

Chapter 1

THE FOSSIL RECORD OF SOILS

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INTRODUCTION

The record of organisms and environments of the past in sedimentary rocks is largely biased towards depositional areas, such as lakes and seas. What of non-depositional areas, such as the landscapes on which we humans play out our lives? Are they lost forever? They have left a different kind of record—one of physical, chemical and biological alteration of sediments and other materials during the quiet times between floods, landslides and other geological events. The resulting fossil soils (paleosols) are found at the hiatuses, diastems, disconformities and unconformities of the non-marine rock record. Those found at major unconformities representing millions of years of non-deposition, are the most problematic. They may have begun forming under very different climatic, or other soil forming factors, than those prevailing when the landscape was buried by younger deposits. It is easier to reconstruct paleosols and factors in their formation when the time over which they formed is short enough that conditions are unlikely to have changed. Burial of land surfaces on this kind of time-scale is found in river valleys and coastal plains. It is the thick sequences of red beds and coal measures of the world which have the best fossil record of soils.

The fossil record of soils is long. The oldest paleosol recognized here is 3100 million years old. There is no reason to doubt that, with increasing awareness of paleosols and increasingly sophisticated methods of detecting them, they will be found in rocks as old as the present sedimentary rock record, some 3800 million years before present.

Over this long sweep of geological time, soils have changed, just as there have been changes in continents, sediments, plants and animals. Some fossil soils of the distant past appear so different from modern soils, that they may represent extinct kinds of soil. It is difficult to be certain, because of the alteration of these paleosols by burial and metamorphism. The interpretation of fossil soils is in some ways similar to reconstructing past vegetation, animals and sedimentary environments from fossils and rocks. Nikiforoff (1943) has compared a fossil soil to the skeleton of a fossil animal. There are uncertainties in reconstructing the past, and many aspects of fossil soils may never be fully understood. Nevertheless, much current research is contributing to a history of soils on Earth. This essay is my attempt to review some major events in that history. I acknowledge a debt of inspiration to my predecessors in this effort (Yaalon, 1971; Hunt, 1972; Ortlam, 1980; Catt and Weir, 1981), and to numerous scientists cited here, who have made my recent review (Retallack 1981a) obsolete.

EARTH'S EARLY ATMOSPHERE

There are so many peculiar features of rocks older than 600 million years that the Earth is suspected to have been a rather different place at that time. Many of these features are better understood if free oxygen were present in much lower amounts than today, its place taken by other gases such as carbon dioxide and methane (Holland, 1984). Oxygen is generally thought to have built up to present levels in the atmosphere because of the activity of photosynthetic microbes (Schopf *et al.*, 1983). Although this view is widely accepted, other scientists have argued for appreciable Precambrian atmospheric oxygen (Dimroth and Kimberley, 1976; Clemmey and Badham, 1982). As for many aspects of Precambrian palaeoenvironments, evidence for atmospheric composition is hard to find and seldom compelling. Paleosols provide important clues.

Although many Precambrian paleosols are different from younger ones, they are not as different as they could be, judging from the kinds of soils now forming on the Moon and planets of our solar system. Soil formation on the Moon is most unlike anything on Earth, now or in the known past. In the absence of air and water there, micrometeorites are the major agent of soil formation (Lindsay, 1976). Impacts of large asteroids and meteorites are also important, but are less significant in the long run because they are much less frequent. Impact breaks down rock fragments and causes local melting. This results in the surficial accumulation of pulverized rock, composite particles welded by glass (agglutinates) and pure particles of glass. On Venus, possible duricrusts

have been seen in images from Soviet probes (Florensky *et al.*, 1983). Considering likely atmospheric conditions on Venus, such duricrusts are probably most similar to what on Earth would be regarded as metamorphic rocks formed under very high temperature and low pressure (Nozette and Lewis, 1982). Paleosols of the Lunar and Venusian type have not yet been found on Earth, but it is unlikely anyone has seriously searched for them. They probably will not be found in rocks younger than 3800 million years. Sedimentary rocks of that age are an indication of free liquid water at the Earth's surface. Very ancient gypsum (3500 million years old) is evidence of surface temperatures below 58°C since that time (Walker *et al.*, 1983). Mars has the most similar of the known planetary surfaces to that of Earth. Images from Viking landers show basaltic boulders, loose sand, silt and clay, as well as partly buried, cracked surfaces (Strickland, 1979). These cracked surfaces may be evaporitic duricrusts, like those forming in desert soils and playas on Earth (Clark and van Hart, 1981). This could be interpreted as evidence of substantial reorganization of some constituents of the planetary surface, but the degree of weathering now demonstrated on Mars is less than in the deserts of Earth.

Precambrian paleosols on Earth clearly formed in a world of running water and atmosphere, and may provide evidence of variation in their fluid environment with time (Holland, 1984). Atmospheric oxygen is the main oxidizing agent in the present atmosphere and its effect on weathering is profound. Iron liberated by dissolution of minerals is oxidized to ferric oxides (Fe_2O_3 or haematite), hydroxides or oxyhydroxides, in which iron is in the trivalent state (Fe^{3+}). These minerals are yellow, brown and red in colour, and are responsible for the warm hues of most modern soils. Under reducing conditions, as in swamps, iron released from minerals remains in its reduced state (Fe^{2+}), within drab coloured silicates (such as chlorite), sulphides (pyrite) or carbonates (siderite). Since bivalent iron is much more soluble than trivalent, iron tends to be lost from soils formed under reducing conditions. The iron content and colour of paleosols are potentially useful guides to the changing oxygen content of past atmospheres. However, a number of other soil forming factors must also be considered in interpreting any particular paleosol.

Among the oldest known paleosols and the most thoroughly studied, are profiles developed on Archaean granite and greenstone at the unconformity below 2300 million year old fluvial sediments of the Huronian Supergroup, north of Lake Huron, Ontario, Canada (Gay and Grandstaff, 1980; Kimberley *et al.*, 1984). A profile developed on greenstone has been called the Denison paleosol (Figure 1). This consists of green sericitic rock, lighter in colour and finer grained than

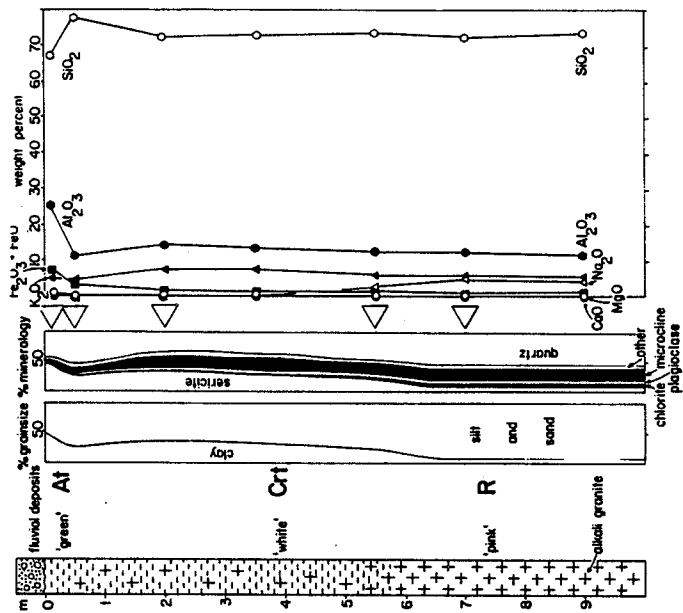
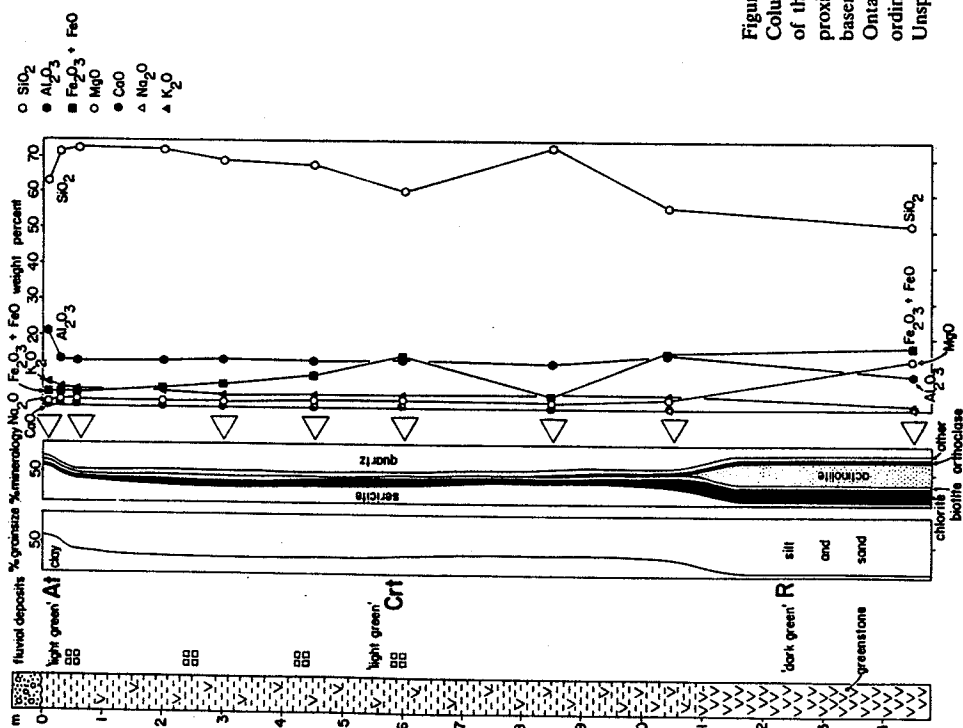


Figure 1. Columnar section, clay content and mineralogical and chemical composition of the Denison (left) and Pronto paleosols (right), both 'green clays', approximately 2300 million years old, on the unconformity between Archaean basement and Huronian fluvial sediments in the area north of Lake Huron, Ontario, Canada (adapted from Gay and Grandsstaff, 1980). Due to extraordinary thickness of these paleosols, the scale is different from other figures. Unspecified symbols are as for Figure 7.



the parent greenstone. Clay content (now sericite) was greatest near the surface (now to a depth of 0.6 m), moderately high to a considerable depth (now 10.5 m) and negligible within the parent greenstone. This very clayey surface layer is interpreted as an A horizon, in view of its contained organic carbon (Gay and Grandstaff, 1980). The less clayey subsurface layer is here identified as a C horizon (contrary to Gay and Grandstaff, 1980), comparable to the saprolite or Cr horizon of modern deeply weathered soils (as characterized by Birkeland, 1984). This layer is much thicker and shows more original rock fabric than is usual in the B horizons of modern soils. The mottled and pallid zones beneath soils on very ancient (millions of years old) landscapes (McFarlane, 1976) may be analogous to this extraordinarily thick, clayey layer.

Formation of clay (now sericite) from actinolite and feldspar, and to a lesser extent from chlorite, appears to have been the major soil forming process. There is also an increase in the amount of quartz at the expense of the same minerals. Since some of the quartz grains are larger than those in the parent greenstone, some quartz may have been newly formed in the paleosol rather than a residual concentrate. Alumina and silica vary in abundance with clay and quartz content, just as they would in a modern soil. The leaching of MgO from the profile is much as would be expected in modern soils of humid climates. The same is also true of CaO, although there was very little of this element in the parent material originally. There is even less Na₂O in the parent material, and it varies little within the profile. Surprisingly, K₂O appears to have accumulated. Total iron is strongly depleted from the profile. There is a little more trivalent iron than bivalent iron near the surface of the profile (above 0.6 m), but below that iron is largely present in the bivalent state.

Another profile along the same pre-Huronian unconformity, developed on pink alkali granite, has been called the Pronto paleosol (Figure 1). This has an especially clayey, light green surface (A horizon, now above 0.5 m) and a deep (now down to 5.5 m), clayey, white, subsurface (Crt) horizon. The main soil forming process appears to have been formation of clay (now sericite) from chlorite and feldspar. Chemical variation in Al₂O₃, SiO₂ and MgO is similar to that observed in the Denison paleosol. The amount of K₂O is rather constant, with slight subsurface (horizon Crt) accumulation and slight surficial (horizon A) depletion. There are very low levels of CaO throughout the Pronto paleosol, and Na₂O shows some surficial depletion, like that found in soils of humid climates. A notable feature of the Pronto paleosol is the surficial enrichment of total iron, largely in the trivalent state, in the upper part of the paleosol (above 6 m).

