

9.11 Coevolution of Life and Earth

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When I hear the sigh and rustle of my young woodlands, planted with my own hands, then I know that I have some slight share in controlling the climate – Anton Chekhov (1899)

9.11.1 Introduction

Earth is one of the oldest words in our language (from Old English 'eorde') meaning both our planet (customarily capitalized in this sense) as well as the soil beneath our feet. Here I explore both ends of its meaning by emphasizing how small-scale biological and soil-forming processes have played a role in the

very significant differences between our planet and other planetary bodies in our solar system. Such processes as the growth of trees are commonly regarded as mundane, in the sense of commonplace or trivial, but they can be mundane also in the more important sense of global, as apparently understood by Chekhov (1899). The media which elevate biological processes to global scope are air, water, and soil. Air and water are important to the coevolution of life and earth (Nisbet and Sleep, 2001), but we will take the path less traveled here, and present a perspective on coevolution of life and earth from my research experience with ancient and modern soils.

Soil is Anglo-French word, introduced from the ancient Roman 'solium' (seat) or 'solum' (ground). Soil is so central to human existence that it is as difficult to define as 'home' and 'love'. To some soil scientists, soil is material altered by the action of organisms, especially plant roots. To engineers, soil is material moved by mechanized graders without recourse to blasting. To farmers, soil is the surficial organic layer that nurtures crops and feeds livestock (Hole, 1981). We prefer a wider definition of soil as materials on the surface of a planetary body altered in place by physical, biological, or chemical processes. This definition has also been adopted by NASA (National Aeronautic and Space Administration of the USA) press releases on Lunar and Martian soils, for the compelling reason that they wished to be understood by the general public. In any case, the surface of the Moon and Mars is now altered locally by robotic landers and footprints, and our future

alterations can be anticipated. A broad definition of soil obviates the need for such terms as regolith or the distinction between physicochemical and biological weathering (Taylor and Eggleton, 2001). More important to our present purpose, there is no longer an assumption that life created soil. The questions now become when and how did life evolve in soil? Did soil nurture the origin of life, or did life evolve in the ocean or volcanic hot springs, then later adapt to soil? These questions remain open (Nisbet and Sleep, 2001).

When soil, air, and water became alive, unlike the apparently sterile but discernably altered soils of the Moon, Venus, and Mars (Figure 1), it must have been an important time in the history of our Earth. Soils gave life access to fundamental nutrients such as phosphorus at their source in the weathering of rocks. Soils allowed life to colonize a significant fraction of Earth's surface, thus enhancing their volumetric effect on surficial fluids. Life gave soils increased

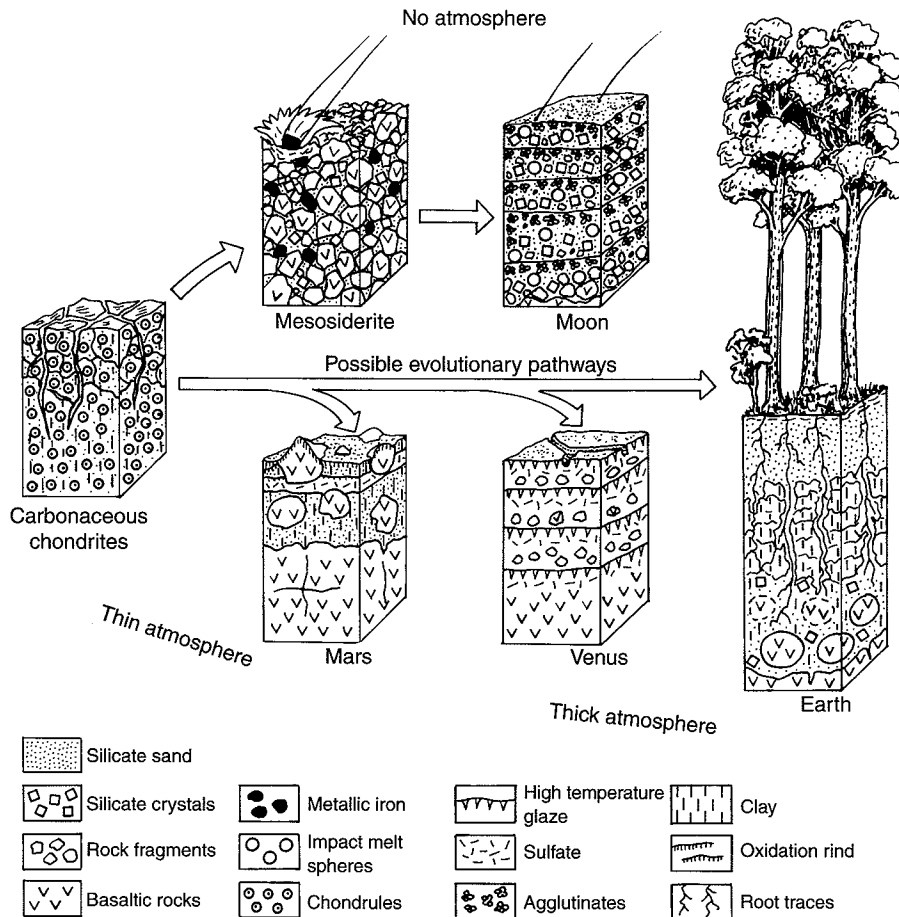


Figure 1 A schematic comparison of soil formation on Earth, Moon, Mars, Venus, and some hypothetical meteorite parent bodies. From Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

depth and stability to enhance rates of physicochemical weathering. Life also gave soils a variety of biosignature horizons and deep weathering products (Retallack, 2001a).

The evolution of life and soil can be viewed as a coevolutionary process, like the coevolution of grasses and grazers, as first proposed by Kowalevsky (1873). Coevolution is the coordinated evolution of phylogenetically unrelated organisms (in this case, plants and animals), which coevolved to enhance their mutual interdependence. Grasses evolved subterranean rhizomes, basal tillering of adventitious roots, intercalary meristems, telescoped internodes, and opal phytoliths to withstand grazing (Figure 2).

Horses evolved cursorial limb structure for predator escape over open country and high-crowned teeth to withstand abrasion from leaves studded with opal phytoliths (MacFadden, 1992). Grasses and grazers coevolved to produce ecosystems, such as grasslands, and soils, such as Mollisols, that became biotic planetary forcings of the atmosphere and hydrosphere (Retallack, 2001b). Because coevolution is directed more toward other organisms than to physicochemical environments, Earth's environment coevolved with life far from primordial conditions exemplified by other planetary bodies (Figure 1).

The coevolution of life and soil is evident from functional intricacies of modern ecosystems, but the

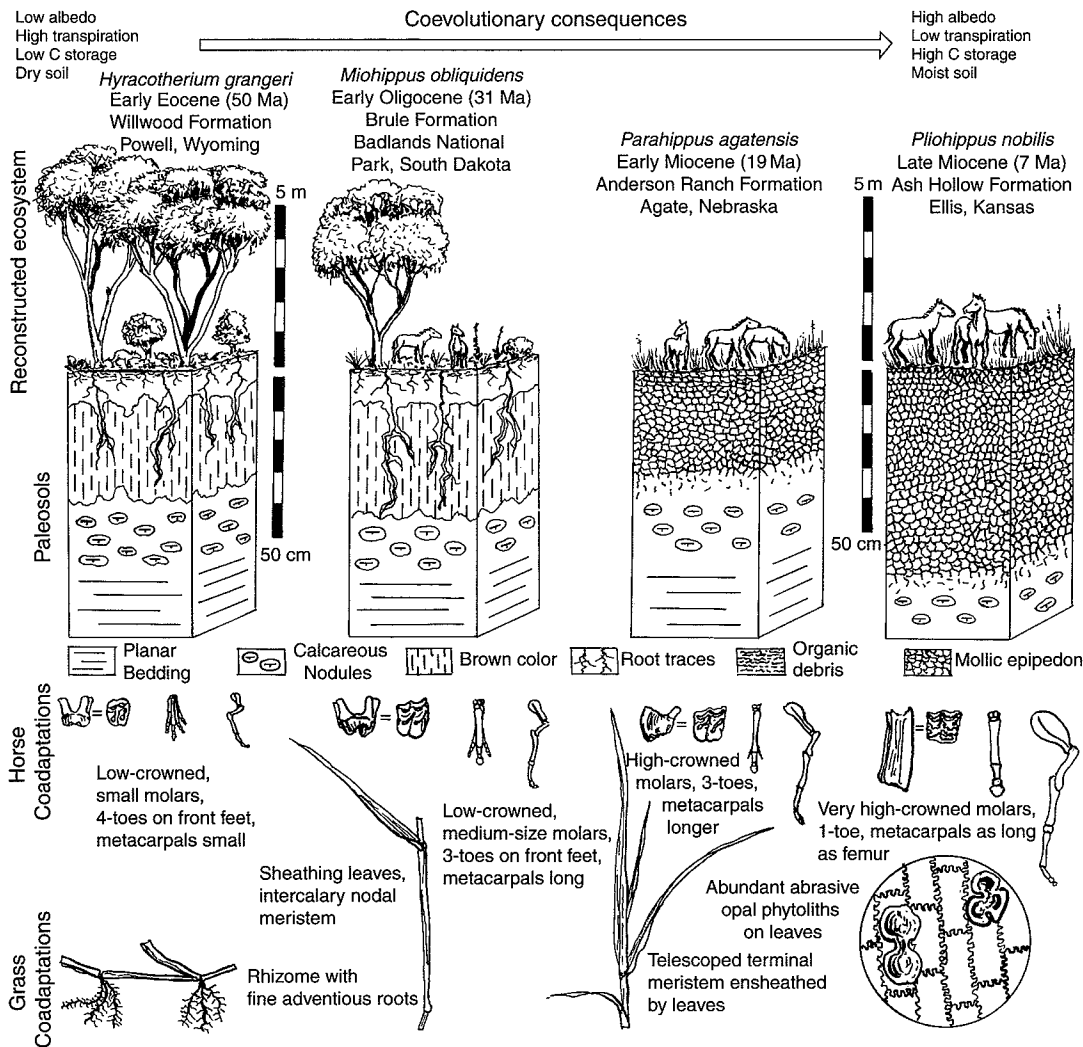


Figure 2 An example of coevolution, grasses and grazers, exemplified by paleosols and fossil horses from North America: grass evolutionary stages are less well constrained by the fossil record. Data from Retallack GJ (2001b) Cenozoic expansion of grasslands and global cooling. *Journal of Geology* 109: 407-426.

details of when and how this happened are shrouded in the mists of time. It is still uncertain, indeed largely hypothetical, when and how life first came to soils. There were changes in atmospheric oxidation and water salinity on the early Earth, but it remains uncertain which of these are due to soil colonization, as opposed to oceanographic, volcanic, or tectonic changes. Fortunately, there is a remarkable record of paleosols in very ancient rocks on Earth land (Rye and Holland, 1998) and on different planetary bodies (Retallack, 2001a), which furnish evidence on past conditions of climate and life. This supplements the geological record of life and environments in marine, volcanic, and metamorphic rocks (Schopf, 1983; Schopf and Klein, 1991). Paleosol development and geochemistry also provide relevant tests for competing hypotheses of the development of continents and plate tectonics.

9.11.2 Barren Worlds as Null Hypotheses

Uniqueness of our planet is best considered by comparison with other planetary bodies, which turn out to have a variety of common features, such as basalt and CO₂-rich atmosphere. Water beyond Earth is frozen or vaporized, with the possible exception of Jupiter's moon Europa (Marion *et al.*, 2005). Impact glasses, shocked minerals, clays, evaporites, carbonates, oxides, and a variety of high-temperature minerals also are found beyond Earth (Squyres *et al.*, 2004). Also found beyond Earth are rugged desert-like landscapes, volcanos, and ice caps (Carr, 1981). Such pervasive similarities to Earth make the search for biosignatures and their coevolutionary effects difficult, but before turning to that topic, consider some other worlds currently considered barren of life as null hypotheses.

