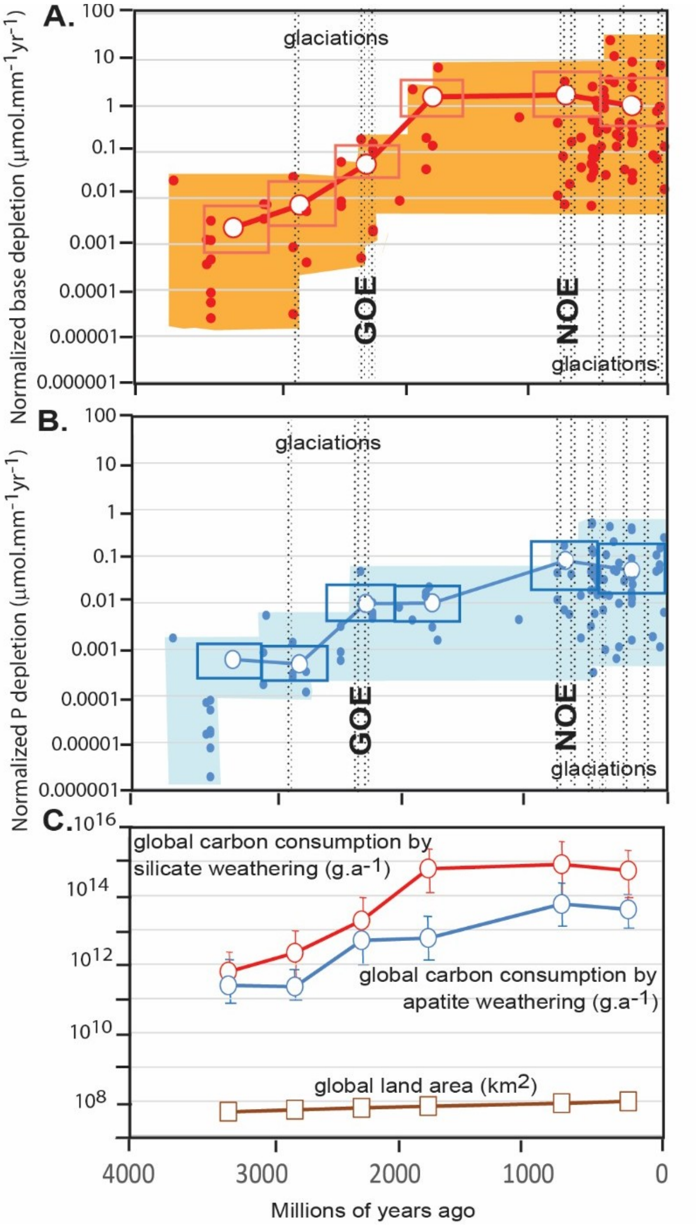

Fig. 4. Nutrient (Ca+Mg+K+Na) base (A) and phosphorus depletion (B), and carbon consumption (C) rates inferred from tau analysis of paleosols over the past 3700 million years. (A–B) Closed symbols are individual paleosols, and large open symbols are mean for 500 Ma intervals. Only a single paleosol is known for 1000–1500 million years ago (C) Annual rates of C consumption from base depletion and apatite weathering and global land area increase are calculated from continental area and freeboard estimates (Cawood and Hawkesworth, 2019). Upper and lower box bounds and error bars are two standard deviations. GOE is Great Oxidation Event, and NOE is Neoproterozoic Oxidation Event. (Data modified from Retallack, 2022d.)

zone of energy and materials flux at the surface (Retallack, 2019).