These two paleosols on the same ancient land surface are both drab coloured, as is typical of paleosols older than 2000 million years (Button and Tyler, 1979). However, they are surprisingly different in their degree of oxidation and amount of total iron. Just as modern soils are known to be products of many interacting factors, the interpretation of Precambrian paleosols must take into account complications such as metamorphism, clay diagenesis, original permeability, organic matter content, palaeodrainage and parent material.

Metamorphic environments are mostly reducing (Winkler, 1979), so it is possible that the lack of oxidized red beds older than about 2600 million years (Walker *et al.*, 1983) is because most rocks of such great age are metamorphosed. However, red oxidized paleosols have persisted in the Silurian Bloomsburg Formation of Pennsylvania (Retallack, 1985), despite metamorphism to lower greenschist facies (Epstein *et al.*, 1974). This is a much higher grade of metamorphism than suffered by the pre-Huronian paleosols (Roscoe, 1968). Furthermore, the existence of both weakly oxidized and unoxidized paleosols on this same ancient land surface is evidence against pervasive metamorphic alteration.

Changes in the composition of clays during burial may further obscure the original nature of paleosols. There is, for example, some reason to suspect that the anomalous K_2O enrichment seen in nearly all Precambrian (Williams, 1968; Button and Tyler, 1979; Gay and Grandstaff, 1980; Sokolov and Heiskanen, 1984) and many Palaeozoic paleosols (Retallack, 1985) may be diagenetic in origin. Comparable illitization of clays with depth has been observed in boreholes in many different parts of the world (Dunoyer de Segonzac, 1970; Van Moort, 1971; Heling, 1974; Hower *et al.*, 1976). A second alternative for the enrichment of K_2O in Precambrian paleosols is that this element was less depleted in soils lacking vascular land plants (Weaver, 1969). The efficacy of land plants in removing potassium is well known. In one set of experiments 15 successive crops of Ladino clover (*Trifolium repens*) removed 410 kg/hectare of K_2O from an illitic-montmorillonitic soil and 356 kg/hectare from a kaolinitic soil (Mehlich and Drake, 1955). In uncropped natural vegetation there is some recycling of potassium (Graustein and Velbel, 1981), but in many situations, especially in humid forested soils, it is ultimately washed out of the soil (Basu, 1981). Out of 128 analyses of North American soils cited by Marbut (1935), 55% show surficial depletion of K_2O . Only 33% show enrichment, and 74% of these are aridland soils. Thus a third explanation for potassium enrichment in paleosols is a dry, evaporitic climate in which alkali and alkaline earth elements accumulate in the soil (Duchafour, 1982). Although this last explanation stretches one's credulity the least, it seems unlikely considering the abundant clay, depth of

weathering, lack of evaporitic minerals or crystal casts, and depletion of Na_2O in pre-Huronian and in many other Precambrian paleosols. At present it is difficult to choose between a residual and metasomatic explanation for the enrichment of K_2O in Precambrian paleosols. If there has been widespread remobilization of potassium during burial, then analyses of other elements of Precambrian paleosols may also be suspect. Some encouragement can be gained from the composition of Precambrian and early Palaeozoic shales, which show enrichment in adjusted ratios of K_2O over $\text{FeO} + \text{Fe}_2\text{O}_3$ compared to igneous parent material, but no consistent trend of their own (Holland, 1984). This may mean that total iron content at least was not affected by potassium metasomatism, if indeed it occurred!

Drab coloured soils are found today in low lying parts of the landscape where atmospheric oxygen is excluded by waterlogging (Coultas, 1980). Such differences in palaeodrainage have been used to explain the different oxidation states of the Denison and Pronto paleosols (Gay and Grandstaff, 1980). This explanation has general appeal for the interpretation of Precambrian paleosols, because waterlogged soils are more abundant than oxidized ones in many sedimentary environments, and so would be more likely to be preserved in the rock record. In the case of the pre-Huronian paleosols, however, poor drainage seems unlikely. Their depth and degree of development is well in excess of modern waterlogged soils. In the Denison paleosol (thought to be waterlogged by Gay and Grandstaff, 1980) clay skins are evidence of a very low water table (below 6 m in the compacted profile). It is possible that the clay and soil structure of both Denison and Pronto paleosols formed under freely drained conditions, but were reduced under waterlogged conditions shortly before burial by overlying fluvial deposits. If this were the case though, it would have affected both paleosols more equally.

It is also possible that the degree of oxidation of Precambrian paleosols was related to their texture. This is an important limitation on depth of weathering of modern soils (Birkeland, 1984). In the case of the pre-Huronian paleosols the finer grained parent material (greenstone) is weathered to a greater depth than the coarse material (granite), the reverse of the usual relationship seen in modern soils. Although previous studies of pre-Huronian paleosols (Roscoe, 1968; Frarey & Roscoe, 1970) noted that greenstone profiles were thinner than granitic ones, the presence of exceptions such as the Denison paleosol, is an indication that grain size was not of overriding importance. Nor was the texture of the resulting soil, because the Denison and Pronto paleosols do not differ appreciably in the amount of clay despite their different parent materials.

Even in the present oxidizing atmosphere, complexes of clay and

stable organic matter in some well drained soils, such as Vertisols, can impart a drab colour (Duchafour, 1982). The effects of organic matter as an oxygen sink are difficult to assess for paleosols, because they are known to have consistently less organic matter than similar modern soils (Stevenson, 1969). Very low amounts of organic carbon have been found in the Denison paleosol (Gay and Grandstaff, 1980), but it is not dark coloured like modern organic soils. The organic matter content of the Pronto paleosol is not available for comparison.

A final consideration is the amount of iron present in the parent material. Under present oxygenic conditions, most iron released by weathering is oxidized. Thus, the reddest soils form on iron-rich parent materials such as basalt and greenstone. Curiously, the situation is reversed with the pre-Huronian and many other Precambrian paleosols: those developed on iron-rich rocks are less oxidized and have lost iron compared to those developed on iron-poor rocks. Holland (1984) has proposed a simplified model for weathering of silicate rocks, which explains this anomaly and allows calculation of some limits to the amount of oxygen in the atmosphere. Under conditions of low amounts of available oxygen within the soil there may not be enough to oxidize the large amounts of iron released from iron-rich rocks by the action of carbonic acid derived from CO_2 . Thus much iron remains reduced and is washed out of the profile. Even low amounts of oxygen may be sufficient to oxidize slowly released iron from iron-poor parent materials. Holland's calculations based on a number of Precambrian paleosols (including these pre-Huronian profiles) reveal dramatic changes in the ratio of the partial pressure of oxygen over the partial pressure of carbon dioxide from 3000 to 1500 million years ago ($\text{P}_{\text{O}_2}/\text{P}_{\text{CO}_2} \leq 1.3 \pm 0.5$) to the present ($\text{P}_{\text{O}_2}/\text{P}_{\text{CO}_2} \doteq 600$). Although very early Precambrian atmospheres now appear to be a little more oxidizing than many scientists would have expected a few years ago, they were much less oxidizing than the modern atmosphere. All things considered, the pre-Huronian paleosols are in accord with the conventional view of very low amounts of oxygen in the atmosphere during the very early Precambrian.

The oldest paleosol yet recognized is some 3100 million years old and is developed on granitic basement under the Pongola System, near Amsterdam, Transvaal, South Africa (Edelman *et al.*, 1983; Holland, 1984). There are several other drab paleosols of similar antiquity (Rankama, 1955; Eskola, 1963; Vogel, 1975; Herd *et al.*, 1976; Button and Tyler, 1979; Schau and Henderson, 1983; Edelman *et al.*, 1983; Holland, 1984). The youngest of these peculiar drab paleosols may be an altered basalt on the western flank of the Anabar Shield, north central Siberia, a little more than 1650 million years old (Chayka and

Zaviyaka, 1968; Chayka, 1970; Salop, 1973). Identification of these paleosols within modern soil classifications is difficult. They were certainly not Spodosols (of USDA; Soil Survey Staff, 1975), because they are much too thick, clayey and lack an eluvial horizon. They are also too thick and well developed for Gleysols (of FAO: Fitzpatrick, 1980), another suggestion of Gay and Grandstaff (1980). The only early Precambrian paleosol which can be identified with confidence is the 2200 million year old Waterval Onder paleosol from Transvaal, South Africa (Figures 4 and 5; Button, 1979). Its clear deep clastic dikes (former cracks) and hummocky surface (gilgai structure) mark it as the oldest known Vertisol (of Soil Survey Staff, 1975). Comparable structures have not yet been observed in the other early Precambrian paleosols, but this may be because their outcrop in mines and boreholes is limited. It also could be that these other drab paleosols represent an extinct kind of deep weathering. For the moment, the author proposes informally to call them 'green clays.'

On some very stable, ancient land surfaces it appears that clay formation and alumina enrichment, characteristic of green clay paleosols, continued to the extent that bauxites formed. The Namies Schist of the Bushmanland Sequence, in the Cape Province of South Africa, contains as much as 74 weight percent Al_2O_3 (Button and Tyler, 1979). The age and stratigraphic relationships of these metamorphic rocks are uncertain, but they are likely to be between 1000 and 2500 million years old (Kent, 1980). This makes them possibly the oldest of the described Precambrian bauxites (Rozen, 1967; Valeton, 1972; Button and Tyler, 1979). Bauxites are thought to form by long term weathering of silicate minerals at moderate pH in humid, non-seasonal climates, on stable land surfaces (Valeton, 1972). Bauxites also form as a residuum or deposit on limestone in regions of low sediment influx. Numerous examples of karst bauxites are known ranging back in age to Devonian (Nicholas and Bildgen, 1979). The oldest example of this kind of bauxite is developed on karst topography on the 2300 million year old Malmani Dolomite in Transvaal, South Africa (Button and Tyler, 1979). An older example of karst, at least 2700 million years old, is known from the Steeprock Group west of Thunder Bay, Ontario, Canada (Schau and Henderson, 1983). This unconformity is overlain by concentrations of manganese and chert, like those found also in parts of the karst on the Malmani Dolomite.

By about 1000 million years ago the atmosphere was discernably more oxidizing. The Sheigra paleosol of about this age in northwest Scotland (Williams, 1968) was developed on a topographic surface of Lewisian biotite gneiss, amphibolite and microcline pegmatite, with a relief of at least 600 m, and covered by Torridonian alluvial fan

deposits (Figures 2 and 3). In profiles developed on biotite gneiss, the surface (A) horizon (30 cm in the compacted paleosol) is stained red. Underlying this is a zone of mottled and light coloured rock including large round corestones of little weathered parent material (C horizon). Alteration extends more deeply into the parent material (up to 6 m from the surface) along joint planes. The profile is more clayey toward the surface but has a crystalline texture throughout. Quartz and microcline persist in the weathered part of the profile, but biotite and plagioclase of the parent gneiss have been extensively weathered to clay (now sericite). Clay formation is reflected in the decreased SiO_2 and increased Al_2O_3 toward the surface. Oxides of labile cations, such as CaO and Na_2O are depleted toward the surface, but K_2O appears to have accumulated, as is usual in Precambrian paleosols. Total iron is enriched in the surface horizon, and slightly depleted in the mottled and light coloured zone compared to parent material. Ferric iron (Fe_2O_3) is well in excess of ferrous iron (FeO) throughout the paleosol. This relationship is reversed in the parent material, as is usual for igneous and metamorphic rocks.



Figure 2. The Sheigra paleosol (bleached and reddened zone about 1 m thick), approximately 1000 million years old, developed on Lewisian amphibolite (left hand side) and biotite gneiss (to right) at the unconformity below Torridonian alluvial fan deposits, near the hamlet of Sheigra, north-west Scotland, UK. The white tape extending down from the unconformity, left of centre, is 2 m long.