9.11.2.1 Moon

At a small fraction of Earth mass (0.01), our Moon has insufficient gravitational pull to retain an atmosphere. Its temperature fluctuations of 111°C to -171°C also rule out liquid water (Taylor, 1982). Its soil appears waterless (Campbell *et al.*, 2006), despite past proposals of small areas of polar ice (Feldman *et al.*, 1998). The main soil-forming process on the moon is micrometeoroid bombardment, which enriches basaltic breccias and rocks exposed at the surface in meteoritic iron-nickel, impact glass, agglutinates of glass and minerals, and shattered mineral

grains (Figure 1). Well developed lunar soils 5–10 cm thick take some 200 million years to form, judging from rate of micrometeoroid bombardment (McKay and Basu, 1983) and the succession of paleosols in the radiometrically dated Apollo 15 core (Heiken *et al.*, 1976). The most profound soil-forming event in the history of the Moon was the footprints, tracks, and excavations of Apollo astronauts.

9.11.2.2 Venus

Venus is slightly smaller (0.8) than the mass of Earth, but its thick (93 bar) atmosphere is a toxic mix of greenhouse gases (97% CO₂) creating surface temperatures of about 470°C. Water is not stable in the soil and water vapor is rare (0.1–0.4%) in the atmosphere. Soil formation on the basaltic parent material may be a rapid process akin to pottery glazing (Figure 1), in which salts are melted to obscure the hard edges of rocks and sediments (Barsukov *et al.*, 1982).

9.11.2.3 Mars

Venus and our Moon stretch concepts of soil formation almost to breaking point, but Mars has more familiar ground. Mars is small (0.1 Earth mass), with a thin (0.008 bar) atmosphere, mainly CO₂ (95%), but with small amounts of N₂ (3%) and Ar (2%). Trace amounts (up to 0.003%) of CH₄ show temporal and geographic fluctuation suggestive of higher local concentrations (Formisano *et al.*, 2004). The Viking lander at Chryse Planitia recorded temperatures from -90°C at dawn to -30°C at noon (Carr, 1981). Martian soils are thin with smectite clays and sulfate minerals (gypsum, kieserite) which indicate that basalt and komatiite parent materials have been altered in aqueous solution by hydrolysis (incongruent dissolution in carbonic acid). This kind of soil formation is slight compared with millions of years for similar soil formation in the Dry Valleys of Antarctica (Campbell and Claridge, 1987) or the Atacama Desert of Chile (Navarro-González *et al.*, 2003). Soil formation may be even slower on Mars, because known soils appear to be paleosols from at least 3.2 Ga, formed at a time of more plentiful water at the surface indicated by outflow channels, then subsequently frozen at the surface during cold and dry subsequent eons (Retallack, 2001a). The red color of Martian soils is a very thin (<10 μm) rind, probably created by long-term surficial photooxidation (Yen *et al.*, 2000). Mars has a stunning array of

familiar landscapes including large volcanos, deep canyons, impact craters surrounded by solifluction flows, eolian dunes, and evaporite basins (Squyres *et al.*, 2004). Life-like microstructures, organic chemicals, magnetite, and carbonates from a Martian meteorite found in Antarctica (McKay *et al.*, 1996) raised hopes as microfossils, implying life on Mars as recently as 3.9 Ga. Although these remains are suggestive, they are not compelling, and Mars is widely considered sterile, now and in the geological past (Schopf, 1999).

9.11.2.4 Asteroids

The soils of asteroids have only been visited recently (Veverka *et al.*, 2001) and are brecciated by impact like lunar soils and some rare kinds of meteorites (mesosiderites and howardites; Bunch, 1975). Very different asteroidal soils are represented by carbonaceous chondrite meteorites, which show no evidence of living creatures, but include complex carbonaceous molecules, such as amino acids, dicarboxylic acids, sulfonic acids, and aliphatic and aromatic hydrocarbons (Pizzarello *et al.*, 2001). These carbonaceous compounds are bound up within fine-grained phases of smectite and other clay minerals formed at low temperatures (Tomeoka and Busek, 1988) in contrast to high-temperature pyroxene, olivine, and other high-temperature minerals of the chondrules (Itoh and Yurimoto, 2003). The clays form gradational contacts like weathering rinds, pseudomorphic replacement of chondrules, concentric cavity linings, and birefringence fabrics of oriented clay like those formed in soils under low confining pressures (Bunch and Chang, 1980). Other evidence of hydrous alteration are framboids and plaquettes of magnetite (Kerridge *et al.*, 1979). There also are cross-cutting veins of calcite, siderite, gypsum, epsomite, and halite (Richardson, 1978).

Clay and salts are common products of hydrolytic weathering in soils of Earth, and can thus be inferred for the parent bodies of carbonaceous chondrites, which are thought to have been small (<500 km diameter) planetesimals. High-temperature chondrules of carbonaceous chondrites are 4.56 Ga in age (Amelin *et al.*, 2002), and the plaquettes of magnetite (Lewis and Anders, 1975), and veins of calcite (Endress *et al.*, 1996) and halite (Whitby *et al.*, 2000) no younger than 4.51 Ga. Degassing of CO₂ and H₂O vapor from planetesimals early in the accretion of the solar system resulted in surprisingly Earth-like weathering (Figure 2).

This account has stressed a weathering interpretation of carbonaceous chondrites, but some of these also include talc and other minerals suggestive of high-temperature (250–300°C) hydrothermal alteration (Brearley, 1997). There is also a school of thought that the clay and organic matter of carbonaceous chondrites are not alteration products, but instead cold condensates from the original nebula from which the solar system formed (Wasson, 1985).

9.11.3 Biosignatures of Earth

Several unique features of Earth compared with other planetary bodies can be considered biosignatures: features indicative of life. Some of these such as inflammable atmosphere, permanent liquid water, carbon-based life, and civilizations feature prominently in the search for life elsewhere in the universe. Biosignatures of Earth's soils include calcic, argillic, and mollic horizons. Salic horizons are known from Mars and carbonaceous chondrites. Carbon-rich surface horizons may be represented by carbonaceous chondrites. It is not yet clear whether deep hydrolytic weathering occurs on other planetary bodies (Squyres *et al.*, 2004). Other features such as convex slopes, spheroidal deep weathering, and granitic rocks unique to Earth are less obviously related to life's influence, but nevertheless worth consideration.

9.11.3.1 Inflammable Atmosphere

Our atmosphere is a combustible mix of oxygen (21%) and methane (0.00017%), with its reacted product carbon dioxide (preindustrial 0.028%, 0.037% in 2001, and rising), diluted with much inert nitrogen (78%). It is thus more like input gases of a carburettor than the exhaust gases of an automobile tailpipe, and quite distinct from the CO₂-rich composition of other planetary atmospheres and volcanic gases (Lovelock, 2000). Burning of other flammable substances such as hydrocarbons and wood in the open air is thus possible on our planet, unlike other planetary bodies. More than trace amounts of oxygen in the atmosphere dates back at least 2.3 Ga (Bekker *et al.*, 2004). Entropy reduction or chemical disequilibrium of planetary atmospheric composition is a way of identifying life elsewhere in the universe, and also an argument for the Gaia hypothesis of life's control of atmospheric composition (Lovelock, 2000).

Table 1 Metabolic processes of geological significance

I. Fermentation						
$C_6H_{12}O_6$	\rightarrow	$2C_2H_5OH$	+	CO_2	+	energy
Sugar		Alcohol		Carbon dioxide		
II. Respiration						
$C_6H_{12}O_6$	+	O_2	\rightarrow	$6CO_2$	+	$6H_2O$ + energy
Sugar		Oxygen		Carbon dioxide		Water
III. Anaerobic photosynthesis						
$6CO_2$	+	$12H_2S$	$\xrightarrow{\text{light}}$	$C_6H_{12}O_6$	+	H_2O + $12S$
Carbon dioxide		Hydrogen sulfide		Bacteriophyll Sugar		Water Sulfur
IV. Aerobic photosynthesis						
$6CO_2$	+	$12H_2O$	$\xrightarrow{\text{light}}$	$C_6H_{12}O_6$	+	$6H_2O$ + $6O_2$
Carbon dioxide		Water		Chlorophyll Sugar		Water Oxygen

The reason for the cosmically peculiar composition of our atmosphere may be the widespread metabolic process of photosynthesis (Rosing *et al.*, 2006) not only by plants, and a variety of protists, but also by a host of bacteria as well. Photosynthesis, or assembly by light, is a process by which sugars are synthesized using light from the Sun and a catalyst (commonly chlorophyll) from CO_2 of the atmosphere (Table 1). The low amount of CO_2 in our current atmosphere is a testament to the success of photosynthesis, but replacement of a primordial reducing atmosphere of CO_2 and CH_4 by O_2 had many ups and downs over 3.5 Ga of Earth's history known from the rock record (Berner *et al.*, 2000).

9.11.3.2 Permanent Liquid Water

Our pale dot of a planet, blue with oceans and wreathed with water vapor, is unique in our solar system now, and also in its deep history of water at the surface (Lovelock, 2000). On a cosmic scale of temperatures from $-270^\circ C$ (2.76 K) in deep space to about $6000^\circ C$ for the surface of the Sun, the 0– $100^\circ C$ range of liquid water is a narrow range not currently found on any neighboring planetary bodies, though perhaps present within Europa (Marion *et al.*, 2005), and on Mars in the distant geological past, some 2.5–3.5 Ga (Carr, 1981). It did not last on Mars, but on Earth there is a continuous record of water-lain sedimentary rocks and precipitates from 3.5 Ga to the present (Allwood *et al.*, 2006). Oxygen-isotopic composition of detrital zircons individually dated to

4.3 Ga in the Jack Hills of Western Australia are unusually heavy (15.3 vs typical mantle values of 5.4‰ $\delta^{18}O$ vs SMOW) and suggest free surficial water at the time their parent granitoids cooled only 200 Ma after the formation of Earth (Mojzsis *et al.*, 2001). This is a long and continuous history of surficial water despite degassing of largely greenhouse gases to a thick (1 bar) atmosphere and a 30% increase in solar luminosity due to stellar evolution over the same period. Permanent water and its longevity through planetary history is a clue to life on other planets, as well as an argument for Gaian environmental regulation (Lovelock, 2000).

The reason for the maintenance of Earth's temperature within the bounds of liquid water for some 4.3 Ga remains uncertain, but life is suspected to have a role because of its coevolved metabolic systems. This general idea of a thermocouple of opposing forces acting as a thermostat is clearly expressed in 'Daisyworld' albedo models of the Gaia hypothesis (Watson and Lovelock, 1983). Imagine a world with only black and white daisies vying for the light of the Sun in order to photosynthesize. White daisies cool by reflecting sunlight back into space. Black daisies warm by absorbing heat. Populations will be dominated alternately by black then white daisies, until mixed populations converge on temperatures that optimize photosynthesis. A more realistic model is the 'Proserpina principle' (Retallack, 2004a), which postulates that photosynthesis cools the planet by drawing down the greenhouse gas CO_2 , whereas respiration warms the planet because it draws down O_2 . If either one of these metabolic

systems had evolved in isolation it would have resulted in a respirator's hell like Venus, or a photosynthesizer's freezer like Mars. Fortunately, photosynthesizers and respirers such as plants and animals are mutually interdependent for food and breath: a coevolutionary thermostat. Plants cool the planet by photosynthesis which is curtailed as they are covered by snow of icehouse atmospheres. Animals warm the planet by respiration but die in high temperatures of greenhouse atmospheres that are less fatal to plants. Population balances between them have the effect of adjusting greenhouse gases in the atmosphere to habitable ranges given external inputs of solar radiation.