A paleosol record of this process may be preserved within carbonaceous chondrites, which can be considered primeval paleosols of the Solar System (Fig. 3). Carbonaceous chondrites have been radiometrically dated at 4567 million years old (Connelly et al., 2012). They formed on planetesimals during original condensation of the Solar System (Bunch and Chang, 1980), and persist on asteroids such as Bennu to this day (Lauretta et al., 2019). Carbonaceous chondrites have a puzzling mix of high temperature minerals such as olivine and pyroxene, often in spheroidal form of melt droplets known as chondrules, as well as low temperature smectite clays and organic

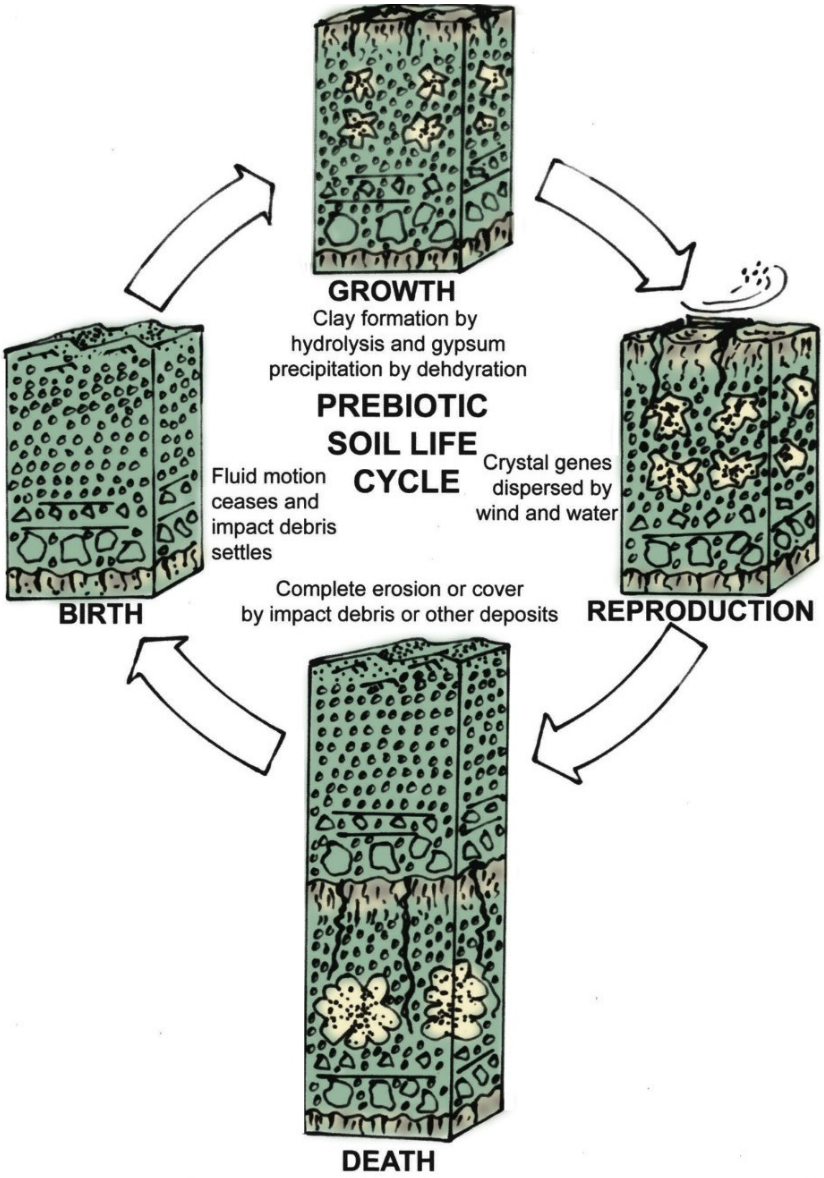

Fig. 5. Life cycle of an hypothetical primeval carbonaceous chondrite soil, on a planetismal in the early stages of formation of our solar system some 4567 million years ago.

matter including amino acids and sugars. The organic matter formed abiotically because it is an even mix of left-handed and right-handed versions of asymmetric molecules, as in the famous Urey-Miller experiments (Boyd, 2012). In contrast, biologically produced organic matter shows marked preferences, left-handed for amino acids and right-handed for sugars.