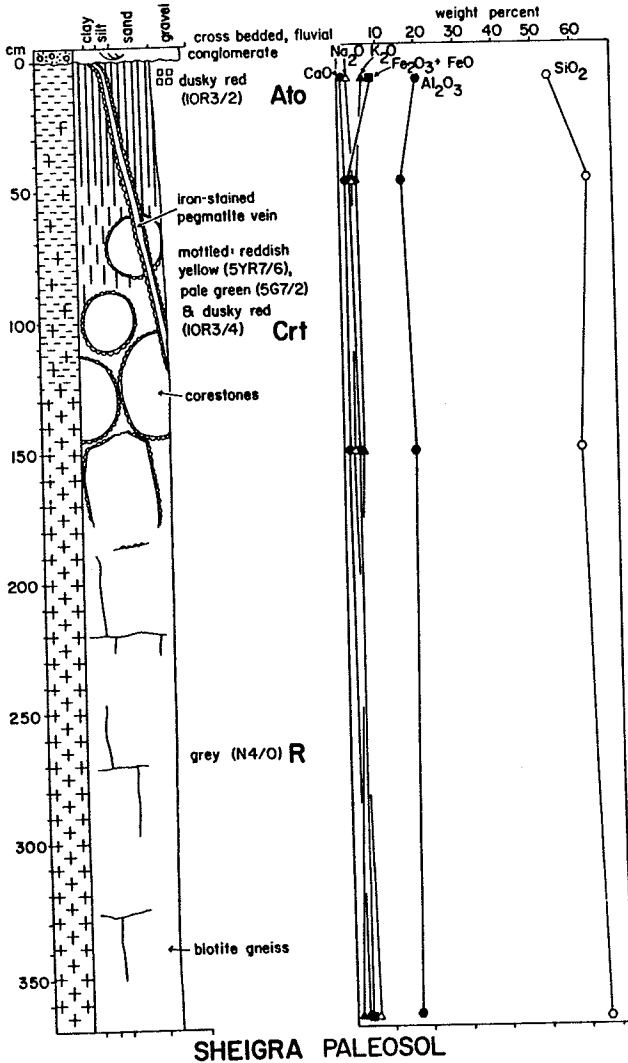


Figure 3. Columnar section and chemical composition of the Sheigra paleosol, probably an Inceptisol (adapted from Williams 1968, and personal field observations). Unspecified symbols are as for Figure 6.

Clay formation and ferruginization appear to have been the principal processes involved in the formation of this paleosol and it developed on a stable, ancient land surface. The bending of weather-resistant pegmatite veins at the surface of the paleosol has been interpreted (Williams, 1968) as evidence of Precambrian soil creep. This, and the depth of weathering of this paleosol are indications that it was well

drained. Incomplete leaching of Na_2O and CaO , and the presence of dolomite veins in paleosols developed on amphibolite on the same ancient land surface, are evidence of a subhumid palaeoclimate. The mild metamorphic alteration, good former drainage and coarse texture of this paleosol would all have aided oxidation. The parent rock had a high demand for oxygen, closer to that of the Denison paleosol than that of the Pronto paleosol (as calculated from its amount of iron and other oxygen consuming elements by Holland, 1984). The clear oxidation and ferruginization of this paleosol is thus evidence of a significant advance in the amount of oxygen in the atmosphere since the early Precambrian.

The oldest paleosol yet reported to be ferruginized is developed on granite underlying the Lower Sossan Group in the Simpson Islands, Great Slave Lake, Canada, thought to be older than 2200 million years (Stanworth and Badham, 1984). Ferruginized paleosols are more common in younger Precambrian rocks (Sharp, 1940; Blades and Bickford, 1976; Elston and Scott, 1976; Grabert, 1976; Kroonenberg, 1978; Knight and Morgan, 1981; Stanworth and Badham, 1984). Even though many of these are on major unconformities, their degree of weathering (where data are available) correspond with Inceptisols (of Soil Survey, Staff 1975). Contrary to Williams (1968), they are unlikely to be 'Podzols' (USDA Alfisols, Ultisols and Spodosols), because they lack recognizable eluvial and illuvial horizons.

When examined more closely, some of these ferruginized paleosols may turn out to have been more deeply weathered, and perhaps referable to Oxisols (of Soil Survey Staff, 1975). This is especially suggested by the 1660 million year old, deep weathering recognized beneath the Thelon Sandstone in the Thelon Basin, North West Territories, Canada (Chiarenzelli, 1983; Chiarenzelli *et al.*, 1983a,b; Miller, 1983) and 1500 million year old weathering beneath the Athabaska Group in the Athabaska Basin in nearby northern Saskatchewan (Ramaekers, 1981; Tremblay, 1983; Holland, 1984). Kaolinite, an indicator of extreme weathering, has persisted in these profiles despite subsequent diagenetic alteration. These paleosols show a haematite zone (up to 24 m thick) at the surface, overlying a transitional horizon (up to 24 m), and then a chlorite-rich horizon (up to 30 m) over fresh bedrock (Chiarenzelli, 1983). These zones are comparable to incipient lateritic zones, mottled zones and pallid zones found beneath deep tropical soils (McFarlane, 1976). No actual laterite has yet been found, although it could have been eroded or otherwise altered. The profiles show strongly eroded surfaces, silcrete caps, and subsurface dolomitization, as well as metamorphic alteration. Until laterite or lateritic detritus is actually found, the oldest likely laterite

remains the Cambrian or Ordovician Chester Emery of Massachusetts, USA (Norton, 1969). It could be that oxygen had not yet accumulated in the atmosphere to the extent needed to form laterite 1660 million years ago. Nevertheless, these deeply weathered Precambrian paleosols may represent the earliest known Oxisols.

Silcretes make their appearance in the rock record by about 1800 million years ago at the base of the Pitz Formation in the Thelon Basin (Chiarenzelli *et al.*, 1983a,b) in the North West Territories of Canada. Other Precambrian silcretes have also been reported (Donaldson and Ricketts, 1978). Silcretes are silica-rich rocks formed on stable cratonic land surfaces where the silica cement may be derived either from deep tropical weathering of silicates at moderate pH or from shallow, highly alkaline weathering, in desert soils (Summerfield, 1983).

Calcrete also appears in the rock record by about 1900 million years before present, in the form of pisolites in the Mara Formation, Goulburn Group, Bathurst Inlet, North West Territories, Canada (Campbell and Cecile, 1981; for age compare Easton, 1981). Numerous other Precambrian calcretes are known (Williams, 1968; Koryakin, 1971, 1977; Kalliokosi, 1975, 1977; Sochava *et al.*, 1975; Lewan, 1977; Donaldson and Ricketts, 1978; Bertrand-Sarfati and Moussine-Pouchkine, 1983; Chown and Caty, 1983). Many of these early calcretes are dolomitic in composition, like carbonate in modern soils of high base status (Doner and Lynn, 1977). Like the calcretes of modern Aridisols (Soil Survey Staff, 1975), they are thought to have formed in dry climates, in which soil solutions were not sufficiently copious or acidic to remove carbonate. The appearance of calcrete in the rock record may be related to the development of dry climates with increased size of continental nuclei and mountains, or perhaps to decreased amounts of atmospheric CO₂ and acidity of rainwater. Calcretes may be another indication of changes in terrestrial weathering roughly 2000 million years ago.

EARLY LIFE ON LAND

The microfossil record of life in marine cherts has now been found to extend back at least 3500 million years (Schopf *et al.*, 1983). Did a similar green slime also form the scum of the Earth? Theoretically, not only Precambrian soil microbiota, but even the origin of life in soil is likely (Nussinov and Vehkov, 1978; Bohn *et al.*, 1979). Unlike oceans and ponds, soils are intermittently wet and often contain concentrated solutions of salts, phosphates, metals and organic matter. The four most common cations in soils (Ca²⁺, Mg²⁺, Na⁺, K⁺) are closer in relative abundance to animals (dominated by Ca²⁺) than the sea

(dominated by Na^+). Soils are also less homogeneous than bodies of standing water. A variety of rare molecules may have been protected from hydrolysis in the nooks and crannies of mineral grains partly pulverized by micrometeorite bombardment, or breached by chemical or physical weathering. Soils are also the principal sites of clay formation on Earth, and clays may have been templates or catalysts for the formation of complex organic molecules. Hydrolysis-resistant short RNA molecules rich in the organic bases guanine and cytosine are viewed as possible molecular beginnings of life (Eigen *et al.*, 1981). In solutions rich in clays, metals, organic bases and a single enzyme, such RNA has been experimentally shown to 'reproduce.' Complementary 'reproductive' reactions could maintain RNA 'quasispecies,' and ultimately, energy transfer pathways and organisms. Life in these formative stages within the soil would be less threatened by competition from other 'quasispecies,' because of reduced mobility compared to open water. Furthermore, microbes and molecules in the soil would have greater protection from ultraviolet and other cosmic radiation than in open water. Although there are some theoretical reasons to prefer a 'primordial sludge' over a 'primordial soup' as the source of life on Earth, these competing hypotheses have barely been tested by hard evidence.

One suggestive line of evidence for early life on land is the modern affinities of some Precambrian marine microfossils, which are similar to modern soil microbes (Campbell, 1979). *Eoastrion*, a stellate microfossil with an opaque central body, is abundant in parts of the 2000 million year old, marine, Gunflint Chert of south-west Ontario, Canada. It is identical to modern manganese-fixing bacterium or microbial trace fossil *Metallogenium* (Barghoorn, 1977; Staley *et al.*, 1982), which is found in soils and rock varnish of modern deserts (Dorn and Oberlander, 1981). *Kakabekia* is another microfossil from the shallow marine, Gunflint Chert, which is similar to modern soil microbes (Siegel, 1977).

Some Precambrian microfossils have been found in situations which may have been terrestrial. Unfortunately, the palaeoenvironments of all of these are presently ambiguous. Microfossil nostocalean cyanobacteria have been found in cherts filling cracks into basement rocks underlying the Pokegama Quartzite in Minnesota, USA (Cloud, 1976; Hofmann and Schopf, 1983). Although this could have been a crevice fauna within a soil, it is likely that these microbes lived in the intertidal or subtidal zone of the sea in which the overlying Pokegama Quartzite was deposited. From a similarly ambiguous near-shore setting are supposed fossil fungi from the 2300 to 2700 million year Witwatersrand Group, South Africa (Hallbauer and van Warmelo, 1974). Although

the form of these 'fossils' are probably artefacts of preparation procedures, and so not fungi, their carbon is probably biogenic (Cloud, 1976; Barghoorn, 1981). It is hoped that future micromorphological studies of Precambrian paleosols will be more revealing.

Evidence for Precambrian soil microbes may also be gained from chemical analysis for organic carbon. The 2300 million year old Denison paleosol (Figure 1) is overlain by fluvial deposits and contains 0.014 to 0.25 weight percent of reduced organic carbon (Gay and Grandstaff, 1980). Although this is a very small amount, it is known from studies of Quaternary paleosols and their equivalent modern soils, that a great deal of organic matter is lost from soils soon after burial (Stevenson, 1969). Very few other Precambrian paleosols appear to have been analysed for organic carbon. This is likely to be a productive and quick source of information on the antiquity of life on land.

Another line of evidence for life in soil is the development of certain kinds of soil structure. Soils appear quite different from blocky cracked clay, of the kind seen in modern brick pits and desert badlands. They have a characteristic appearance and structure, which can be called *tilth* (Russell, 1973). This appearance is especially well demonstrated by the 2200 million year old Waterval Onder paleosol, from Transvaal, South Africa (Figures 4 and 5). Compared to the yellow and green clayey portion (C horizon) of the paleosol, the dark surface (A) horizon in the swales of the Vertisolic gilgai structure, appears well structured. The transitional horizon (AC) between yellow and dark parts of the profile is cracked into subangular, blocky clods (former soil peds) separated by contorted veins of dark material washed in from the surface (deformed illuviation ferrimangano-argillans in the terminology of Brewer (1976)). The surface (A) horizon has some relict bedding, and probably continued to accumulate sediment during development of the paleosol (as a cumulative horizon in the sense of Birkeland (1984)). This horizon also contains rounded clay granules (crumb peds), which are encrusted by a thin rind of iron and manganese (diffusion ferrimangan) better developed on the top than the bottom of the ped (Retallack *et al.*, 1984, Figure 5). This is very like modern rock varnish, thought to be fixed from atmospheric dust by fungi and bacteria (Dorn and Oberlander, 1981; Staley *et al.*, 1982). As for other lines of evidence for life on land discussed, this is unlikely to be the oldest because so few Precambrian paleosols have been examined carefully from this point of view.