9.11.3.3 Carbon-Based Life

Science fiction movies make alien life instantly recognizable by emphasizing the salient features of humans or insects. Recognizing alien life on sight may not be so easy because much life is immobile or microscopic, and its activities too varied to be easily characterized (Table 2). Life on Earth has a marked preference for six elements: carbon, hydrogen, oxygen, nitrogen, phosphorus, and sulfur (CHONPS: Schoonen *et al.*, 2004). Of these only oxygen is a common component of most rocks. Three general features are regarded as necessary for life: bodies, metabolism, and reproduction. Even bodies of such simple organisms as the common gut bacterium *Escherichia coli* have an astonishing complexity of interacting parts, including high-molecular-weight organic compounds which provide both structure and function (Figure 3). If there is a common theme to the great variety of metabolic reactions, it is lack of chemical equilibrium that drives them and that has left an imprint on our cosmically peculiar atmosphere (Lovelock, 2000). Reproduction is also quite varied, ranging from simple cell division, to clonal budding, parthenogenesis, and the romantic complexity of sex. If there is a common theme to biological reproduction it is remarkable speed. Astonishing numbers of individuals can be produced in days or years as long as food is available. Whole worlds can be populated in a geological instant, which should be encouraging to the search for life on other worlds. If life is anywhere on a planet, it is likely to be everywhere.

The difficulties of recognizing life are well illustrated by debate over evidence of life on the early Earth and a Martian meteorite. Very ancient photosynthetic activity is suggested by the carbon-isotopic ratios of amorphous organic matter in the 3.8 Ga Isua

Greenstone Belt of west Greenland (Rosing and Frei, 2004), although the high metamorphic grade and discernable modern organic contamination are concerns (Nishizawa *et al.*, 2005). The oldest microfossil evidence for life has long been considered chains of cyanobacteria-like carbonaceous cells in chert from the 3.5 Ga Apex Chert of the Warrawoona Group near Marble Bar, Western Australia (Schopf, 1983). The poor preservation and hydrothermal setting of these fossils has recently been urged as evidence that they are not microfossils, but inorganic carbonaceous material (Brasier *et al.*, 2002). In my opinion the distortion and poor preservation of the microfossils, and especially the rounded cavities in them (Schopf *et al.*, 2002), are evidence of both cyanobacterial primary production and actinobacterial decay. Carbonaceous and mineral pseudofossils have greater regularity of construction and resistance to degradation than the Warrawoona microfossils (Garcia-Ruiz *et al.*, 2003). In any case, alternative evidence of life from microbial etching of pillow lavas of the Barberton Greenstone Belt of South Africa is also dated to about 3.5 Ga (Furnes *et al.*, 2004).

The Martian meteorite (ALH84001) with putative microfossils recovered from ice near Allan Hills, Antarctica (McKay *et al.*, 1996) is of an unusual type (SNC) thought to have been derived from Mars because of its igneous (nonchondritic) microtexture, geological age younger than the 4.6 Ga age of most meteorites, unusual oxygen-isotopic composition, lack of magnetic paleointensity, and evidence for long exposure to interplanetary cosmic rays (Wood and Ashwal, 1981). Meteorite ALH84001 ground-mass is 4.5 Ga in age, but was shock metamorphosed at 4.0 Ga, then infiltrated by secondary carbonate with putative microfossils at 3.9 Ma (Borg *et al.*, 1999). It was lofted from Mars by about 16 Ma and orbited the Sun until falling into Antarctic ice at about 13 Ka (McKay *et al.*, 1996). The putative microfossils are cylindrical shapes arranged end to end in short strings, with associated polycyclic aromatic hydrocarbons and lozenges of magnetite comparable with magnetotactic bacteria and nannobacteria on Earth (Folk and Lynch, 1997). Unfortunately, the cell-like structures are too small (0.02–0.1) to house the complex molecular machinery of even the simplest known bacteria (Figure 3), and are more likely organic coagulates or artefacts of scanning electron microscopy (Schieber and Arnott, 2003). The organic compounds have isotopic composition (Jull *et al.*, 1998) and racemization like Earthly contaminants (Bada *et al.*, 1996). The magnetite has nonbiological

Table 2 Kinds of microbes, their metabolic requirements and role

General role	Kind of organism	Example genus	Energy source	Electron donor	Carbon source	Oxygen relations	Comments	
Carbon cyclers	Algae	<i>Chlamydomonas</i>	Sunlight (P)	H ₂ O (L)	CO ₂ (A)	Aerobic	Primary producer normal-wet soils, ponds, and ocean	
	Cyanobacteria	<i>Nostoc</i>	Sunlight (P)	H ₂ O (L)	CO ₂ (A)	Aerobic	Primary producer normal-wet soils	
Nitrogen cyclers	Purple non-sulfur bacteria	<i>Rhodospirillum</i>	Sunlight (P) sometimes organic compounds (C)	Organic compounds (O) sometimes H ₂ S (L)	Organic compounds (H) sometimes CO ₂ (A)	Amphibiaerobic	Primary producer in swamps and stagnant ponds and ocean	
	Methanogenic bacteria	<i>Methanobacterium</i>	H ₂ (C)	H ₂ (L)	CO ₂ (A) sometimes formate (H)	Anaerobic	Creators of 'swamp gas' (CH ₄) in stagnant ponds and ocean	
	Aerobic spore-forming bacteria	<i>Bacillus</i>	Organic compounds (C)	Organic compounds (O)	Organic compounds (H)	Aerobic	Decomposer of organic compounds soils, lakes, and oceans	
	Fermenting bacteria	<i>Clostridium</i>	Organic compounds (C)	Organic compounds (O)	Organic compounds (H)	Anaerobic	Decomposer in swamp soils, lakes, and oceans	
	Protoctistans	<i>Amoeba</i>	Organic compounds (C)	Organic compounds (O)	Organic compounds (H)	Aerobic	Predator in soils, lakes, and oceans	
	Nitrogen-fixing bacteria	<i>Azotobacter</i>	Organic compounds (C)	N ₂ (L)	Organic compounds (H)	Aerobic	Creates ammonium (NH ₄ ⁺) in soils, lakes, and oceans	
	Root nodule bacterioids	<i>Rhizobium</i>	Organic compounds (C)	N ₂ (L)	Organic compounds (H)	Aerobic	Supplies ammonium (NH ₄ ⁺) to host plant in well-drained soils	
	Denitrifying bacteria	<i>Pseudomonas</i>	NO ₂ (C) sometimes organic compounds (C)	N ₂ (L) sometimes organic compounds (O)	CO ₂ (A) sometimes organic compounds (O)	Anaerobic	Releases nitrogen (N ₂) to atmosphere from swampy soils	
	Sulfur cyclers	Sulfur-reducing bacteria	<i>Desulfovibrio</i>	Organic compounds (C)	SO ₄ ²⁻ (L)	Organic compounds (H)	Anaerobic	Creates 'rotten egg gas' (H ₂ S) and pyrite in swamps, lakes, and oceans
		Sulfur bacteria	<i>Chromatium</i>	Sunlight (P)	H ₂ S or S (L) sometimes organic compounds (O)	CO ₂ (A) sometimes organic compounds (H)	Amphibiaerobic	Remobilizes sulfur in poorly oxygenated soils, lakes, and oceans
Sulfur metabolising bacteria	<i>Thiobacillus</i>	S, FeS (C)	S, SO ₄ ²⁻ , Fe ₂ O ₃ (L)	CO ₂ (A)	Aerobic	Remobilizes sulfur in wet-normal soils, lakes, and oceans		

Initials for trophic groups are autotrophic (A), chemotrophic (C), heterotrophic (H), lithotrophic (L), organotrophic (O), phototrophic (P). From Pelczar MJ, Chan ECS, Krieg NR, and Pelczar ME (1986) *Microbiology*, 918 pp. New York: McGraw-Hill.

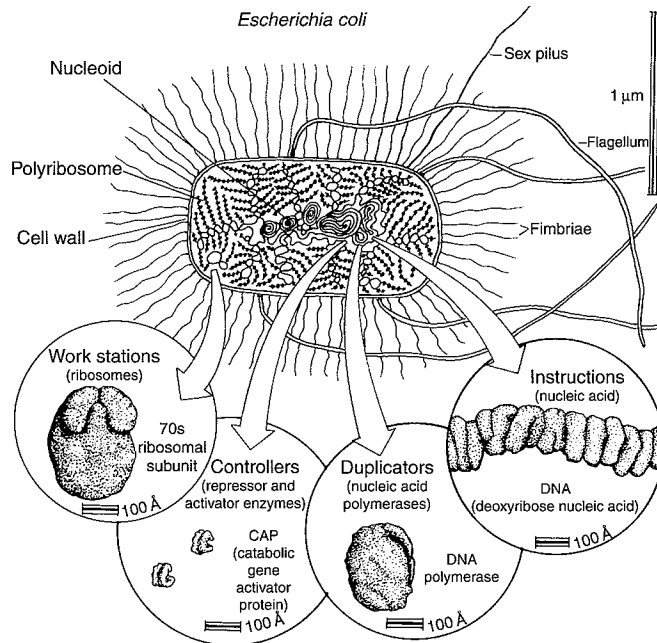


Figure 3 *Escherichia coli*, a common gut bacterium, considered as an organic machine of astonishing complexity. From Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

screw dislocations and the carbonate may have formed at high temperatures (McSween, 1997). Life on Mars thus remains unproven.

9.11.3.4 Spheroidal Weathering

Spheroidal weathering is the deep weathering of rocks of uniform composition, especially granitic rocks, to create rounded concentric rings of alteration inward from deep cracks (Figure 4(a)). Onion-skin

weathering is a particularly aggressive form of spheroidal weathering in which the whole rock has been altered to concentric shells. More common is a few shells of alteration around a rounded remnant of little altered rock in the center, a corestone. Rounded corestones resist erosion more strongly than soft surrounding rock and overlying soil profile (Strudley *et al.*, 2006), and are exhumed as distinctive rounded boulders called tors (Figure 4(b)). No landforms or deep-weathering of these kinds have yet been found

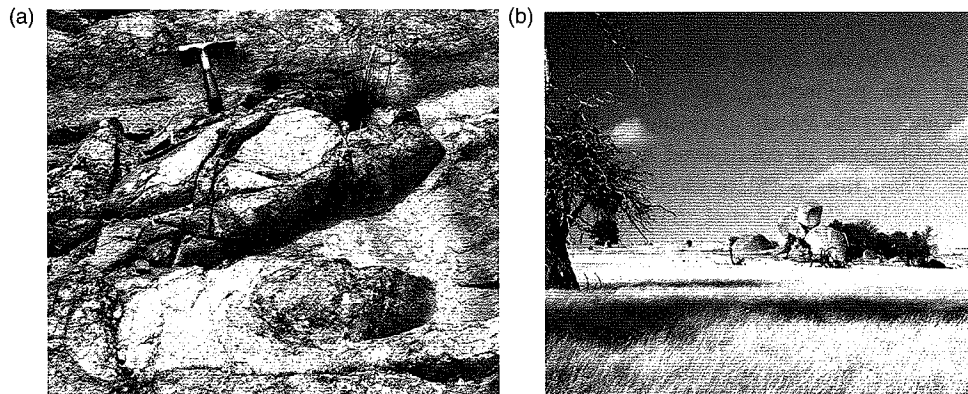


Figure 4 Spheroidal weathering and tors. (a) Paleoproterozoic (2.3 Ga) spheroidally weathered corestones in a paleosol developed on pink alkali granite and covered Huronian Supergroup conglomerates, near Elliot Lake, Ontario, Canada (hammer gives scale). (b) Tors of Baynton Granite exhumed from Miocene paleosol, 4 km north of Pyalong, Victoria, Australia (tree is 5 m tall).

on the Moon, Venus, Mars, or asteroids, but this kind of deep weathering is common in deeply weathered Precambrian paleosols as old as 2.3 Ga (Figure 4(a)). If spheroidal weathering has such geological antiquity one would expect tors to be ancient landforms as well, but the oldest likely documented case known to me is on the pre-Stoer (1.2 Ga) paleosol near Stoer, northwest Scotland (Stewart, 2002).