Carbonaceous chondrite soils can be imagined (Retallack, 2019), and tentatively named “Chondrosols” as a new, but so far informal, order of US taxonomy (Fig. 5). The high temperature minerals formed by impacts and igneous activity were the parent material of these soils, which were weathered by hydrolysis in water to smectite clays and salts such as gypsum. The principal metabolic reactions of these soils were hydrolysis of feldspar to clay and Urey-Miller synthesis of complex organic matter. Clay and organic matter stabilize the soil, as a form of natural selection. Clayey, carbonaceous, abiotic soils at matric potential for water were like multicellular creatures, with tiny reactive centers within menisci of water between grains. Clay and organic matter can also be dispersed by wind or water as fragments to seed other soils, in a form of asexual reproduction. Different combinations of clay and organic matter are needed for persistence in different local climatic regimes, but eventually the life cycle of a soil is ended by erosion or by cover sediment.

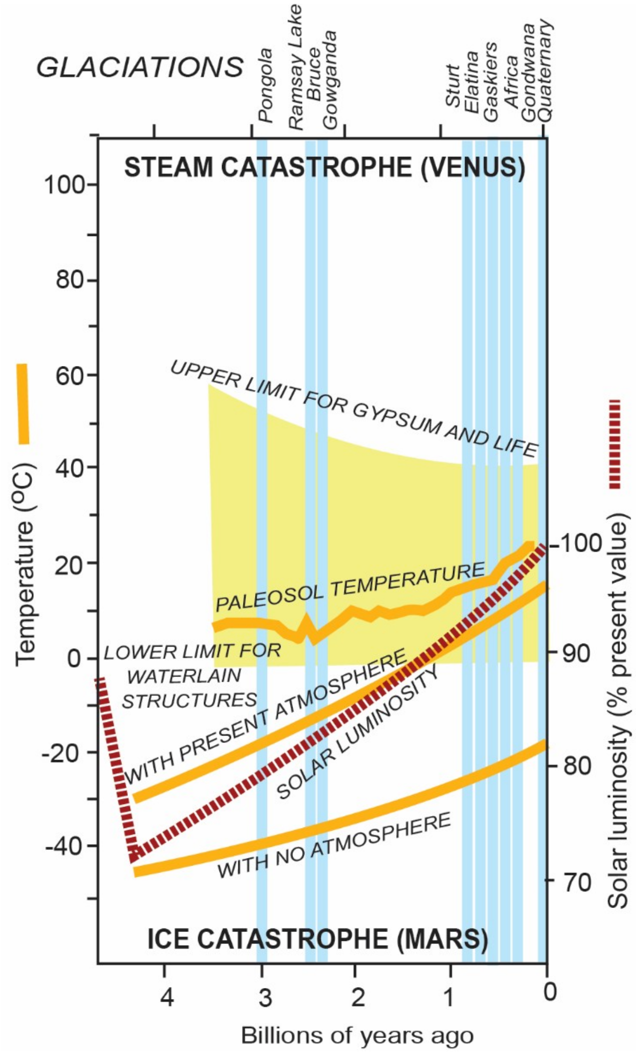

Fig. 6. Observed Earth temperature from paleosols and their evaporite and freezing constraints (primary y axis), theoretical temperature given present or no atmosphere (primary y axis), and calculated solar luminosity (secondary y axis) over the past 4600 million years (modified from Ribas, 2009).

Similar paleosols dating back 3300 million years are known in the Apollo 15 core on the Moon, but these lunar paleosols have no clay or organic matter, and can be tentatively call "Menesols", from Greek mene (f.) for moon (Fig. 3). Soil formation without atmosphere on the moon is not a process of chemical weathering, but of physical disturbance and addition of sand-to-silt grain-sized micrometeoroids. These include Fe–Ni micrometeoroids that enrich metal content and magnetic susceptibility of the surface soil (Heiken et al., 1976). Many of these small impacts are energetic enough to melt target soil into donuts of broken grains cemented by melt glass, known as agglutinates. Tiny droplets and dumbbells of melt glass also build up over time. Successive paleosols, each forming over many tens of millions of years, were isolated by cover of ejecta blankets from much larger bolide impacts nearby. "Menesols" are highly unlikely to have fostered or supported life, and are a null hypothesis for the origin of life from more promising "Chondrosols".