A final circumstantial line of evidence for the advent of life in soils is related to the increased resistance to erosion associated with soil microbes (Campbell, 1979). Armouring of desert landscapes with rock

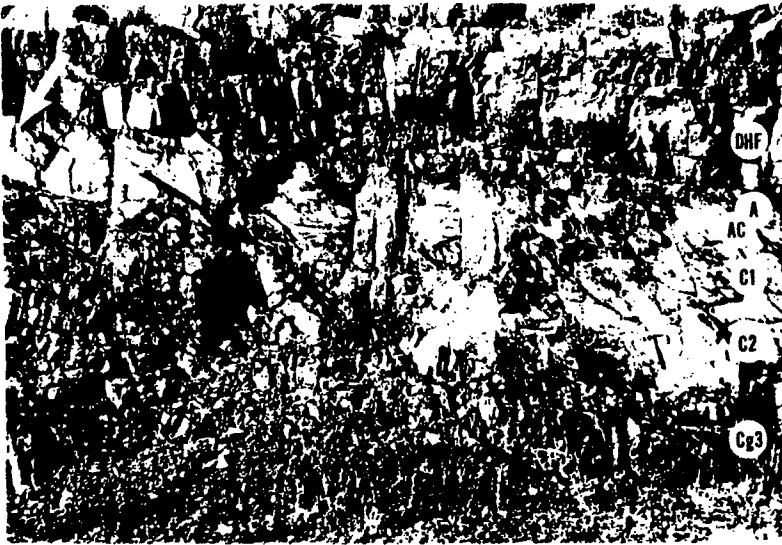


Figure 4. The Waterval Onder clay paleosol, approximately 2200 million years old, developed on clayey sediments overlying the Hekpoort Basalt and overlain by the fluvial Dwaal Heuvel Formation, in a road cutting east of Waterval Onder, Transvaal, South Africa. Thickening of the dark surface (A horizon as in Figure 5) to the right hand side, and clastic dyke (at arrow) below the thinner dark layer (A horizon) to the left, is gilgai structure. Hammer on light coloured rock (of C2 horizon to right) gives scale.

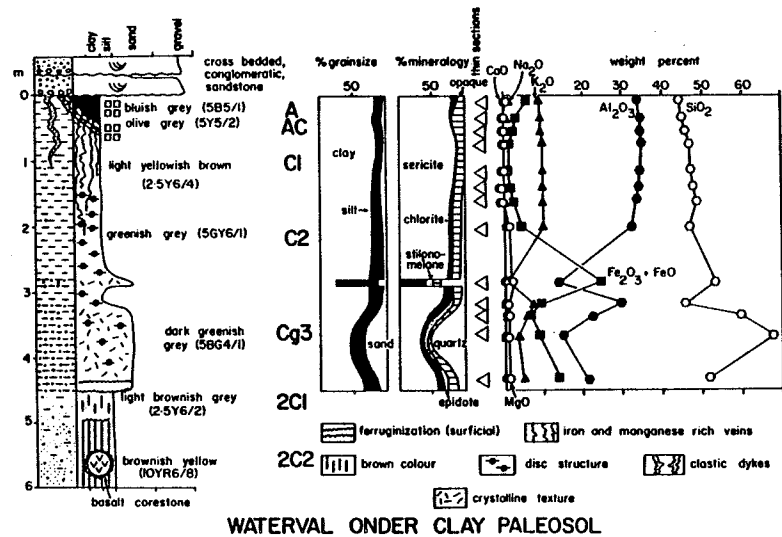


Figure 5. Columnar section, grainsize and mineralogical and chemical composition of the Waterval Onder clay paleosol, a Vertisol (adapted from Button 1979 and personal field observations and grainsize data). Scale is as for Figure 1 and unspecified symbols after Figure 6.

varnish (Dorn and Oberlander, 1981) and carbonate crusts (Krumbein and Geile, 1979) are well known. Microbial mats alone may appreciably stabilize desert soils (Booth, 1941). By contrast, in an hypothetical prebiotic landscape grains loosened from bedrock by physical or chemical weathering would tend to be washed away rather than weather in place to form clay (Schumm, 1968, 1977). Precambrian paleosols as old as 3100 million years are more clayey and deeply weathered than this hypothetical initial state. They represent land surfaces which were probably stable for many thousands, if not millions, of years.

In some cases (Gay and Grandstaff, 1980; Edelman *et al.*, 1983), very ancient soils are more oxidized than would have been expected from other lines of evidence for atmospheric composition, such as the nature of marine rocks and geochemical models for formation of the Earth (Holland, 1984). Could it be that oxygen producing organisms appeared earlier in the soil than in the sea? Was the present situation of an oxidized atmosphere and more reducing CO_2 -rich soil reversed during the earliest Precambrian? Could the redistribution of iron in Precambrian soils have been controlled more by microbial chelation than levels of atmospheric oxidants? These and other consequences of the evolution of soil biota remain to be explored.

LARGE PLANTS AND ANIMALS ON LAND

Although there is reason to believe that the greening of the land may have occurred well back in geological time, the appearance of multicellular land plants was a major advance in the complexity and biomass of terrestrial ecosystems, with consequences for soil formation. A few studies of paleosols relevant to this event have been published. They provide some constraints on when, which and how large plants and animals came to live on land.

Considering the very different lines of evidence used, it is not surprising that there has been some controversy concerning the antiquity of large plants and animals on land. Plants with a distinct vascular strand almost certainly lived on land and the oldest megafossils of these are mid-Silurian (Wenlockian) in age (Boucot and Gray, 1982). The extinct nematophytes also had tubular conducting tissues (Niklas and Smocovitis, 1983), although this differs from that of vascular plants. These fossils are poorly understood and peculiar in other ways, so that there is some doubt whether they were aquatic, semi-aquatic or land plants (Lang, 1937; Niklas, 1982). Nematophytes may be as old as Early Silurian (Llandoveryan: Pratt *et al.*, 1978; Strother and Traverse, 1979). All of these remains are restricted to

sedimentary environments where aerobic decay of organic matter was limited by waterlogging. Fossil spores of plants are preserved in similar environments, but are more widespread, extending also into shallow marine environments. Although there has been some dispute over which of the many organic-walled palynomorphs extracted from early Palaeozoic rocks can be regarded as belonging to land plants (Banks, 1975; Gray and Boucot, 1977), this line of evidence for land plants extends back to the Late Ordovician (Caradocian: Gray *et al.*, 1982). Late Silurian (Ludlovian and Pridolian) fossil millipedes are the oldest well accepted evidence of large land animals (Rolfe, 1980; Shear *et al.*, 1984). Early land animals have an even more patchy fossil record than plants and spores, because they were rare, delicate, and destroyed in acidic as well as oxidizing environments.

Compared to this poor fossil record, paleosols are preserved in a wider range of environmental conditions and will probably provide more copious evidence of early multicellular organisms on land. To date, however, only two paleosols pertinent to this question have been reported. One of these is in the Late Ordovician (Ashgillian) Juniata Formation near Potters Mills, Pennsylvania, USA (Retallack, 1985). This is a red, clayey profile interbedded with other, similar paleosols and fluvial sandstones (Figure 6). The paleosol is riddled with burrows of large (3 to 16 mm diameter) invertebrates, which are subhorizontal in the surface of the paleosol (up to 4 cm from the top) and subvertical below that (to depths of 50 cm in the compacted paleosol). The burrows are thought to be original parts of the soil because calcareous nodules are preferentially distributed around them and because the increase in clay at the expense of mica toward the surface corresponds with increased density of burrows. The soil is not thought to have been marine influenced because palaeogeographic reconstructions place the sea 200 km to the west at the time it formed, and because there are no marine fossils (including palynomorphs) in these rocks. Nor are there any associated fossils or sedimentary structures which could be construed as evidence of lakes (Retallack, 1985). It thus seems likely that relatively large animals had colonized dry soils by Late Ordovician (Ashgillian) time. What kind of animals these could have been is uncertain, but it is difficult to imagine how they could have survived without plants for food and shelter.

Additional evidence for Late Ordovician (Caradocian or Ashgillian) land plants is provided by a paleosol from the Dunn Point Formation of the Arisaig area, Nova Scotia, Canada (Dewey in Boucot *et al.*, 1974). This consists of 1.3 m of red, calcareous claystone on top of weathered corestones and columnar jointed flows of andesite. Near the surface of the paleosol are irregular pockets, about 1 m wide and

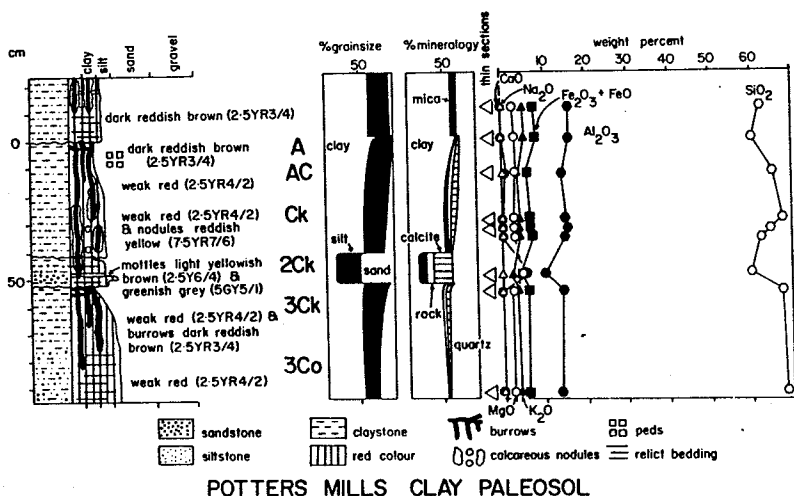


Figure 6. Columnar section, grainsize distribution and mineralogical and chemical composition of the Potters Mills clay paleosol, an Oxic Ustropept (Inceptisol), in the Late Ordovician (Ashgillian), Juniata Formation, near Potters Mills, Pennsylvania, USA (from Retallack 1985).

20 cm deep, filled with red shale redeposited from erosion of the paleosol. Small white reduction spots within the mounds between erosional pockets, are similar to drab mottles forming during diagenetic alteration of organic matter buried in red paleosols (Retallack, 1983b). Perhaps the mounds were stabilized against erosion by clumps of non-vascular land plants which lacked rooting structures substantial enough to leave obvious traces.

These two indications of relatively large, Late Ordovician, animals and plants on land may turn out to be among the oldest, because there are no large terrestrial fossils likely to be much older. Many more paleosols need to be examined, however, before the matter is closed.

Which large organisms were among the first to live on land is a problem for which the evidence of paleosols provides only the broadest of constraints. No obvious traces of rhizomes or roots were seen in either of the Late Ordovician paleosols. Considering the low preservation potential of organic matter in oxidized paleosols (Retallack, 1984a), there may have been a substantial biomass of multicellular filamentous or thallose plants and microbes growing in them. This may have included algae, lichens, bryophytes or completely extinct plants (Gray and Boucot, 1977). This earliest multicellular flora is unlikely to have included the enigmatic *Eohostimella* or vascular land plants, such as rhyniophytes. *Eohostimella* is preserved erect in the Early Silurian

(Llandoveryan) Frenchville Formation of Maine, USA (Schopf *et al.*, 1966). This could be regarded as a poorly developed coastal paleosol (such as an Aquent of Soil Survey Staff, 1975). It could also have been sediment at the bottom of a lake or tidal pool, although this seems less likely in view of the chemical affinities of *Eohostimella* with vascular land plants (Niklas, 1982). Wispy, tubular bioturbation has been found in a well drained, red, calcareous Late Silurian (Ludlovian) paleosol, near Palmerton, Pennsylvania, USA (Figure 7). This is the oldest likely trace of early vascular land plants, such as rhyniophytes, on dry land yet recognized. This paleosol is much better developed, and has better differentiated horizons, than the otherwise comparable Late Ordovician paleosol from Pennsylvania (Figure 6).