Spheroidal weathering and exhumed tors require deep weathering by acidic solutions which are promoted by abundance of water and warmth (Taylor and Eggleton, 2001). Strongly acidic solutions of the sort that form during oxidation of pyritic mine spoils are not indicated, because these form intense localized plumes of alteration. More likely the acid involved in spheroidal weathering is carbonic acid formed from dissolution of CO₂ in soil water exploiting the network of cracks in the rock due to expansion by thermal stresses or following unloading of overburden (Retallack, 2001a). Warm, wet carbonic solutions should be feasible on other planetary bodies, but on Earth they are greatly promoted by life. Soil respiration, especially of microbes deep in soil profiles, creates partial pressures of CO₂ up to 110 times that of atmospheric pressures of CO₂ on Earth (Brook *et al.*, 1983). Texturally uniform rocks, such as anorthosites on the Moon and komatiites on Mars, are produced on other planetary bodies, though not granitic rocks (Campbell and Taylor, 1983). Thus it remains to be determined for sure whether spheroidal weathering and tors are truly biosignatures.

9.11.3.5 Granitic Rocks

Granitic rocks are unique to Earth, and not yet discovered on other planetary bodies or meteorites (Campbell and Taylor, 1983). Granitic rocks are ancient but not primordial, with granodiorites appearing at about 3 Ga among more abundant tonalites and trondjemites of continental crust (Engel *et al.*, 1974). They thus postdate evidence for life (Schopf, 1983), water (Mojzsis *et al.*, 2001), and deep hydrolytic weathering on Earth (Buick *et al.*, 1995), so it is plausible, though far from certain, that granites are biosignature rocks.

It may seem outrageous to suggest that granitic rocks forming under high temperatures (600–800°C) at depths of 2–10 km within continental crust could be in any way related to life, but additional features beyond uniqueness to Earth and geological antiquity suggest a role for life in granitization (Rosing *et al.*,

2006). Granites are low-density, hydrated, aluminous and siliceous compared with other igneous rocks, and these are all qualities imparted by deep hydrolytic weathering known to have been widespread on Earth back some 3.5 Ga (Maynard, 1992). Light rare-earth enrichment, including preference for neodymium over samarium, is also a geochemical similarity of granitization and weathering (MacFarlane *et al.*, 1994). The 'Daly Gap', or rarity of rocks intermediate in composition between basalt and rhyodacite, in many continental volcanic regions points to different pathways of genesis (Bonnetoi *et al.*, 1995). Migmatites and S-type granites include melted sedimentary rocks derived from continental weathering, judging from their elevated ⁸⁶Sr/⁸⁷Sr ratios, oxygen-isotope composition, and alumina content, as well as minerals such as corundum, muscovite, garnet, and cordierite (Rosing *et al.*, 2006). By this view, life's role in forming granite begins with its promotion of deep weathering and surficial liquid water. Some of this low-density, aluminous and siliceous soil and sediment would be melted upon deep burial by overlying rocks, or entrained by the foundering of dense basaltic slabs in plate tectonic recycling to form a new kind of rock, granite. From this perspective it is also plausible that plate tectonics driven by density differences and surficial water leading to granitization is also a biosignature of Earth. There is no plate tectonics on the Moon (Taylor, 1982), and it seems unlikely on Venus, which has some poorly understood mechanism of crustal renewal (Fowler and O'Brien, 2003), but early Mars may have had seafloor spreading (Nimmo and Tanaka, 2005).

A role for life and weathering in granitization is rarely considered in igneous petrology. The prevailing view of granite formation is by magmatic processes, such as fractional crystallization in which mafic minerals drop out of molten magmas to form dense residues leaving more silicic fluids to form granitic rocks (Rudnick, 1995). Remarkably, Bowen's (1922) famous reaction series, which defines the order in which specific minerals crystallize from silicate melt, is the same order in which they are destroyed in deeply weathered soils, as determined by Goldich (1938). Olivine and pyroxene are early to crystallize in magmas and first to be weathered, in contrast to quartz which is last to crystallize and weather away.

9.11.3.6 Convex Slope

Concave and straight slopes are common in desert badlands and alpine glacial terrains on Earth, the

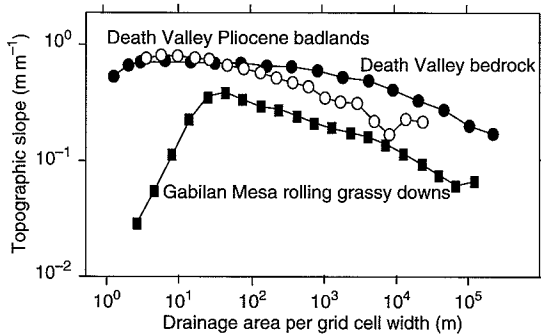


Figure 5 Slope–area relations in soil-mantled compared with desert bedrock and badlands in landscape areas of 0.24–5.8 km² in California. Drainage area per grid cell is the size of individual catchments within successively larger map squares beginning with 2 m squares. From Dietrich WE and Taylor PJ (2006) The search for a topographic signature of life. *Nature* 439: 411–418.

Moon, and Mars. Broadly convex slopes on the other hand may be unique to soil-mantled upland topography on Earth (Dietrich and Taylor, 2006). The effect is subtle in topographic profiles, especially those with low topographic relief, but can be striking in slope–area plots of small areas (<6 km²) compiled from high-resolution digital elevation models (Figure 5). Soil-mantled terrain has low slopes around upland small drainage basins but steep slopes in large valleys, whereas badlands and mountains show comparably steep slopes in both small and large drainage basins. Put another way, landscapes at large scale and with limited life are almost fractal in distribution of slope with catchment size, but soil-mantled landscapes are nonfractal in small catchments. This potential biosignature deserves further attention through image analysis of other planetary bodies and paleosols.

Concave slopes are created by fluvial or other fluid erosion, by Gilbert's (1877) 'law of divides' in which slope increases with restricted discharge upstream. Straight slopes on the other hand are the angle of repose of scree and glide planes of other forms of mass movement, as first noted by Penck (1953). Broadly convex slopes in contrast are associated with thick soil mantles. Soil communities bind the earth on which they depend while smoothing with bioturbation irregularities in the overall expansion of parent materials due to weathering and cracking. Mathematically this is diffusive dilation. Soil stabilized by life creeps downslope until failure angles are reached to create broad convex slopes above oversteepened straight to concave slopes. In contrast, desert badlands and alpine glacial terrains

have narrowly convex divides, steep headwall drainages, and long straight-to-concave slopes.

9.11.3.7 Calcic Horizon

A calcic horizon is a particular level within a soil studded with nodules of calcite or dolomite (Figure 6(a)), and is especially common in desert soils (Aridisols) on Earth. Despite the desert-like images of the Moon (Taylor, 1982), Venus (Barsukov *et al.*, 1982), and Mars (Squyres *et al.*, 2004), no pedogenic carbonate has yet been detected beyond Earth. Spectroscopic detection of carbonate minerals in Martian dust (Bandfield *et al.*, 2003) may have the same source as carbonate in Martian meteorites and carbonaceous chondrites, which are vein-filling crystalline carbonate, some of it formed at high temperatures (McSween, 1997). Fossil and modern pedogenic carbonate in contrast is micritic and replacive in microtexture (Retallack, 1991).

Many aspects of calcic horizons are controlled by their soil ecosystems. Calcareous rhizoconcretions and termitaries are biogenic structures modeled on roots and termite nests, respectively (Retallack, 2001a). Micritic replacement within nodules is microbially mediated (Monger *et al.*, 1991). The depth to calcic horizons in soils is strongly correlated with mean annual precipitation and soil respiration (Retallack, 2005). Calcic horizons are found largely within the mean annual precipitation range of 300–1000 mm, as shown particularly well in transects from the hyper-arid Atacama Desert to higher-elevation shrubland and grassland (Rech *et al.*, 2006). The microbe-poor Atacama Desert soils have gypsic horizons of soluble salts, like soils of Mars (Figure 1) and the Dry Valleys of Antarctica (Campbell and Claridge, 1987). Calcic horizons are found in soils of desert shrublands largely leached of such soluble salts, which inhibit biological activity with high osmotic stress.

9.11.3.8 Argillic Horizon

An argillic horizon is a level within a soil markedly enriched in clay compared with horizons above and below (Figure 6(b)), and is a hallmark of forested soils such as Alfisols and Ultisols. The argillic horizon is defined on the basis of amounts of clay enrichment (roughly 10% more, with exact definition texture dependent; Soil Survey Staff (2000)), but also has characteristic microstructure of extremely fine-grained and high-birefringence wisps of pedogenic clay along with inherited clay (Retallack, 2001a). These kinds of

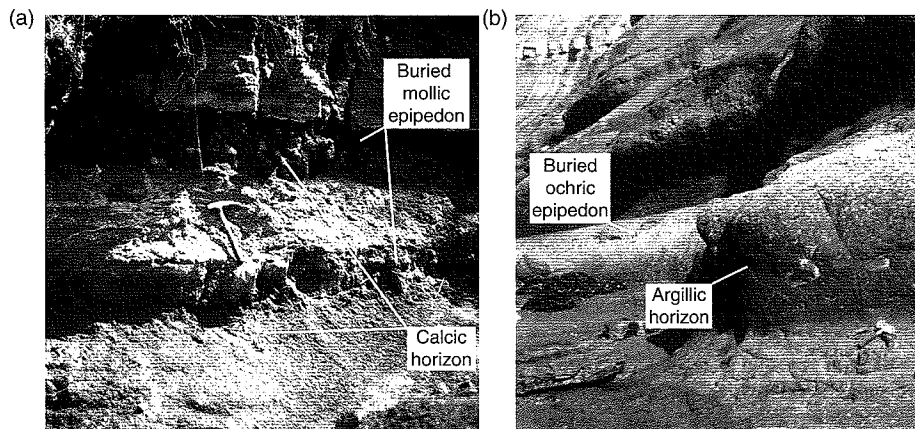


Figure 6 Calcic and argillic horizons and mollic epipedon in paleosols. (a) Two Chogo clay paleosols (Calciustolls) showing mollic epipedon (dark surface) and calcic horizon (white nodular bands), both profiles truncated by nephelinitic granular ash in the Middle Miocene Fort Ternan Member of the Kericho Phonolites, near Fort Ternan, Kenya (Retallack, 1991). (b) Long Reef clay paleosol (Paleudult) showing argillic horizon (dark red clay-enriched) beneath light-colored surface horizon covered by clayey alluvium in the late Early Triassic Bald Hill Claystone, near Long Reef, New South Wales, Australia (Retallack, 1997b). Hammers give scale in both frames.

horizons have not been found on the Moon, Mars, and Venus, and are not found in paleosols on the early Earth either. The geologically oldest known example of an argillic horizon is from Middle Devonian (Givetian, 390 Ma) Aztec Sandstone of Victoria Land, Antarctica (Retallack, 1997a). Argillic paleosols and other evidence of ancient forests are common in every subsequent geological period (Retallack, 2001a).

Both the micromorphology and geological history of argillic horizons suggest that they are products of forest ecosystems, known independently from fossil stumps and wood to have originated by Middle Devonian (390 Ma). Their micromorphology of wispy to laminated very fine-grained clay suggests that they form from both washing down, and weathering in place around large tapering roots (Retallack, 2001a). During interplanetary exploration it is likely that forests would be encountered before argillic horizons, but relict argillic paleosols could be indicative of forests of the geological past, just as they are in some desert regions of Earth today.