6 Soil planetary thermostat

The long-term habitability of Earth has been a balance between thermonuclear evolution of the sun becoming increasingly bright through time and biotic regulation of greenhouse gases (Ribas, 2009). During the Precambrian the sun was 30% dimmer and Earth should have been frozen over were it not for greenhouse atmospheric gases (Fig. 6). There were some unusually extensive ice caps at about 2400 and 720 million years ago (Hoffman et al., 2017), but even then, some regions of frost-patterned paleosols were without ice cover (Retallack et al., 2015).

Paleosols are evidence of widespread and continuously ice-free soils back 3700 million years with evidence for incongruent dissolution of feldspar to clay in water and precipitation of evaporite minerals at less than boiling temperatures (Retallack and Noffke, 2019). Such benign temperatures require warmth and greenhouse gases, such as CO_2 and CH_4, known from paleosol paleobarometers to have been much more abundant in the atmosphere than today (Breecker and Retallack, 2014; Retallack et al., 2021b). Furthermore, these greenhouse gases are partly controlled by soils, with ever increasing efficiency of biomass creation and chemical weathering through time (Fig. 4). Carbon consumption thus mitigated a greenhouse atmosphere in the long term, but short-term greenhouse spikes from large volcanic eruptions could also be mitigated by expansion of productive ecosystems. For example, short term recovery has been documented from global expansion of highly productive Mollisol and Oxisol paleosols during the middle Miocene CO_2 spike to 588 ppm from eruption of Columbia River Basalts of Oregon and Washington (Retallack, 2022a). Plants and soils cool the planet by sequestering carbon, but animals warm the planet, by the Proserpina Principle, so that coevolution of animals and plants can adjust to each other, as well as to volcanic and other physical forcings (Retallack, 2019).

7 Summary

As life diversified through geological time, so too did soils. Precambrian soils supported mainly microbes, and amoebozoa appeared in soils during the Proterozoic, if not earlier. Late Precambrian paleosols included large multicellular organisms such as lichens. Three great Phanerozoic milestones for soils were evolution of non-vascular land plants within Entisols and Aridisols, then trees creating clayey subsurface horizons of Alfisols and Ultisols and ferruginous horizons of Spodosls, and finally, sod grasslands creating distinctive surface horizons of Mollisols.

Although organisms are commonly considered adapted to particular environments, they also are adapted to other organisms in coevolutionary pairs such as grasses and grazers, and trees and termites. Some of the most bizarre adaptions such as antlers and manes, are for the opposite sex, due to sexual selection. Coevolution and sexual selection are notable for creating maladaptive organisms and communities that can change the environment, such as ice ages brought on first by the evolution of forests, and then of grasslands.

If life can be defined as a system capable of Darwinian evolution by natural selection, the same can also be said of soils. Even before the evolution of life clay formation in soils that were like carbonaceous chondrites would have maintained a place in the sun and rain for the formation of more clay. Organic and clayey fragments eroded from these "Chondrosol" soils could have seeded soils nearby in a form of asexual reproduction. These early clayey, carbonaceous. Soils at matric potential for water were like multicellular creatures with tiny reactive centers within menisci of water between grains.

The last word belongs to Jabr (2024, p. 62). "Much like life, soil has thwarted all attempts to contain it within a succinct and precise definition. Most textbooks and scientific organizations rely on lengthy, convoluted definitions of soil that list its many properties and refer to it as a material or medium. There seems to be growing recognition in mainstream scientific circles, however, that soil may be best understood, not as a substrate for life, but rather as itself a living entity."

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Declaration of competing interest

The author declares that he has no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Credit Author statement

Gregory J. Retallack: Conceptualization; Data curation; Formal analysis; Visualization; Writing—original draft; Writing—review & editing

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