Although evidence is sparse at present, it appears that primitive vascular plants were preceded on land by a variety of poorly known non-vascular plants. The nature of the earliest large soil animals remains uncertain. The only generally accepted terrestrial fossils from rocks older than Devonian are millipedes, known definitely from rocks as old as Late Silurian (Ludlovian and Pridolian), but represented by very doubtful specimens from older Silurian (Llandoveryan) rocks (Rolfe, 1980; Shear *et al.*, 1984). Silurian scorpions have been widely considered as among the oldest land animals, but from careful assessment of their fossilized respiratory organs it appears that these were aquatic creatures until at least Late Carboniferous time (Kjellesvig-Waering, 1966). The extinct, largely aquatic eurypterids are known from rocks as old as Ordovician (Størmer, 1955). Respiratory organs of some especially well preserved Silurian eurypterids are indications that some of them may have been amphibious (Størmer, 1977). A case for the partly terrestrial habitat of some Late Carboniferous king crabs (xiphosurs) has been made but most of these creatures, extending back to Early Cambrian times, appear to have been marine (Størmer, 1955). Velvet worms (onychophorans) and earthworms (oligochaetes) are two common living terrestrial organisms with a virtually non-existent fossil record in non-marine rocks. Curiously they have been found in marine rocks as old as Middle Cambrian and Middle Ordovician, respectively (Thompson & Jones, 1980; Morris *et al.*, 1982). Although the fossil record of water bears (tardigrades) does not extend back beyond the Cretaceous, they are an evolutionarily isolated group, are small, easily dispersed and remarkably resistant to desiccation (Rolfe, 1980). Among these possible early inhabitants of land, only a few are unlikely to have made the burrows observed in the Late Ordovician paleosol from Pennsylvania. Both earthworms and velvet worms are soft bodied, and probably would have dehydrated in these permanent dwelling burrows. Tardigrades are much too small. Accepting the

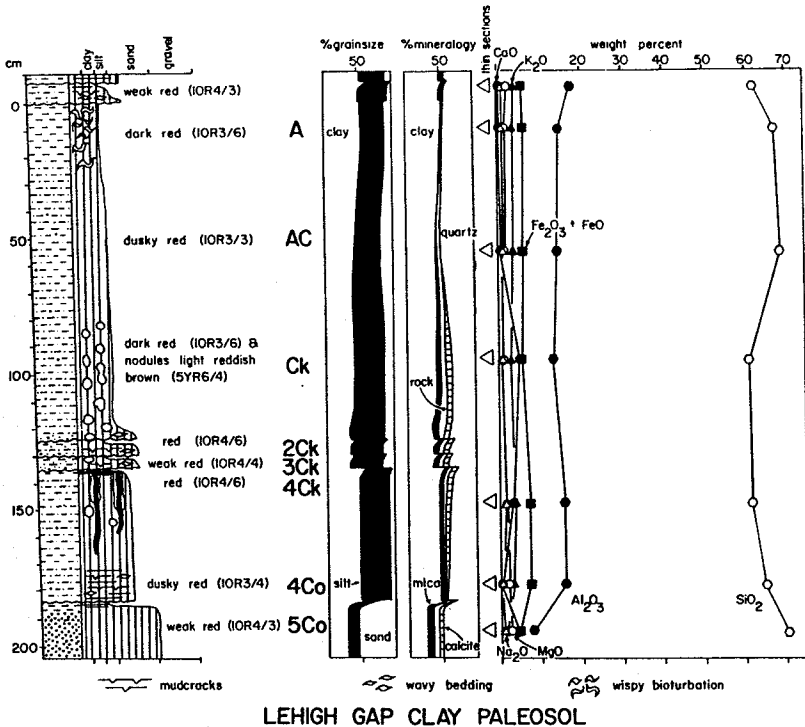


Figure 7. Columnar section, grainsize distribution and mineralogical and chemical composition of the Lehigh Gap clay paleosol, an Oxidic Ustrocept (Inceptisol), in the Late Silurian (Ludlovian), Bloomsburg Formation, near Palmerton, Pennsylvania, USA (from Retallack 1985). Unspecified symbols are as for Figure 6.

stratigraphic range of fossil millipedes at face value, this only narrows the field to amphibious chelicerates. Perhaps more detailed studies of burrows in early Palaeozoic paleosols will give a better idea of the earliest large animals on land.

Early Palaeozoic paleosols also provide some indications of how large plants and animals may have come on land. As presently understood, there is some support for the concept that the land was colonized first by microbes, then non-vascular plants, and then vascular plants. Terrestrial ecosystems appear to have been built up by degrees, becoming increasingly complex with the passage of time. This is compatible with the view that large land plants evolved from unicellular soil algae, independent of multicellular aquatic algae (Stebbins and Hill, 1980). It is also compatible with the alternative view that some aquatic plants may have invaded the land where it was prepared by

microbial or non-vascular plants. Only the idea that bare, sterile earth was colonized by large plants now seems unlikely.

The advent of large animals on land may have been similar to the extent that they did not appear there until suitable food and shelter were available. If some of the earliest animals on land were as small as modern tardigrades, they would be very difficult to detect in paleosols. In the invasion of the land by large amphibious animals, pre-existing burrows may have been important microhabitats, in addition to abundant herbiage. In a similar way, rodent burrows in modern deserts may be small, semi-autonomous communities, including also algae, fungi and beetles, all protected within the moist, cool burrow from the harsh external environment (Halffter and Matthews, 1966; Martin and Bennett, 1977).

Increased biomass of terrestrial ecosystems in waterlogged terrain resulted in the appearance of a new kind of soil, peaty soils or Histosols (of Soil Survey Staff, 1975). The oldest known Histosol is the Early Devonian (Siegenian) Rhynie Chert of Scotland, which is a petrified peat with remains of vascular land plants in growth position (Kidston and Lang, 1921). Coaly layers of large, non-vascular plants in Silurian (Llandoveryan to Ludlovian) rocks of the eastern United States (Willard, 1938; Pratt *et al.*, 1978; Strother and Traverse, 1979) do not meet the organic matter content and compacted thickness required of Histosols. They should be re-examined from this point of view.

Early Palaeozoic paleosols at major unconformities do not appear fundamentally different from Late Precambrian ones (Sharp, 1940; James *et al.*, 1961; Gariel *et al.*, 1968; Morey, 1972; Blaxland, 1973; Patel, 1977; Cummings and Scrivner, 1980). These also may have been Inceptisols and Oxisols (of Soil Survey Staff, 1975). Even less well developed Inceptisols and Entisols have been identified from Early Palaeozoic alluvial sequences (Retallack 1985). These are poorly developed soils and the existence of these orders is implied by better developed Precambrian paleosols. They are more easily recognized in Ordovician and younger rocks because of the presence of burrows and root and rhizome traces.

A cover of large land plants would have mitigated soil erosion compared to pre-existing microbial communities. The reduction spotted mounds in the surface of the Late Ordovician paleosol from Nova Scotia (Dewey in Boucot *et al.*, 1974) may represent direct evidence of this. There were probably also general effects of increased landscape stability, for example, in the increased abundance of high sinuosity, suspended load streams compared to low sinuosity, bedload streams (Schumm, 1968, 1977). The alluvial architectures of Ordovician to Permian red beds of the eastern United States show features, such

as increased abundance of clay and fining upwards cycles, which may reflect just such changes in fluvial style with time (Cotter, 1978).

AFFORESTATION OF THE LAND

Many of the changes in terrestrial ecosystems initiated with the early Palaeozoic appearance of large plants on land would have been further accentuated upon the evolution of trees. Evidence of these large plants and associated animals in paleosols can be used to assess the antiquity and nature of early woodland and forest ecosystems.

Presumably forests could have arisen during the Middle Devonian (Givetian), as large stumps of this age have been found in New York state, USA (Banks, 1980). More massive secondary wood may be almost this old (late Givetian), but it is especially common in Late Devonian (Frasnian) rocks of New York (Beck, 1964). Some of these massive woody trunks attained a diameter of 1.6 m (Beck, 1971). The actual time of origin of woodlands and forests may be hard to pinpoint, because measurement of stems of Palaeozoic fossil plants has revealed a steady increase in girth with time (Chaloner and Sheerin, 1979).

As would be expected from known Late Devonian fossil plants of New York and Pennsylvania, USA, there are abundant large root traces in rocks of that age there (Figure 8; other examples illustrated by Barrell, 1913; Walker and Harms, 1971). One of these horizons with large fossil root traces near the town of Hancock, New York, has recently been studied in detail (Figure 9: Retallack, 1985). This paleosol is very weakly developed and has conspicuous relict bedding throughout the profile. This is especially clear at the surface, which probably accumulated from floodwaters restrained by vegetation as the soil continued to form (a cumulative horizon of Birkeland, 1984). The large root traces and a fossil litter or partly decomposed leaves of *Archaeopteris halliana* within the cumulative horizon, provide evidence that this paleosol was forested. Other evidence for a forest cover is the differentiation of a laterally continuous, less clayey and iron poor surface (A) horizon over a more clayey, purple and iron rich, subsurface (B) horizon. This horizonation is so weakly expressed that the paleosol is identified as an Inceptisol (Soil Survey Staff, 1975). Translocation of clay and iron to subsurface horizons (lessivage) is partly mechanical, related to the penetration of roots and activity of soil fauna. Soil materials such as iron can also be translocated by the chemical action of rainfall leachates from leaves and of organic acids generated by decay of leaf litter (Fisher and Yam, 1984). Some of the most effective substances in this respect are phenols, manufactured by plants for defence against herbivores and pathogens. Such substances



Figure 8. Large fossil root traces (disturbed clayed areas) in a paleosol in the Late Devonian (Frasnian) Oneonta Formation, in a roadcutting on Interstate Highway 88, 1 km west of the Unadilla exit, New York State, USA. Hammer is for scale.

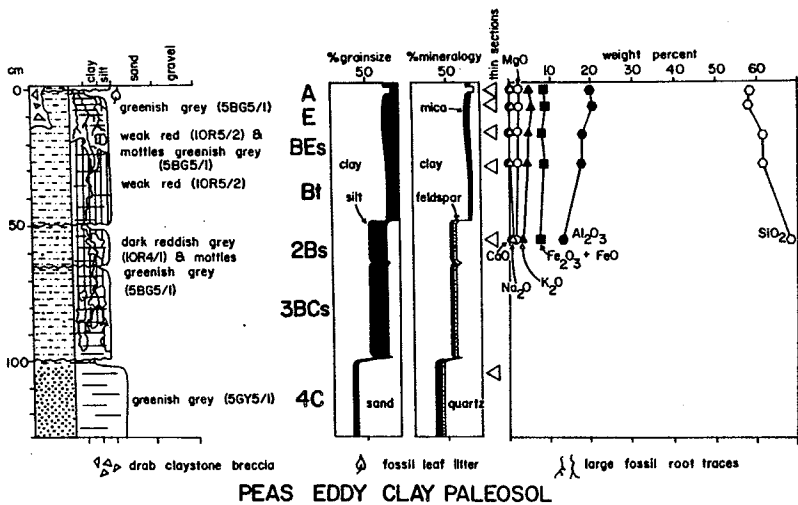


Figure 9. Columnar section, grainsize distribution and mineralogical and chemical composition of the Peas Eddy clay paleosol, a Tropaquept (Inceptisol), in the Late Devonian (Frasnian) Walton Formation, near Hancock, New York State, USA (from Retallack 1985). Unspecified symbols are as for Figure 6.

were probably widespread among early land plants, including *Archaeopteris*, just as they are in living plants (Swain and Cooper-Driver, 1981). This and other Late Devonian paleosols with large root traces in New York and Pennsylvania are non-calcareous and associated with sandstone of former stream channels. In contrast, red nodular calcareous paleosols away from palaeochannel deposits, lack large root traces and differentiation of A and B horizons. Thus, these early forests appear to have been restricted to streamside galleries, dissecting low shrubby or herbaceous vegetation of interfluves (as long ago proposed by Barrell, 1913). This distribution of early trees cannot be entirely attributed to their pteridophytic reproduction (Beck, 1971), because seed plants had not yet evolved (Gillespie *et al.*, 1981). Perhaps their demand for water and nutrients was not satisfied by soil microbiota or water in soils away from streams.

This early forested paleosol appears to be an Inceptisol, but well developed forested paleosols, such as Alfisols, Ultisols and Spodosols (of Soil Survey Staff, 1975) probably also date back to late Palaeozoic time. These each have laterally continuous A and B horizons, and include soils loosely called 'podzolic' in prior classifications (for example, Marbut, 1935; Kubiena, 1953). The USDA orders of forested soils are not always easy to distinguish in paleosols, because they are based on features such as pH and base status which must be inferred rather than measured (Retallack, 1981a, 1983a).