9.11.3.9 Mollic Epipedon

The mollic epipedon is a unique surface horizon of grassland soils (Mollisols), which presents a number of paradoxes. It is rich in base-rich clay such as smectite yet does not ball up or shear into unwieldy clods and flakes. It is rich in organic matter yet well aerated with cracks and channels in which organic matter should be oxidized. It can be well drained yet still maintains soil

moisture. It is fertile with phosphorus and other mineral nutrients despite long periods (thousands of years) of soil development. Mollic epipedons achieve these highly desirable agricultural qualities through a unique structure of fine (2–3 mm) ellipsoidal (crumb) clods (peds), which consist of clay stabilized by organic matter. The geologically oldest fossilized mollic epipedons are Early Miocene (19 Ma), and known from the Anderson Ranch Formation, near Agate, Nebraska, and the upper John Day Formation, near Kimberly, Oregon, USA (Retallack, 2004b). Nothing like a mollic epipedon has been reported from the Moon, Venus, Mars, or meteorites.

The temporal range and micromorphology of mollic epipedons suggest that they are created by grassland ecosystems. Some of the rounded crumb peds are excrements of earthworms common in these soils; other peds are products of the three-dimensional network of the slender adventitious roots of sod-forming grasses (Retallack, 2001a). The organic matter is partly from root exudates and earthworm slimes, but cake-like dung of large ungulates also plays a role in soil conditioning. The mollic epipedon can be considered a trace fossil of sod-grassland ecosystems, even after destruction by desertification or burial (Retallack, 2004b).

9.11.3.10 Civilization

One would think the Great Wall of China as a human construction obvious from space. In poor-resolution

images it could be confused with another 'Great Wall of China', which is a tourist attraction south of Blinman in the northern Flinders Ranges, Australia, and a natural outcrop of Neoproterozoic Mount Caernarvon Greywacke flanking a diapiric uplift like a fortified city. The so-called 'Face of Elvis' and 'pyramids' on Mars were subsequently shown to be inselbergs and yardangs (Pieri, 1999). Similarly the canals of Mars were products of imagination working at the limits of optical resolution of the time (Zahnle, 2001). Close optical and physical surveillance will be needed to detect civilizations, even on planets with dead atmospheres and boiled oceans in which civilizations may once have failed. Successful civilizations on the other hand may come to us. The effort to detect radio communications from outer space continues with ever larger telescopes and more powerful computer power (Whitfield, 2003). Yet we remain alone and isolated in the universe, and know no failed civilizations either.

9.11.4 Origin of Life on Earth

From a coevolutionary perspective the origin of life is not the heroic struggle of an individual molecule or organism from a primordial soup (Nisbet and Sleep, 2001), but a system of supportive molecules or organisms for growth and replication in an environment that selected for some desirable property while

metering the supply of fundamental nutrients to last for geological timescales. To oversimplify in human terms, the first life needed a job, a family, a place to live, and a legacy for future generations. These theoretical requirements are made clear by Cairns-Smith (1971), who has argued that the precursors of organic life were clays which grew by assimilating cations from weathering solutions and reproduced by cracking into smaller pieces or flakes with diagnostic layering or cationic substitutions. These general characteristics would not coevolve in any lifelike way unless the clays also had some property subject to natural selection. He proposes the simple but important quality of stickiness. The layering and cationic substitution of smectite clays determines their interlayer expansion upon wetting, which can vary stickiness from a thin slurry to a thick paste. Stickiness becomes important because it can be selected to stabilize an environmental interface, and thus create a legacy for creation of more stickiness to maintain that interface in a critical zone of energy and materials flux.

Cairns-Smith thought in terms of a sandy aquifer, but I imagine a primordial soil of a planetesimal, in which minerals are weathered by carbonic acid solutions of degassed CO₂ and water, and warmed by a faint young Sun (Figure 7). Sloppy clay expands violently when wet and is then washed too deep into the soil for warmth and acid to produce copies. Tough clay on the other hand expands little and covers source

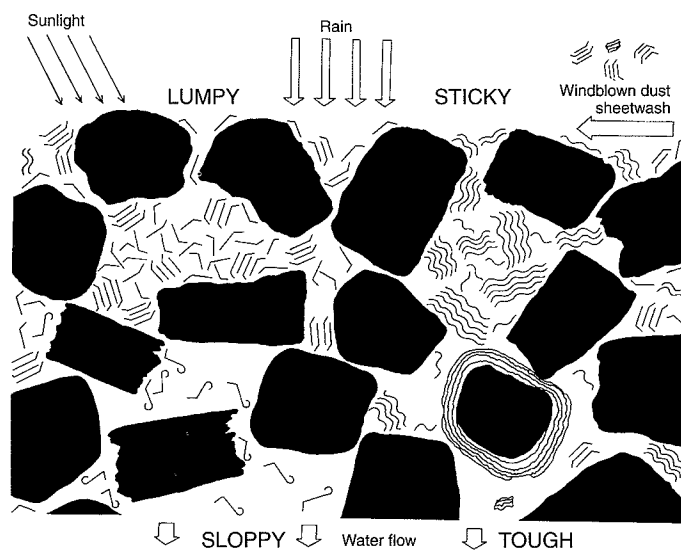


Figure 7 Hypothetical natural selection of four kinds of clay differing in stickiness (Sloppy, Lumpy, Sticky, and Tough) and weathering from the minerals of a primordial soil. From Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

grains in a thick rind so that the mineral is no longer available to water and acid to produce more clay. Lumpy and sticky clay in contrast has a consistency that expands on wetting sufficiently to bridge the gaps between grains, yet crack away from the source grains to allow access for water and acid to create more clay. These clays will be selected against the soil-eroding activities of wind and water because their grains are glued together by sticky clays. In an environment where loose sandy soils are eroded, all soils become more clayey, and by persisting produce more clay. Lumpy clays may be better adapted to heavy rains, and sticky clays to drizzle, but both are selected by their ability to hold the soil. They are in a coevolutionary escalation which refines their distribution, composition, and degree of stickiness for particular climatic conditions, because rivers erode the least sticky clays. Similarly the first organic molecules, produced by abiotic reactions could also be selected in soils for their stickiness. Like molasses on a cornfield, organic soils survive erosion to make more organic matter and create soils with observed fabrics of carbonaceous chondrites. The molecules of heredity, DNA and RNA are sticky too, and could have begun what Cairns-Smith calls a "genetic takeover" of the older and cruder clay system of soil stabilization. Cells, trees, and civilizations continue to depend on the stability and bounty of soil and its nutrients. This simple parable makes clear the importance of natural selection by the environment in converting the long odds of chance evolution of complex life to a necessity.

9.11.4.1 Experimental Studies

Laboratory experiments have succeeded in demonstrating the ease with which so-called 'organic molecules' are created abiotically. Miller's (1953) classical experiment sparking ammonia and hydrogen gas above water in a sealed apparatus produced amino acids and sugars, common building blocks of life as we know it on Earth. These experiments have since been repeated with a variety of gas mixtures, with lower but still significant yields of organic matter in CO₂-rich mixtures (Schopf, 1983). Yields are promoted by addition of clays, and chemically reduced minerals such as pyrite and siderite (Schoonen *et al.*, 2004). Clays also promote assembly of oligomers of RNA up to 50 units long (Ertem and Ferris, 1996). Clays act as catalysts, templates, compartments, energy-storage devices, and chemostats in promoting organic molecule synthesis. Siderite can mediate photooxidation of CO₂ to produce reduced

carbon in a way comparable with photosynthesis (Braterman *et al.*, 1983). Concentrated amino acids (proteinoids) can form spherical structures (Fox and Dose, 1972), which appear superficially like living cells, but are far from functioning cells (Figure 3). Nothing close to a living cell has ever emerged from experiments designed to examine the origin of life, so such experimental research into the origin of life has reinforced the enormous gulf between quick and dead organic matter.

9.11.4.2 Geological Records

Carbonaceous chondrites, with their mix of smectite clay and diversity of reduced organic compounds (Pizzarello *et al.*, 2001), are evidence for abiotic synthesis of organic compounds in the presence of clay and reduced iron minerals such as siderite early during the origin of the solar system. The carbonaceous clayey matrix postdates olivine chondrules which are 4.56 Ga in age (Amelin *et al.*, 2002) and predates cross-cutting veins of calcite (Endress *et al.*, 1996) and halite (Whitby *et al.*, 2000) no younger than 4.51 Ga. The characteristic carbon-isotopic fraction of rubisco, an important enzyme in photosynthesis, is indicated by light carbon-isotopic ratios of amorphous organic matter in the 3.8 Ga Isua Greenstone Belt of west Greenland (Rosing and Frei, 2004) and microfossils comparable with cyanobacteria are found in chert from the 3.5 Ga Apex Chert of the Warrawoona Group near Marble Bar, Western Australia (Schopf, 1983). Thus the origin of life occurred during the 700 My interval between 4.5 and 3.8 Ma, or approximately the first 15% of the history of our planet. This period is a dark age in Earth history, its sedimentary records obliterated by metamorphism and subduction. Surfaces of the Moon and Mars dating to this early time in the history of the solar system also reveal another reason for paucity of geological records: heavy bombardment of larger planets by planetesimals and other debris of the evolving solar system. One of these very large impacts at about 4.4 Ga is thought to have created our Moon from a Mars-size impactor ('Theia') melting itself and a large fraction of the Earth (Halliday, 2004). Impact melting of rocks would have destroyed both life and any precursor clay or organic matter (Maher and Stevenson, 1988). Another impediment to life and its precursors in surface environments would have been ultraviolet and other forms of protein-denaturing radiation from a young Sun unfiltered

by a fully formed atmosphere and ozone layer (Sagan and Pollack, 1974).

9.11.4.3 Soil

The idea that life came from soil is probably the most ancient human view of the origin of life. A cuneiform text dating to 2000 BC from the Sumerian city of Nippur in present-day Iraq (Kramer, 1944) describes a feast of the gods presided over by Ninmah (mother earth) and Enki (god of the waters), in which Ninmah created different kinds of humans from clay and Enki decreed their role in life. The idea of the spontaneous generation of life from soil was popular for centuries, endorsed by such influential thinkers as Aristotle (*c.* 384–322 BC) and Lucretius (*c.* 99–55 BC), but lost much appeal in the late nineteenth century when Louis Pasteur showed that life did not arise spontaneously from the soil or organic matter, but from microscopic propagules. Nevertheless, soil has much promise as a site for the origin of life because of its self-organizing complexity, lack of chemical equilibrium, and benign physicochemical conditions (Retallack, 2001a). The soil is alternately full and then drained of water after rain-storms, but thin films of water persist in pockets and margins of grains in which complex chemical reactions stop and start with the influx and evaporation of soil water. Weathering is a slow process taking millions of years to reach chemical equilibrium in the production of nutrient-starved kaolinitic clays. Soils are full of insulating clays, which moderate heat differences; diaphanous grains, which filter harsh electromagnetic radiation; and base-rich clays, which neutralize environmental acids.

9.11.4.4 Sea

The idea that “life like Aphrodite was born on the salt sea foam” (Bernal, 1967) has a long pedigree extending back to Ionian Greek philosophers Thales (*c.* 585 BC) and Anaximander (*c.* 565 BC). The appeal of this idea comes partly from the abundance of life in the sea at present and the aqueous geochemistry of most vital processes. There is also the saline composition of blood plasma and cytoplasm of land animals, suggestive of marine origins. Charles Darwin’s (1959) posthumously published vision of life’s origins in “some warm little pond, with all sorts of ammonia and phosphoric salts, light, heat, electricity, etc.” hint at the theory’s greatest weakness: the remarkably uniform and dilute composition of the ocean, which

tends toward chemical and physical equilibrium (Nisbet and Sleep, 2001). The smaller the pond and the more intermixed with mineral matter, the better opportunity for complex serial biosynthesis.