Alfisols are high base status, forested soils and are easily identified when calcareous and clayey. Sandy and non-calcareous Alfisols are known, and paleosols like this must be judged by the abundance of easily weathered minerals remaining in the profile, the nature of likely original clays and the abundance of elements likely to have been dominant cations (K_2O , Na_2O , CaO and MgO). Although not identified as such, it is likely that Alfisols are present among strongly developed, calcareous paleosols described from the Late Devonian (Fammenian), Aztec Siltstone, in Victoria Land, Antarctica (McPherson, 1979). Other possible fossil Alfisols, not yet studied in detail, are below the Early Permian (Leonardian) Minnekahta Limestone, South Dakota (Retallack, 1981a), within the Late Triassic (Carnian-Norian) Chinle Formation of Arizona (Retallack, 1981a), and in the Early Eocene (Wasatchian) Willwood Formation of Wyoming (Bown and Kraus, 1981a,b). The oldest fossil Alfisol studied adequately and identified as such, appears to be the Paleudalf (Interior clay paleosol) in the late Eocene (early Chadronian), Chadron Formation, Badlands National Park, South Dakota (Retallack, 1983a,b).

Ultisols are base poor, clayey, forested soils. The author suspects that very early Ultisols may be found in clayey sediments in the

unconformity between the Middle Ordovician (Whiterockian) Jefferson City Dolomite and the Late Carboniferous (Desmoinesian or Westphalian) Fort Scott Limestone in the 'diaspore clay region' of south-eastern Missouri, USA (Keller *et al.*, 1954). A variety of paleosols and sediments were recognized on this ancient karsted landscape. In places there are leached, grey clays (possible A horizons) grading down into reddish clays (possible B horizons). The clays contain abundant diaspore, and some boehmite, goethite and lepidocrocite. Alkalis and alkaline earths ($\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{MgO} + \text{CaO}$) in this material total only 1.15 weight percent. This and the mineralogy of the clays, provides evidence of prolonged deep weathering of this material and very low base status. Early Triassic 'paleoplanosols' of southern France (Lucas, 1976) may also have been Ultisols. The oldest Ultisol actually identified and described as such is a partly redeposited profile, the Yellow Mounds paleosol, developed between the Late Eocene (Duchesnean) Slim Buttes Formation and Late Eocene to Early Oligocene (Chadronian) Chadron Formation, in Badlands National Park, South Dakota, USA (Retallack, 1983a,b).

Spodosols are acidic sandy soils, with a B horizon cemented by sesquioxides and organic matter (spodic horizon of Soil Survey Staff, 1975). The most promising place to search for the oldest Spodosols is among British Carboniferous ganisters: silicified sandstones mined for silica bricks and furnace linings. To date, however, there have been no detailed reports of iron or organic cemented horizons within ganisters which would qualify as spodic horizons in the USDA classification. Local ferruginized and carbonaceous zones have been noted in the Late Carboniferous (Westphalian) Sheffield Blue Ganister, near Sheffield, England (Searle, 1930; Percival, 1983 and this volume). This is tentatively accepted as the oldest likely Spodosol. The Avalon Series are comparable ganister-bearing paleosols from the Early to Middle Triassic (Scytho-Anisian) Newport Formation, near Sydney, Australia (Retallack, 1976, 1977). These have well differentiated sandy eluvial horizons (ganisters), but clear spodic horizons have not been recognized other than weakly enriched zones of amorphous organic matter within underlying siltstones. The author originally regarded prominent sub-surface horizons of iron carbonate (siderite) nodules in underlying siltstones as B horizons of these profiles. These are not spodic horizons, and there do not appear to be similar features in any reported modern Spodosols. Siderite implies a pH rather more neutral to alkaline than is typical for Spodosols. The nodules may be better regarded as products of waterlogging below or subsequent to burial of the original soil (deep gley or diagenetic pseudogley). Perhaps the British and Australian ganisters represent an extinct kind of soil formed under less acidifying

vegetation than modern Spodosols. There are other problems with the Long Reef clay paleosol, within the Early Triassic (Scythian) Bald Hill Claystone in the same area (Retallack, 1976, 1977). In petrographic thin sections of the B horizon of the type profile an opaque, sesquioxide stain cements sand grains, as it should in a spodic horizon. However, the grains are for the most part weathered volcanic rock fragments, and the A horizon is much more clayey than usual for Spodosols. More detailed analytical work is warranted on both of these possible Early Triassic Spodosols. The oldest unquestionable Spodosols are between the Late Eocene (Bartonian) Sables de Beachamps and Formation d'Ezanville, near Paris, France (Pomerol, 1964, 1982). The profile at Ermenonville (Pomerol, 1982, Figure 3.24) appears to be an Orthod (of Soil Survey Staff, 1975).

From this preliminary outline of the appearance of the main kinds of well-drained forested soils on Earth, it appears that base-rich, alkaline, forested soils may have predated acidic, highly weathered ones. Whether this is a reflection of increasingly stable forest communities, or of palaeoclimatic and other peculiarities of the few areas best studied, remains to be established by further work. Fortunately, there is no shortage of well developed, red, Late Palaeozoic and Mesozoic paleosols (Hubert, 1960; Danilov, 1968; Kabata-Pendias and Ryka, 1968; Chalyshev, 1969; Power, 1969; Ortlam, 1971, 1980; McBride, 1974; Meyer, 1976; Sturt *et al.*, 1979; Annovi *et al.*, 1980; Franks, 1980; Retallack and Dilcher, 1981a,b; Ortlam and Zimmerle, 1982; Besly and Turner, 1983).

The advent of trees had substantial consequences for Histosols, the peaty soils of swamps and marshes (Soil Survey Staff, 1975). Moderately thick (10 cm) coal overlying claystone with large root traces has been found as old as Late Devonian (Fammenian) in Virginia, USA (Gillespie *et al.*, 1981). Much thicker organic horizons (later altered to coal) of wetland soils were widespread during Carboniferous and later time. Drab paleosols of the Euramerican Carboniferous coal measures have been given a number of local names, such as underclay, seat earth, fire clay, tonstein and ganister (Huddle and Patterson, 1961; Roeschmann, 1971; Feofilova, 1973, 1977; Feofilova and Reksinskaya, 1973). A number of Permian and Mesozoic fossil Histosols have also been reported (Retallack, 1976, 1977, 1980; Retallack and Dilcher, 1981a,b).

In addition to gains in biomass of terrestrial ecosystems there were also gains in complexity and diversity of life associated with woodlands and forests. There is a distinct Late Carboniferous to Early Permian peak in the diversity of fossil land plants (Tiffney, 1981). By this time also, many of the most important soil animals had evolved, including

mites, collembollans, centipedes, insects, land snails, amphibians and reptiles (Rolfe, 1980; Milner, 1980). Increased diversity of soil fauna with time is also indicated by studies of trace fossils in paleosols. A low diversity assemblage of burrows and faecal pellets has been found in a thin paleosol A horizon in the Early Carboniferous (Viséan), Llanelly Formation, of South Wales, UK (Wright, 1983). Burrows of three kinds, including those of earthworms, cicada-like insects and vertebrates were recognized in paleosols formed under woodland and heath in the Early to Middle Triassic (Scytho-Anisian), Newport Formation, near Sydney, Australia (Retallack, 1976). Nine kinds of trace fossils have been found in woodland paleosols of the Early Eocene (Wasatchian), Willwood Formation, of Wyoming (Bown and Kraus, 1983). Beautifully preserved nests of termites and 15 other kinds of animal burrows were found in forested paleosols of the Oligocene, Jebel Quatrani Formation, in the Fayum Depression of Egypt (Bown, 1982). Five kinds of trace fossils, including indications of dung beetles and sweat bees were recognized in paleosols of woodland, savanna and grassland in the Oligocene White River and basal Arikaree Groups, in Badlands National Park, South Dakota (Retallack, 1983b, 1984b). In addition to the three kinds of vertebrate burrows found in Early Miocene desert paleosols of the Harrison Formation of north-eastern Nebraska (Martin and Bennett, 1977), the author can personally testify to the existence of an equivalent number of invertebrate burrows in these paleosols. These few accounts represent nearly all the descriptions of trace fossils explicitly recorded from paleosols. This promising source of information on the evolution of soil fauna has been sadly neglected, compared to trace fossils of other environments (Ratcliffe and Fagerstrom, 1980).

Forests would also have had a greater stabilizing effect on the landscape than pre-existing vegetation. Erosion of land protected only by primitive land plants may be the reason why Silurian marine red beds are so common, whereas there are few red beds with marine fossils in younger rocks (Ziegler and McKerrow, 1975). The author has seen very few sequences of Palaeozoic paleosols (setting aside the spectacular Devonian '*Psammosteus*' Limestones of the Welsh Borderland: Allen, 1974b) which have comparably well developed profiles to the red and variegated badlands of non-marine early Tertiary rocks in western North America (Figure 12: Bown and Kraus, 1981a,b; Retallack, 1983a,b). There are very few pre-Carboniferous coals, and even the Euramerican Carboniferous coal measures have more and thicker clastic partings than the spectacularly thick early Tertiary coals of North America, East Germany and south-eastern Australia (Schumm, 1968). Although the depth of weathering observed

on Precambrian and early Palaeozoic unconformities may be impressive (Morey, 1972; Gay and Grandstaff, 1980), it does not rival some late Mesozoic and early Tertiary lateritic weathering profiles (McFarlane, 1976).

Even today the land is not completely forested, and more ancient kinds of soils persist in other parts of the landscape. Weakly developed paleosols (Entisols and Inceptisols of Soil Survey Staff, 1975), recognized mainly by their root traces, are found in many late Palaeozoic and Mesozoic alluvial sequences (Allen, 1947, 1959, 1976; Batten, 1973; Grigor'ev, 1973; Hlustik, 1974; Retallack, 1976, 1977, 1979, 1980, 1983c; Retallack & Dilcher, 1981a,b). Late Palaeozoic and Mesozoic calcretes, formed within a variety of aridland soils such as Inceptisols, Aridisols and Alfisols, are largely calcitic, rather than dolomitic like older calcretes (Blom, 1970; Allen, 1973, 1974a,b,c; Freytet, 1973; Steel, 1974; West, 1975, 1979; Folk and McBride, 1976; Lucas, 1976; Watts, 1976, 1978; Hubert, 1977a,b; Sochava, 1979; Adams, 1980; Adams and Cossey, 1981; Boucot *et al.*, 1982; Freytet and Plaziat, 1982; Wright, 1982; Parnell, 1983; Blodgett, 1984). Changes in water table and in rainfall have been interpreted from the morphology and depth of karst topography and residual paleosols on late Palaeozoic and Mesozoic limestones (Keller *et al.*, 1954; Dunham, 1969; Faugeres and Robert, 1969; Bernoulli and Wagner, 1971; Maiklem, 1971; Wardlaw and Reinson, 1971; Bosellini and Rossi, 1974; Walkden, 1974; Walls *et al.*, 1975; Goldbery, 1979; Poty, 1980; Wright, 1981, 1982; Buchbinder *et al.*, 1983). Bauxites are indicators of continuously wet tropical climates and laterites of wet seasonal climates, and both are widespread on late Palaeozoic and Mesozoic land surfaces (Goldich, 1938; Sloan, 1964; Sombroek, 1971; Valetton, 1972; Loughnan, 1975; Philobos and Hassan, 1975; Singer, 1975; Blank, 1978; Goldbery, 1979; Abed, 1979; Nicholas and Bildgen, 1979). Silcretes may be associated with bauxitic or lateritic deep weathering or form in deserts, and these also are known in Palaeozoic, Mesozoic and Tertiary rocks (James *et al.*, 1968; Dury and Habermann, 1978; Selleck, 1978; Wopfner, 1978; Rubin and Friedman, 1981; Isaac, 1983b). Strong seasonality is indicated by Vertisols, which also have been found in non-marine rocks of various ages through Phanerozoic time (McBride *et al.*, 1968; Jungerius and Mûcher, 1969; Allen, 1973, 1974b; Galloway, 1978; Goldbery, 1982a,b). Persistence of these ancient kinds of paleosols and soil features along with forested soils is evidence that the diversification of soils with time was more a process of addition of new kinds of soils, than of replacement of pre-existing kinds of soils.