9.11.4.5 Deep-Sea Vent

Ocean floor volcanic vents discovered by deep-ocean submersibles are yet another plausible site for the origin of life (Corliss *et al.*, 1981). Hot water billows out of these vents like black smoke because of the rapid precipitation of sulfides and other dissolved substances into cold sea water. The vent fluids are like toxic waste, highly acidic, oxygen starved, and extremely hot (temperatures up to 380°C). There is little organic carbon or organisms immediately around them, but inches away the seafloor is crowded with life: large white clams and crabs, and peculiar pale tube worms. Only a few meters beyond, cold temperatures and lack of nutrients severely curtail biodiversity. At such great depths there is no possibility of photosynthesis, so the community is fed by chemosynthetic microbes (Table 2). Life is envisaged to have originated in spongy rock and vent deposits between the hot, caustic vents and cold, dilute ocean. Hydrothermal vent origin of life is an appealing explanation for widespread heat-shock proteins and primitive thermophilic bacteria (Nisbet and Sleep, 2001), though it is debatable how primitive are these adaptations (Forterre, 1995). Unlike soils or tidal flats, deep-sea vents are not controlled in their location or persistence by microbial scums. Black smokers erupt and become dormant like volcanoes, by virtue of subsurface faulting creating conduits for fluid migration in their oceanic rift valleys.

9.11.4.6 Other Worlds

Another possibility is that life evolved elsewhere in the universe and colonized our planet as propagules that could withstand long-distance transport in space. This concept of ‘panspermia’ goes back to the turn of the century and the Swedish chemist Svanté Arrhenius. A related idea is deliberate colonization of the Earth by advanced extraterrestrial civilizations (Crick, 1981). All manner of organisms could have been broadcast or dispatched, ranging from influenza viruses or unicellular bacteria, to sophisticated aliens in space vehicles. Such views have some appeal in this age of space exploration, but merely remove the question of the origin of life to another planet. The environment where that life evolved is likely to have

been Earth-like in many respects because life has long been well suited to our planet. Thus it remains useful to consider the origin of life from natural causes here on Earth.

9.11.5 Coevolutionary Histories

Biosignatures of Earth can be understood by their long geological histories, which present natural experiments with variation that may reveal oscillatory mechanisms of coevolution. The metaphor of an arms race is commonly applied to coevolutionary trajectories because they appear to change toward more competitive organisms (Dawkins and Krebs, 1979). Increasingly discriminating bees are more faithful to the pollination of particular flowering species. Grassy sod with basal-tillering and telescoped internodes is best able to absorb the rough grazing of tall-toothed and hard-hooved bison and horse. Similarly in warfare, arms races between competing nations result in more sophisticated and powerful weapons. Unlike biological coevolution, which has spawned diversity, arms races commonly result in imperialism, and loss of other cultures, their languages, and archives.

Another metaphor for coevolution is the Red Queen hypothesis (Van Valen, 1973) named for Lewis Carroll's character who had to run faster and faster in order to stay in the same place, presumably within a social pecking order. This implies that all creatures are caught up in a rat race, and does not account for the persistence of such slow creatures as millipedes and possums.

We prefer a more fundamental human metaphor of technological escalation, which applies also to activities such as information technology (Vermeij, 2004). Hand-copying, printing, and electronic publication are successive improvements in information technology driven by inventions (adaptations) of companies (producers) to satisfy demands of customers (consumers), who in turn adjust their consumption (coadaptations). Technological escalation in information technology has resulted not only in increasing quality and volume of communication between groups, but also increasing diversity and specialization.

By any of these models, the history of life and Earth environments should demonstrate a pattern of adaptations in one segment of the ecosystem followed by compensatory coadaptations in another part of the ecosystem. The following sections attempt

to identify such reciprocal changes in air, water, soils, and rocks. There are also catastrophic alterations of surface environments after large bolide impacts, giant volcanic eruptions, and metamorphism of limestone and coal (Retallack, 2002). There are biological perturbations, such as seasonal leaf-shedding and soil fertility cycles, on such short timescales (10^0 – 10^5 years) as well (Retallack, 2004a). Biological flexibility on these timescales may have aided biotic recovery from abiotic catastrophes. The present account however will deal only with evolutionary timescales (10^6 – 10^9 years).

9.11.5.1 Life and Air

Bioterraforming, or re-engineering of planetary environments by life, as may be possible for Mars (Fogg, 1998), could already have happened on Earth during the Precambrian evolution of cyanobacterial photosynthesis to create our oxygen-rich atmosphere. Evidence from paleosol geochemistry suggests an especially marked oxidation at about 2.3 Ga (Figure 8), as does the coeval demise of banded iron formations (Bekker *et al.*, 2004), fluvial uraninite placers (England *et al.*, 2002), and mass-independent fractionation of sulfur isotopes (Farquhar *et al.*, 2000). One difficulty for this scenario is the discovery of geologically older cyanobacterial microfossils (3.5 Ga; Schopf, 1983), stromatolites

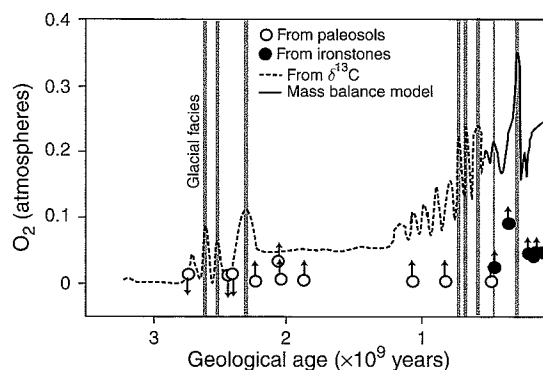


Figure 8 Reconstructed history of atmospheric oxygen abundance, showing constraints from selected Precambrian paleosols and Phanerozoic pedogenic goethites. The Phanerozoic curve (solid line) is from a sediment mass-balance computer model (Berner *et al.*, 2000) and the Neoproterozoic curve (dashed line) inferred from carbon-isotopic data (Anbar and Knoll, 2002). The gray bars represent times of known glacial sedimentation, which vary considerably in duration and recurrence interval. Updated from data of Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

(3.5 Ga; Allwood *et al.*, 2006), mats (3.4 Ga; Tice and Lowe, 2004), and 2α -methylhopane biomarkers (2.7 Ga; Brocks *et al.*, 2005). As was clear to Cloud (1976) in proposing photosynthetic oxidation of Earth, bioterraforming may take geologically significant spans of time.

Reasons why bioterraforming takes so long are clearer from more recent atmospheric oxidation events, such as the peak value of 35% O₂ during 390 Ma spread of forests, followed by a 250 Ma low of about 15%, before climbing again to the modern value of 21% with the 19 Ma spread of sod grasslands (Figure 8). Forest ecosystems of the Middle Devonian (390 Ma) culminated a steady increase in woody plant size from the first appearance of matchstick-sized primitive land plants during the early Silurian (430 Ma). These evolutionary changes were matched by comparably drawn out increases in depth and intensity of weathering indicated by paleosols, suggesting that not only individuals evolved but soil communities distributed over broad continental areas (Retallack, 1997a). Woody steles of the earliest land plants frustrated early Silurian herbivores, which were mainly millipedes adapted to feeding on unvascularized liverworts (Retallack, 2001c). A coevolutionary swing toward woody plants to the disadvantage of herbivorous animals would not be enough in itself to change the world unless such communities became globally distributed, commandeering the source of mineral nutrients in soil, metabolic gases in the atmosphere, and the distributive capacity of rivers and the ocean. The evolution of laminar leaves by 320 Ma would also have enhanced photosynthetic production of wood, and transpiration of water (Beerling *et al.*, 2001). Altering atmospheric composition requires not only living communities as a force for photosynthetic oxidation, but deep and protracted silicate weathering and burial of carbon in swamps and oceans in large quantities over many millions of years (Berner *et al.*, 2000).

Silicate weathering and carbon burial could be regarded as physicochemical components of global change, but both are under strong biotic control. The simple inorganic weathering agent, carbonic acid from dissolution of CO₂ in water comes largely from soil respiration, in turn fed by primary production, so that some rainforest soils have 110 times more soil CO₂ than in the atmosphere (Brook *et al.*, 1983). Other organic acids also enhance hydrolytic weathering (Berner, 1997) and are also more abundant in larger and more productive ecosystems. The physical evidence of this biotic enhancement of

weathering are the new kinds of forest soils (Alfisol, Ultisol), which appear as paleosols in Devonian and younger rocks (Retallack, 1997a). Carbon burial can be regarded as a purely physical process, but the biological invention then proliferation of lignin first occurred in a world without effective ligninase enzymes or creatures such as termites and dinosaurs which could reduce the load of woody debris undecayed and finding its way into rivers, lakes, swamps and the sea. The physical evidence of Devonian-Carboniferous increase in carbon burial is not only the increase in marine and lacustrine black shales, but the appearance of a new kind of peaty soil (Histosol) and wetland-forest (swamp) ecosystem in Devonian and younger rocks. Thus the soaring oxygen content of 35% in the Late Carboniferous (*c* 310 Ma) was largely a product of the balance between producers with the newly evolved product of wood getting ahead of consumers. By Late Carboniferous, there were few vertebrate herbivores, but many winged herbivorous insects, wood-eating cockroaches, and fungi. Carboniferous swamp woods were punky with soft cells (parenchyma) between the files of lignin-rich cells (tracheids), and so were less prone to wildfire than many modern woods. Carboniferous ecosystems thus included features that offset the oxidizing effects of trees, culminating in the evolution of tree-eating termites and dinosaurs during the long Mesozoic CO₂ greenhouse (Retallack, 2004a).

Sod grasslands formed by a coevolutionary process between grasses and grazers (Figure 1) on all continents except Australia and Antarctica over a geologically significant interval of time between the Late Eocene (35 Ma) and Pliocene (4 Ma). Studies of root traces in paleosols indicate that before grasslands the stature of woody vegetation became uniformly smaller from humid climate forests to arid shrublands, as it still does in Australia where mallee vegetation of 2–10 m multistemmed shrubs dominate a semiarid climatic belt of grasslands on other continents. From Late Eocene to Late Miocene, grasslands were confined to the rainfall belt of 300–400 mm mean annual precipitation, but after Pliocene evolution of large hard-hooved hypergrazers such as horses and wildebeest and megaherbivores such as elephants the prairie-forest ecotone was rolled back to nearer to the 1000 mm isohyet of mean annual precipitation (Retallack, 2001b). Evidence for this territorial advance of grasslands comes from the spread of the characteristic surface horizons of grasslands (mollic epipedon) in

soils with deep calcareous nodular horizons (calcic horizon) of subhumid climates. Human coevolution with grass-based agroecosystems has now created grasslands even in rainforest climate belts. This territorial expansion is evidence that this coevolved ecosystem is not merely an assemblage of creatures with mutually compatible climatic tolerances, but a biological force for global change.

Like forests of the Carboniferous, Cenozoic grasslands were a force for atmospheric oxidation because they increased plant biomass, particularly underground, compared with mallee vegetation in comparable climatic belts. Grasslands also promoted hydrolytic weathering, accelerating its pace with a rich soil fauna of earthworms and a variety of microbes. The result is a dramatic increase in soil organic carbon contents as high as 10 wt.% for depths of up to a meter, compared with mallee soils of comparable climatic belts which have such high organic content only in the surficial few centimeters of leaf litter. Soil organic matter also stabilizes a characteristic soil structure of crumbs, which withstands wetting and can be transported considerable distances by streams to be buried in agricultural stock ponds, lakes, river floodplains, and the ocean. Thus Cenozoic grasslands increased efficiency of biomass creation, hydrolytic weathering, and carbon burial in aridlands, just as forests did during their Devonian–Carboniferous evolution in humid areas and wetlands. Grasslands also effected global change by controlling two additional landscape attributes traditionally regarded as purely physical: albedo and evaporation (Retallack, 2001b). Grasses are lighter colored than trees, especially when hay-colored during dry or snowy seasons, and bow down under snow more readily than trees. Grasslands thus reflect more radiation back into space, with a net chilling effect on the planet. Grass colonization of the bare soil patches of desert shrubland or mallee has the effect of reducing evaporation of soil water. Grassy sod is moist and the air above grasslands dry, in contrast to woodlands and forests which have dry soil and moist air because of persistent transpiration of water by trees. Lowered vapor pressure of water over grasslands compared with forests also has the effect of planetary cooling, because water vapor is a powerful greenhouse gas. Thus the rise of O₂ to 21% with Cenozoic drying and cooling culminating in the Pleistocene ice age, may have been in part due to the coevolution of grasslands. The process is not without limits, but even the warming forces of greenhouse gas (CO₂ and CH₄) emissions from wildfires and large herds of ungulates are part of a system that promotes grassland

expansion at the expense of forests, because grasses recover from fire and herbivory more readily than trees. If anything can undo the cooling effect of grasslands, it would have to be their domineering evolutionary product, us (Palumbi, 2001).