GRASSES IN DRY CONTINENTAL INTERIORS

Before the advent of grasses, dry regions of the world were probably vegetated by a variety of woody plants. Many of these may have been even more bizarre than the Joshua tree (*Yucca brevifolia*), Boojum (*Idria columnaris*) and Saguaro cactus (*Cereus giganteus*) of the North American Southwest. Although large plants and animals play a conspicuous role in such desert ecosystems, there is much bare earth exposed. Soils forming in such environments (Aridisols of Soil Survey Staff, 1975) do not appear greatly different from calcareous paleosols as old as 1900 million years. The advent of grasses in subhumid to semi-arid plains of continental interiors, and in physiologically dry mountain regions, signalled the beginning of a new kind of ecosystem and new kinds of soils, the Mollisols. These are soils with nutrient rich, well structured surface horizons (mollic epipedon of Soil Survey Staff, 1975). Nutrients are not leached from these soils as is usual in humid climates, nor slowly redistributed into duricrusts as in very dry climates. Instead, a dense, low growth of herbaceous, annual grasses, abundant soil invertebrates and a variety of large mammals recycle nutrients to such an extent that these are some of the most productive ecosystems on land.

Evidence for grasslands among plant and animal fossils is poor and indirect. Grasses are not fossilized in the dry, well-drained environments of most grasslands. The direct fossil record of grasses is confined to swampy and near-stream habitats. Fossil grass pollen can be traced securely back to Palaeocene time (Muller, 1981) and megafossil grasses are known as old as Eocene (Daghlian, 1981). Grass fossils found within Late Miocene and Pliocene paleosols are largely biogenic silica encrustations (Thomassen, 1979) which may have coevolved with mammalian grazers long after the appearance of grasslands (Stebbins, 1981). There are similar problems with other adaptations thought to indicate former grasslands, such as the high-crowned cheek teeth (hypsodonty) and elegant slender limbs (cursoriality) of Tertiary mammals. Such skeletal evidence is an indication that grasslands were already established in South America by Eocene time (Webb, 1978), in North America by early Miocene (Webb, 1977), and in Africa, Eurasia and Australia by Late Miocene to Pliocene (Van Couvering, 1980; Sanson, 1982; Flannery, 1982). As a completely independent line of evidence for grasslands, paleosols may reveal not only the antiquity of grasslands, but how they developed.

There is abundant paleopedological evidence that woodland and forest was much more widespread than at present during late Mesozoic and early Tertiary time. This can be seen especially from the wide distribution of laterites and deeply weathered paleosols of this age

(Pettyjohn, 1966; Sombroek, 1971; Thompson *et al.*, 1972; Chernyakhovskii and Khosbayar, 1973; Abbott *et al.*, 1976; McFarlane, 1976; Nilsen, 1978; Nilsen and Kerr, 1978; McGowran, 1979; Singer and Nkedi-Kizza, 1980; Abbott, 1981; Thompson *et al.*, 1982; Isaac, 1983a). In Tertiary alluvial sediments, also, paleosols formed under woodlands and forests are abundant and widespread (Pomerol, 1964, 1982; Morand *et al.*, 1968; Reffay and Ricq-Debouard, 1970; Ritzkowski, 1973; Braunagel and Stanley, 1977; Buurman, 1980; Bown and Kraus, 1981a,b; Retallack, 1981b,c; Winkler, 1983).

A transition from forest to woodland, savannah and open grassland has been documented in volcanoclastic Late Eocene to Oligocene alluvium of the White River and lower Arikaree Groups in Badlands National Park, South Dakota, USA (Figure 10: Retallack, 1983a,b). Under an initially humid Late Eocene climate a forested Ultisol (of Soil Survey Staff, 1975) developed on an unconformity of smectitic Late Cretaceous (Maastrichtian) marine shale and thin residuals of Late Eocene (Duchesnean) alluvial sediments. In subhumid climates of the Eocene–Oligocene transition (early Chadronian), the first soil formed on the volcanoclastic alluvium was a strongly developed Paleudalf. This was followed by numerous superimposed Early



Figure 10. The Pinnacles area, Badlands National Park, South Dakota, USA. Above the weathered Cretaceous marine shale (lower foreground) are 87 superimposed paleosols of Late Eocene and Oligocene age within 143 m of alluvial deposits. These spectacular, colourful outcrops show well the field appearance of sequences of clayey paleosols.

Oligocene (Chadronian) Paleustalfs (for example, Figure 11). Judging from the laterally continuous, drab, surface (E) horizons and abundant large, drab-haloed root traces in these fossil Alfisols, they supported woodland vegetation. By early Late Oligocene (Orellan and Whitneyan) time there were a variety of soils, and the climate appears to have been distinctly drier and more seasonal. Paleustalf paleosols at this stratigraphic level have iron stained calcareous stringers (petrocalcic horizons) and are closely associated with sandstone palaeochannels. Red Fluvents (Entisols), showing relict bedding and only small root traces, are interbedded with stream deposits. Away from palaeochannels are Andic Ustochrept (Inceptisol) paleosols, in which there are limited development of A and B horizons, abundant small root traces, fine granular structure (peds), but only scattered large, drab-haloed root traces (Figure 11B). Early Late Oligocene vegetation is interpreted as a mosaic of early successional herbaceous vegetation in streamside swales, streamside gallery woodland and widespread savannah on the interfluvies. In late Late Oligocene (Arikarean) rocks only one kind of paleosol, identified as Fluvaquentic Eutrochrept (Inceptisol), has large drab haloed root traces. These are interbedded with the sandstones of deeply incised streams. Other paleosols at this stratigraphic level all have simple (A-Ck) profiles and abundant small root traces. Some of these are light coloured Calciorthids (Aridisols: Figure 11), and others with darker surface horizons were identified as Ustollic Eutrandepts (Inceptisols). During late Late Oligocene time, trees appear to have been confined to stream margins within deep erosional gullies, whereas the open floodplain depressions and dry areas supported open grassland. Sediments at this stratigraphic level in the badlands are more calcareous and less clayey than older sediments. The climate was becoming increasingly dry, and was probably semi-arid by Late Oligocene time. Climatic drying can be attributed to the lengthening rain shadow cast by the Rocky Mountains to the west, but was also related to global climatic changes at this time (Wolfe, 1978; Kennett, 1982).

This sequence of paleosols can be used to assess a number of factors proposed for the maintenance of grasslands over other kinds of vegetation (Vogl, 1974; Walker *et al.*, 1981). For example, grasses flourish over trees in dry or otherwise unfavourable, highly seasonal and unpredictable environments. This is because of their small stature, unspecialized pollination and dispersal mechanisms, and protection of most of their tissues in rhizomes and other underground structures. Climatic drying is a plausible explanation for the origin of grasslands in Badlands National Park. There is good evidence for increasingly dry climate with the appearance of grasslands, as far as can be judged from

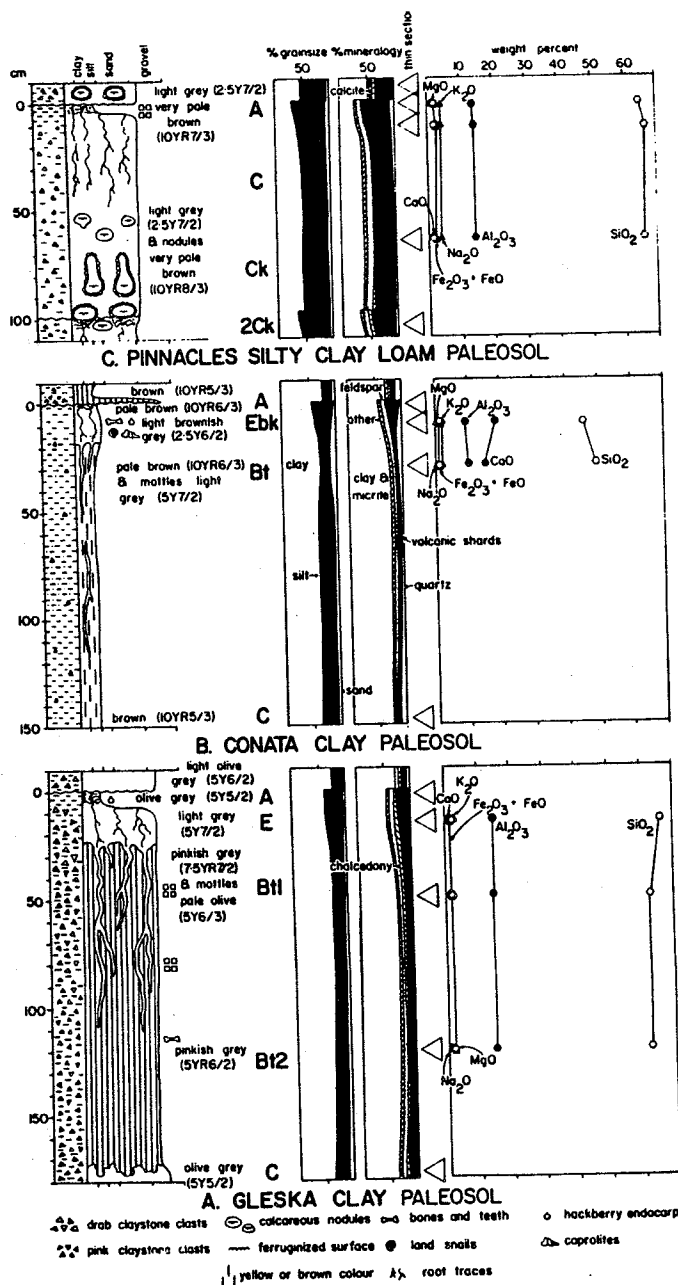


Figure 11. Columnar sections grainsize distributions and mineralogical and chemical compositions of C, the type Pinnacles silty loam paleosol, a Calcicorthid (Aridisol) in the late Late Oligocene (Arikarean) Sharps Formation; B, the type Conata clay paleosol, an Andic Ustochrept (Inceptisol) in the early Late Oligocene (Orellan) Scenic Member of the Brule Formation; and A, the type Gleska clay paleosol, an Udic Paleustalf (Alfisol) in the early Oligocene (Chadronian) Chadron Formation, in Badlands National Park, South Dakota, USA (adapted from Retallack 1983b). Unspecified symbols are as for Figure 6.

the less severe weathering of volcanoclastic parent material and the shallower depth of calcareous horizons in paleosols with time.

Grasses also recover more quickly from fire than trees, and many modern grasslands are thought to be maintained by periodic fires. This is a difficult feature of ancient environments to interpret from grassland paleosols, because phytoliths and ash left by grass fires may appear no different in the soil than organic and mineral matter left by ordinary decay of grasses. The burning of trees, on the other hand, creates charcoal which can be distinguished from coalified wood, and is known to persist in paleosols under a wide variety of conditions (Retallack, 1984a). No charcoal was seen in savannah and woodland paleosols of the sequence in Badlands National Park. Although fire may have played a role in the maintenance of grassland once it appeared, it does not appear to have been important in the development of grassland from pre-existing savannah and woodland. The author suspects that fires were much less frequent and important to grasslands before the advent of humans.

Animal activity may also play a role in the maintenance of modern grasslands, but there is little evidence of this among fossil mammals of Badlands National Park. A new mammalian fauna appreciably better adapted to open country than Eocene or Paleocene Faunas of North America, appeared during Early Oligocene time, when the area was mostly wooded, judging from fossil soils. This evidence of woodland, independent of fossil mammals, confirms long-held suspicions that these creatures invaded from savannahs elsewhere, rather than evolving locally (Clark *et al.*, 1967; Emry, 1981). These faunas thrived in savannah environments which appeared during early Late Oligocene time. Surprisingly, they also persisted into late Late Oligocene time, when trees were very restricted in distribution, and most of the area was open grassland. At this time there were many mammals smaller than their presumed ancestors and more burrowing animals, so the observed environmental deterioration did have some effect on fauna. However, most endemic evolutionary lineages persisted, diversity remained much higher than usual for open grasslands and their tooth and limb structure was not appreciably better adapted to open country than before (MacDonald, 1963, 1970). This fauna was eventually replaced by a different assemblage of mammals, including a number of new evolutionary lineages with clear open country adaptations, during the very latest Oligocene, a time of extreme desertification in the Great Plains (Webb, 1977). Since open grasslands appeared several million years before mammalian faunas were especially well adapted to them, it is unlikely that grazing pressure was the most important factor in the origin of open grasslands in Badlands National Park.

A final factor encouraging development of grasslands is competition between plants. Although trees may have an advantage in shading out and chemically poisoning lower growing plants, grasslands can still persist by choking out small seedlings of trees. Plant competition is unlikely to have been a significant factor in the Badlands, because the oldest grassland soils there formed in very dry climates. These would have been marginal even for modern grasses, as in much of the intermontane rangelands of the present North American West. Well structured portions of these Late Oligocene paleosols are variable in thickness. This may be an indication that grasses and forbs of these early prairies were clumped, and that plant competition was less important than being able to tolerate dry conditions.