9.11.5.2 Life and Water

Ocean composition also has changed on evolutionary timescales. Oceans of the early Earth may have been 1.5–2 times more saline than modern oceans, judging from fluid inclusion and evaporite volume data (Knauth, 1998). The principal mechanism for oceanic dilution is burial of large volumes of salt in evaporite deposits of barred basins, like the modern Persian Gulf. The growth of continental granitic crust promoted the tectonic formation of such basins and emergent continental storage of evaporites. Barred evaporite basins also are common behind coral, algal, and stromatolitic reefs, which extend back at least to the Paleoproterozoic (2 Ga; Grotzinger, 1988). Paleoproterozoic was also a time of peak abundance of banded iron formations (**Figure 9**), the distinctive laminated hematite and chert deposits of oceans during the transition from chemically reducing to chemically oxidizing oceans due to the spread of photosynthetic microbes (Cloud, 1976).

Roles for life in Precambrian changes in salinity and redox remain controversial because so remote in geological time. A clearer indication of life's role in oceanic chemistry on evolutionary timescales is the Phanerozoic oscillation between aragonite and calcite seas (**Figure 10**). This geochemical oscillation is seen in both carbonates and evaporites of the Cambrian to Mississippian (540–320 Ma) and Jurassic to Paleogene (170–20 Ma), which are dominantly calcite and sylvite-gypsum. In contrast, carbonates and evaporites of the Mississippian to Jurassic (320–170 Ma) and Neogene (20–0 Ma) are dominantly aragonite and kieserite-gypsum. The salts are simple precipitates, and although the carbonates are largely precipitated as skeletons of marine organisms, the major rock-forming organisms use passive forms of extracellular metabolically induced precipitation, rather than shell formation mediated by an organic-matrix (Stanley and Hardie, 1999). Calcite seas correspond broadly in time with high-CO₂ greenhouse intervals and aragonite seas with low CO₂ icehouse intervals, but this does not appear to be a simple pH effect from a more acidic atmosphere at times of high CO₂ and carbonic acid. Calcite is favored over aragonite only for a narrow range of

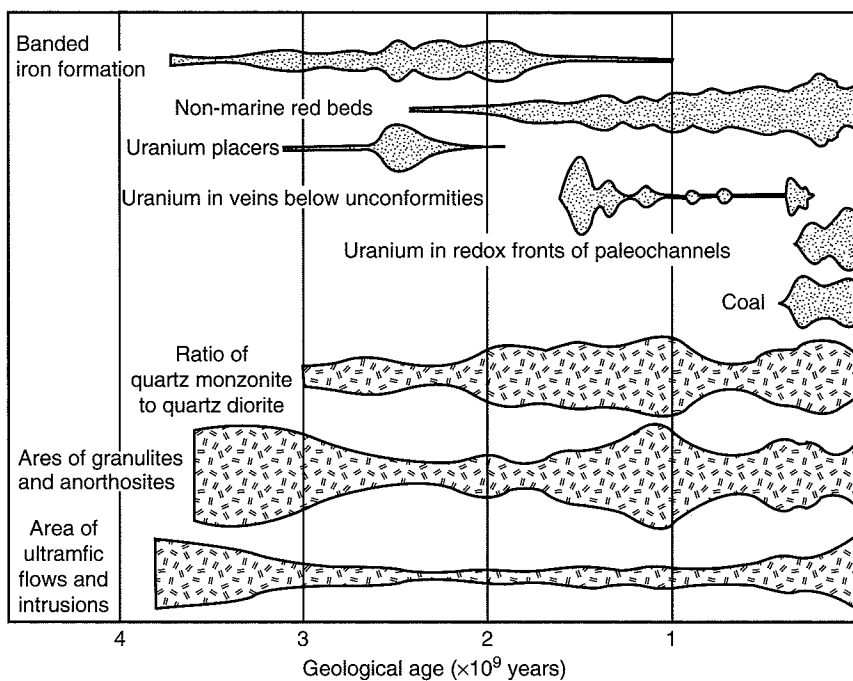


Figure 9 Relative abundance of sedimentary rocks, ores, and igneous rocks through geological time. Adapted from Engel *et al.* (1974), Meyer (1985), and Barley and Groves (1992).

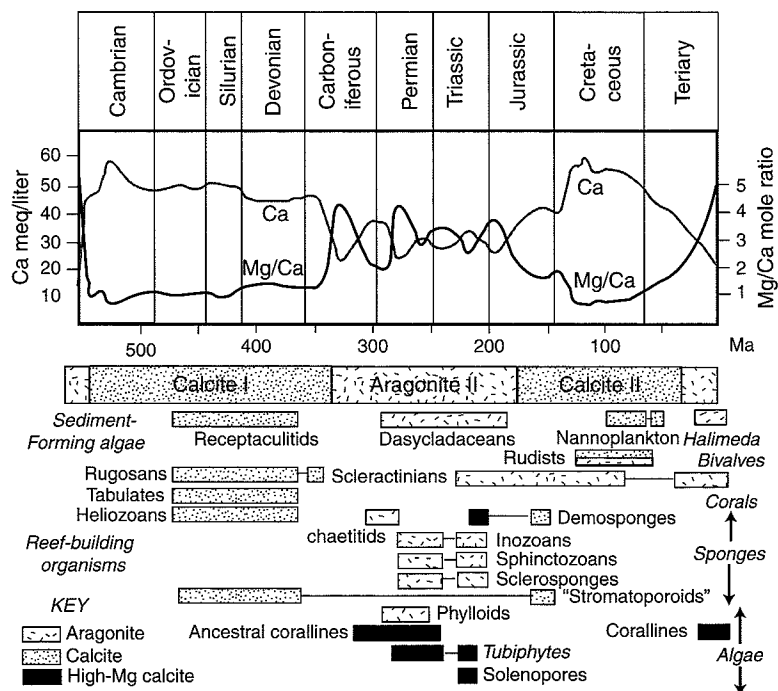


Figure 10 Correspondence between oceanic alkaline earth composition and the carbonate mineralogy of hypercalcifying marine organisms. Simplified from Stanley SM and Hardie LA (1999) Hypercalcification: Paleontology links plate tectonics and geochemistry to sedimentology. *GSA Today* 9(2): 1-7).

unrealistically high partial pressures of CO₂. Instead, calcite versus aragonite is favored by the Mg/Ca ratio of the ocean, with calcite favored when the ratio is less than 2 and aragonite favored when it is more than 2. The Mg/Ca ratio of the ocean is in turn controlled by the balance of cations in solution from rivers and from hydrothermal activity at mid-ocean ridges. Stanley and Hardie (1999) argue that variation in mid-ocean ridge spreading is the primary control, with active spreading inducing a greater hydrothermal flux and sub-seafloor sequestration of Mg. However, rates of seafloor spreading are not appreciably varied over the past 180 Ma in which spreading can be inferred accurately from seafloor magnetic striping (Rowley, 2002). An alternative control on oceanic Mg/Ca ratio is riverine input, considering the greater abundance of dolomitic caliche in paleosols of high-CO₂ greenhouse climates than those of low-CO₂ icehouse climates (Retallack, 2004a). Dolomitic versus calcitic caliche in turn is probably under microbial control of methanogens encouraged by soil CO₂ higher than usual for arid-land soils (Roberts *et al.*, 2004). From their source as they are weathered from soils to their sink in fossil coral reefs, Ca and Mg cations are under constant biotic surveillance.

9.11.5.3 Life and Soil

Compared with a long history of research on Precambrian air and water, even such basic questions as when soils became living remain uncertain. There was land and thick deeply weathered soil at least as far back as 3.5 Ga, as indicated by paleosols of that age (Buick *et al.*, 1995), and isotopic geochemical evidence for substantial amounts of Archaean continental crust (Bowring and Housh, 1995). Sterilization of land surfaces by intense ultraviolet radiation before development of the ozone layer is unlikely in an early atmosphere of methane, water vapor, or other dense early atmospheric gases (Rye and Holland, 1998). The very existence of clayey paleosols that were well drained, is regarded by some geomorphologists as evidence of stabilization by life, including microbes (Schumm, 1977). Remarkably light C-isotope values ($-40 \delta^{13}\text{C}\%$ vs PDB) indicative of methanotrophs have been recorded from the 2.8 Ga Mt Roe paleosol of Western Australia (Rye and Holland, 2000) and isotopically heavy C-isotope values (-16 to $-14 \delta^{13}\text{C}\%$) like those of hypersaline microbes from the 2.6 Ga Schagen paleosol South Africa (Watanabe

et al., 2000). Many paleosols have carbon isotopic values in the normal range for cyanobacteria and other photosynthetic organisms (-24 to $-30 \delta^{13}\text{C}\%$): 2.3 Ga pre-Huronian in Ontario (Farrow and Mossman, 1988), 1.3 Ga post-Mescal paleokarst in Arizona (Beeunas and Knauth, 1985), and 0.8 Ga pre-Torrisonian in Scotland (Retallack and Mindszenty, 1994). To this list of methanogenic, hypersaline and photosynthetic life in Precambrian paleosols must be added decomposers such as actinobacteria, because most Precambrian paleosols have only traces of organic C. Something must have cleaned up the dead bodies, just as fungi and other microbes create carbon-lean modern soils. Filamentous cyanobacterial microfossils have been found permineralized on a paleokarst paleosol in the 1.3 Ga Mescal Limestone of Arizona (Horodyski and Knauth, 1994) and within angular flakes of laminated crust in the 2.8 Ga Mt Roe paleosol of Western Australia (Rye and Holland, 2000). Comparable modern microbes have wide environmental tolerances and comparable fossils are common in permineralized stromatolites (Schopf, 1983), so were they soil microbiota or just opportunistic colonists of puddles? Lichen-like microfossils from the Carbon Leader of the Witwatersrand Group of South Africa (Hallbauer *et al.*, 1977), now known to be 2.9 Ga in age (England *et al.*, 2002), have been discounted as artefacts of acid maceration (Cloud, 1976), although it is clear from petrographic thin sections that they predate metamorphic veins (MacRae, 1999). The organic matter of these structures has isotopically light carbon, pristane, phytane, and pentose \approx hexose indicative of photosynthesis (Prashnowsky and Schidlowski, 1967). Life in soil may go back 3.5 Ga, and perhaps further (Figure 7).

Soils played a role in coevolutionary oscillations of atmospheric and oceanic chemical composition (Figures 8–10). One of the best lines of evidence for a chemically reducing early atmosphere on the early Earth are paleosols older than 2.2 Ga with the deep spheroidal weathering suggestive of a well-drained soil yet chemically reduced clay and iron minerals. These appear to be a completely extinct form of soil, which we have labeled 'Green Clays' (Retallack, 2001a). The degree and depth of weathering of paleosols on the early Earth is remarkable, with some of them attaining the alumina enrichment of bauxites (Figure 11). The appearance of laterites and oxidized paleosols is one of the most obvious lines of evidence for Paleoproterozoic atmospheric oxidation (Rye and Holland, 1998). The oldest

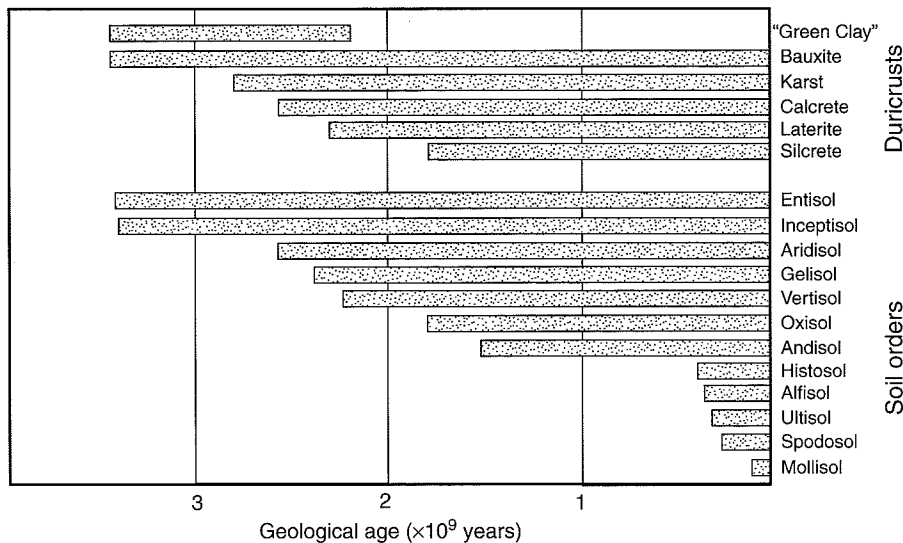


Figure 11 Stratigraphic range of duricrusts and soil orders. Entisols and Inceptisols are assumed to have been precursors of other paleosols, even though difficult to recognize in the rock record before the advent of roots and burrows in soils. Updated from data of Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

known caliche may be the Schagen paleosol (2.6 Ga) of South Africa (Watanabe *et al.*, 2000). The first silcrete appears along with the first desert dunes and sizeable (>600 km) continents in the pre-Pitzz Formation paleosols (1.8 Ma) of northern Canada (Ross and Chiarenzelli, 1984). Silcretes and caliche are characteristic of deserts on Earth today, and it is surprising not to find them on the earlier Earth. Primordial rolling terrains of Green Clay paleosols would have seemed unusual by modern standards. Without benefit of multicellular vegetation these landscapes would have appeared barren and desolate, and yet these deep clayey soils lacked the harsh rocky outlines and dunes of modern desert landscapes. Their deep weathering can be attributed to a humid, maritime climate and warm temperature assured by the greenhouse effect of elevated levels of atmospheric water vapor, methane, and carbon dioxide (Figure 12).

The role of soil in Phanerozoic environmental oscillations is especially clear from trace and body fossils associated with paleosols. Burrows and tracks of millipedes first appear in mid-Ordovician (460 Ma) paleosols (Retallack, 2001c), and termite nests in Triassic (230 Ma) paleosols (Hasiotis and Dubiel, 1995). These animals and their burrows had profound effects on soil aeration and respiration. The effect of plants is even more obvious, with reduction spotting and rhizome traces first appearing in early Silurian (430 Ma) paleosols, and large root traces of

trees in middle Devonian (390 Ma) paleosols (Retallack, 1997a). The evolution of trees coincided with evolution of forest soils (Alfisols), and was soon after followed by forest adaptations to low-nutrient conditions of mineral-poor peats (Histosols) by Late Devonian (365 Ma), silica sands by Mississippian (330 Ma), and base-poor clays (Ultisols) by Pennsylvanian (320 Ma; Retallack, 2001a). The evolution of sod grasslands with their characteristic surface horizons (mollic epipedon) ushered in the first grassland soils (Mollisols) in the Early Miocene (19 Ma). The oscillation of terrestrial communities dominated by millipedes, forests, termites, and grasslands corresponds to grand swings in greenhouse to icehouse paleoclimates (Retallack, 2004a).

9.11.5.4 Life and Rocks

The abundance of different kinds of rocks has also waxed and waned with life-mediated paleoenvironmental changes on Earth (Figure 11). Uranium ores for example, show three distinctive forms, which succeed one another over geological history. Uranium in fluvial sandstones is found as rounded grains that were evidently transported by rivers in well-aerated turbulent water and trapped in less-turbulent portions of the stream with gold and other heavy grains in a kind of deposit called a placer (England *et al.*, 2002). These dense grains of uraninite are highly unstable in modern rivers

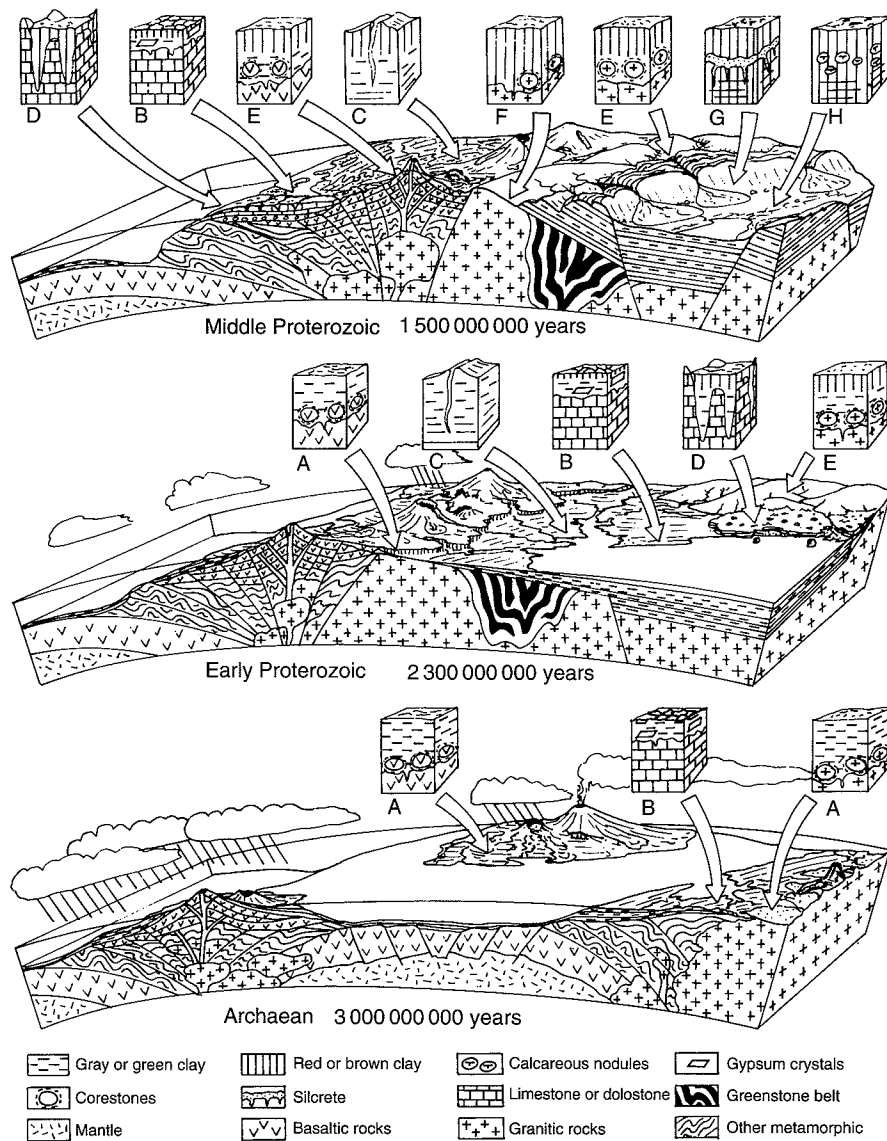


Figure 12 A speculative scenario for the evolution of soils, atmosphere, and continents on the early Earth, in which desert soils appear long after deeply weathered soils. The soil types illustrated are extinct 'Green Clays' (A), salty soils or Salids (B), swelling clay soils or vertisols (C), karst and drab cave earth or Orthents (D), oxidized incipient soils or Ochrepts (E), red and deeply weathered soils or Oxisols (F), desert soils with silcretes or Durids (G), and desert soils with calcareous horizons or Calcids (H). From Retallack GJ (2001a) *Soils of the Past*, 2nd edn., 600pp. Oxford: Blackwell.

because they readily oxidize to soluble yellow carno-
nite. Evidence from chemically reduced but deeply
weathered paleosols beneath these uraniferous sand-
stones support the notion that uraninite placer
deposits reflect a distant time (1.9–3.1 Ga) of much
lower atmospheric oxygenation (Farrow and
Mossman, 1988). In geologically younger rocks
(0.2–2.0 Ga), uraninite ores fill veins and breccias
associated with unconformities, where they precipi-
tated at a zone of chemical reduction after

transportation in overlying sandy aquifers from
granitic sources as oxides in surface and ground
waters. Geologically younger again (0.3–0 Ma) are
roll-type uranium ores, which reflect similar oxi-
dized transportation in groundwater until
encountering a redox front in fluvial sandstones
locally rich in organic matter such as fossil logs
(Meyer, 1985). Thus uranium ore types are related
to atmosphere and aquifer redox, which in turn are
biologically mediated (Retallack, 2004a).

Common kinds of igneous rocks also show variations in relative abundance that may reflect biologically mediated paleoenvironmental changes (Figure 11). The early Earth like the Moon, Venus, and Mars was dominated by basaltic and ultramafic igneous rocks. The proportion of quartz-rich igneous rocks such as quartz monzonite and granite to diorite and other more mafic rocks increases steadily after 3 Ga, reaching peaks at about 1.8, 0.8, 0.4, and 0.2 Ga. The latter two corresponding to Ordovician and Jurassic are the familiar CO₂ greenhouse (Figure 8), calcite-sea (Figure 9), and animal-dominated soil episodes (Retallack, 2004a). These were times of widespread deep weathering in a warm-wet world lacking ice caps.

Other rocks may also be under surficial environmental control. For example, eruptions of basaltic andesites have been related to surface conditions, because volcanoes of northern California preferentially erupt during isostatic rebound after unloading of melted ice caps and sea level rise (Jellinek *et al.*, 2004).

9.11.6 Conclusions

A coevolutionary perspective explains many aspects of our Earth's distinctive geological history. As in classical cases of coevolution, such as grasses and grazers, interactions between unrelated but mutually dependent organisms promote the persistence of new and earth-changing ecosystems, such as grasslands. The global-change capabilities of such ecosystems arise because organisms are evolving primarily in response to other organisms, and only indirectly in response to their physicochemical environment. As coevolved ecosystems proliferated to commandeer the nutrient supply of soil, the metabolic gases of the air, and the medium of water, they altered the atmosphere, hydrosphere, pedosphere, and deep lithosphere.

At the heart of any coevolutionary process is the natural selection of specific coadaptations to other components of the biota. These inventions are rare events that appear to flaunt the laws of thermodynamics and entropy, because they promote continuing metabolism and disorder. As rare events, such adaptations are followed at long geological intervals by coadaptations, with the directed trends of coevolution achieved through a series of discrete oscillations. As these ecosystems spread and proliferated, coevolutionary oscillations of life were

transferred to the geological history of our air, water, soil, and rocks.

From this perspective, human agroecosystems are not the first coevolved ecosystem to induce global warming. Just as cyanobacterial mats cooled a world of methanogenic slimes about 2 Ga, forests cooled a world of millipedes about 390 Ma, and grasslands cooled a world of large mammals about 35 Ma, we can hope that new coevolutionary initiatives will restore Earth to a livable temperature. If we fail in this mission, other organisms may succeed.

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