It remains to be seen whether grassland in other parts of the world appeared under similar circumstances. There is a comparable sequence of paleosols in Eocene and Oligocene volcanoclastic alluvial deposits of Argentina. These contain abundant fossil mammals, phytoliths and trace fossils of invertebrates, including dung beetles (Frenguelli, 1939; Andreis, 1972; Spalletti and Mazzoni, 1978). An early Oligocene (Deseadean) paleosol near Paso Flores, Neuquén, has been briefly characterized (Frenguelli, 1939, p. 344). It may have been an Inceptisol or Alfisol (of Soil Survey Staff, 1975). Other paleosols of about the same age near Lago Colhue Huapi, Chubut, have been identified as Udolls and Aquolls (Spalletti and Mazzoni, 1978). These are the oldest Mollisols yet recorded. Since high crowned teeth and slender limbs are found in Eocene mammals of Argentina (Webb, 1978), it is likely that older Mollisols will be reported from there.

Perhaps, as appears to have been the case in South Dakota, the oldest grassland soils in Argentina will turn out to belong to different soil orders, such as Inceptisols. Similarly, the oldest forested paleosol now recognized cannot be attributed to one of the modern orders of forested soils. The early stages in the evolution of ecosystems may be very different, perhaps less integrated than their modern equivalents. The general question of coevolution within ecosystems is another issue for which paleosols may provide useful information.

Grasslands are only one of a number of plant communities poorly represented in the record of fossil plants, because they grow far from waterlogged sedimentary environments favourable for plant preservation (Retallack, 1984a). Clues to the geological history of chaparral, desert and alpine vegetation eventually may be gleaned from the study of paleosols.

PALEOSOLS AND HUMAN EVOLUTION

Even if our bones and the debris of civilization soon become diagnostic

fossils of one of the briefest biostratigraphic zones in geological history, our effect on land surfaces of the world is already conspicuous and irreversible. Modern cities, dams, parking lots and highways are reshaping the landscape (Bidwell and Hole, 1964). Introduction of carbon that accumulated in the Earth over millions of years into the atmosphere as carbon dioxide, spreading of peculiar local phosphate deposits over fertile bottomlands and burial of various bizarre chemical wastes will have longer term effects on the nature of the Earth's surface; effects which already seem alarming. The writings of Plato reveal that some of these effects, such as deforestation and soil erosion, have long been a concern of intellectuals (Glacken, 1956). Presumably there was once a 'Golden Age,' though perhaps not as Plato imagined it, when human impact on the landscape was minimal and when many aspects of our character were imposed by an environment beyond our control. Evidence of that time and our long evolutionary career since, can be sought from paleosols, as well as from fossils of our early ancestors and the remains of their cultures.

A variety of extinct Miocene ape-like fossils have been considered ancestral to both humans and the living great apes. Ramapithecines especially have aroused interest, because some had short canines and thick-enameled molars arranged in divergent rows: intermediate in character between apes and humans. Recently discovered skulls of these creatures from Pakistan and China, on the other hand, indicate that Asiatic ramapithecines are more likely to have been ancestors of orangutans (Pilbeam, 1984). This interpretation is compatible with palaeoenvironments interpreted from paleosols associated with these fossils in India and Pakistan (Johnson, 1977; Johnson *et al.*, 1981). The author has made detailed studies of paleosols in the Late Miocene part (palaeomagnetically dated at 8.3 million years by Tauxe and Opdyke, 1982) of the Dhok Pathan Formation, Siwalik Group, near Khaur village in Northern Pakistan (Behrensmeyer and Tauxe, 1982; Retallack, 1985). Exceptionally well preserved ramapithecine fossils have been found here, including a face of *Sivapithecus* (Pilbeam *et al.*, 1980; Pilbeam, 1982). Former vegetation may be interpreted from six different kinds of paleosols (Retallack, 1985). Intimately associated with buff-coloured channel deposits of former streams draining Himalayan foothills, are paleosols with numerous large root traces and burrows, and obvious sedimentary relicts, such as bedding. These probably supported well drained, streamside, early successional woodlands. A second kind of paleosol associated with buff palaeochannels is brown to yellow and has a prominent zone of manganese staining below the surface (placic horizon of Soil Survey Staff, 1975). These may have supported early successional vegetation in poorly drained

swales, associated with stream margins. A third kind of paleosol is thick, copiously bioturbated, clayey and red, with well differentiated subsurface (B) horizon and large root traces (Figure 12). It presumably formed under well drained streamside gallery forest. A fourth kind is yellow to brown and silty. The abundant fine root traces, rare large ones and weakly developed subsurface clayey (B) horizon are all features of savannah soils. These paleosols are interbedded with laterally impersistent red paleosols of the third kind. Thus, vegetation of dry floodplains may be more accurately spoken of as savannah groveland. The two remaining kinds of paleosols are associated with large, grey-green, sandstone palaeochannels, thought to represent the ancestral Indus River, which had a source, then as now, well within metamorphic rocks of the rising Himalayan Mountains. Both kinds of paleosols are grey in colour, contain large root traces, and presumably formed under different kinds of bottomland forest. Clayey, deeply cracked and partly brecciated, grey paleosols probably formed in seasonally dry swamps. Silty paleosols with abundant calcareous nodules presumably formed in levees and other slightly elevated parts of these alluvial bottomlands. This concept of the Late Miocene vegetative mosaic is well in accord with the nature of the abundant

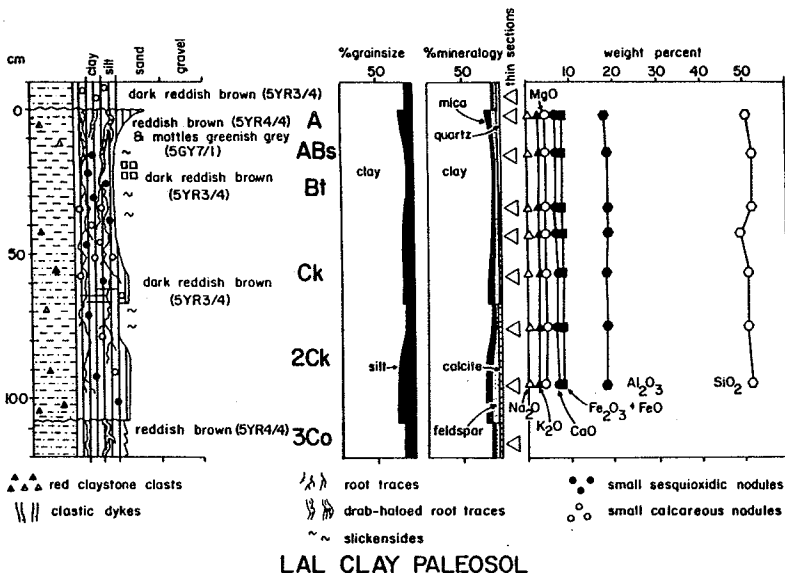


Figure 12. Columnar section, grainsize distribution and mineralogical and chemical composition of the Lal clay paleosol, an Oxic Haplustalf (Alfisol) in the Late Miocene, Dhok Pathan Formation, near Kaulial village, northern Pakistan (from Retallack 1985). Unspecified symbols are as for Figure 6.

vertebrate fauna known from these rocks, which includes crocodiles, elephants, antelopes, pigs and three-toed horses (Badgley and Behrensmeyer, 1980). Ramapithecines were very rare elements of these tropical faunas. They have only been found in palaeochannel deposits and paleosols of associated levees. These paleosols are of the first kind mentioned, and have been identified as Fluvents (of Soil Survey Staff, 1975). Presumably ramapithecines lived within these early successional woodlands and in adjacent gallery forests on the red (type 3 or Oxic Haplustalf) soils (Retallack, 1985). The life of these orangutan-like creatures would have had little impact on their environment, but these creatures were already a separate branch of the human evolutionary tree (Pilbeam, 1984). For evidence of the transition from ape to human, the more complete record of human evolution in East Africa is more promising.

During recent fieldwork in Kenya and Tanzania, the author was able to confirm reports of paleosols at a number of well known late Cenozoic vertebrate fossil localities. In Early Miocene (about 18 million years before present) localities for dryopithecine apes on Rusinga Island and near the villages of Songhor and Koru in south-western Kenya (Andrews, 1981; Pickford and Andrews, 1981), abundant red paleosols are broadly similar to those thought to have formed under tropical forest in northern Pakistan. New kinds of soils formed in this region after uplift of the western margin of the East African Rift and extrusion of plateau phonolites by Middle Miocene time, some 14 million years ago. These are well illustrated by paleosols near the village of Fort Ternan, south-western Kenya (Shipman *et al.*, 1981). A paleosol presumably formed under woodland ('Braunlehm' of the classification of Kubiena, 1953) has been identified from here (Bishop and Whyte, 1962). This paleosol is poorly fossiliferous and situated stratigraphically below the most fossiliferous paleosols which are of a quite different type (Figure 13). These have simple (A-Ck) profiles, scattered large root traces and a dark granular surface layer like that of Mollisols. They probably formed under open grassy woodland or savannah, an interpretation amply supported by the nature of the fossil mammalian fauna from here (Evans *et al.*, 1981; Shipman *et al.*, 1981). The ramapithecine *Kenyapithecus* also has been found at Fort Ternan. Its ecology and habitat here, 14 million years ago, was probably very different from that of *Sivapithecus* in northern Pakistan, 8 million years ago. These preliminary findings underscore the need to obtain fossils of *Kenyapithecus* better than the few teeth and fragments of jaw now available. Perhaps it and savannah environments played some role in early human evolution? These ideas are neither new nor satisfactorily answered (Pilbeam, 1984). Evidence from paleosols may aid in guiding future efforts to answer them.



Figure 13. Middle Miocene (about 14 million years old) paleosols interbedded with tuffs in the main excavation at Fort Ternan National Monument, Kenya. The top of the upper paleosol is at the hammer head, and only the top of the lower paleosol is exposed at the base of the excavation. Both paleosols were probably Mollisols.

Paleosols now are recognized in association with fossil hominids from the East African Rift Valley from Ethiopia to Tanzania and ranging back as old as 4 million years (Hay, 1976; Leakey *et al.*, 1978; Aronson and Taieb, 1981; Burgraff *et al.*, 1981; White *et al.*, 1981; Cohen, 1982). In Plio-Pleistocene sediments of the East Turkana region of Kenya, and Olduvai Gorge, in Tanzania, studies of paleosols may be useful in interpreting the habitat preferences of sympatric species of hominids. Numerous root traces and paleosols have been recognized in the main sequence (Beds I to IV) of Olduvai Gorge, some 2.1 to 0.5 million years old (Figure 14). These sediments accumulated on alluvial uplands and the margins of alkaline lakes whose variation in extent and location with time has been beautifully documented by Hay (1976). During his own examination of these rocks, the author was especially struck by their palaeoenvironmental similarities with modern Amboseli National Park, Kenya (Williams, 1972; Sombroek *et al.*, 1982). Hay's lake margin sequences include zeolitic paleosols, which were presumably alkaline Entisols supporting scrubby, salt tolerant vegetation. Also in lake margin facies are dark granular paleosols with calcareous subsurface layers, which may have been Mollisols, formed under grassland and savannah. Hay's alluvial facies



Figure 14. The Plio-Pleistocene (2.1 to 0.5 million years old) sequence in Castle Rock (foreground) and Olduvai Gorge (background), northern Tanzania. Red paleosols form the thick dark layer (Bed III) in the upper part of Castle Rock. Underlying drab sediments include paleosols and lake deposits of Beds I and II.

is a prominent red unit (Bed III) composed of a number of well drained, calcareous paleosols, perhaps Inceptisols formed under open woodland. Calcrete-bearing paleosols have been found in the overlying Masek, Nduvu and Naisusu Beds (Hay and Reeder, 1978), which appear to have formed in a more arid and less stable environment. Stone tools are associated with hominids in Olduvai Gorge and structures which may have been associated with permanent encampments have been found at low stratigraphic levels (Bed I; Hay, 1976). Chemical and petrographic studies of occupation floors could prove revealing. One of the later appearing hominid species (*Homo erectus*) found in Olduvai Gorge is thought to have discovered the use of fire by 0.7 million years ago in other parts of the world (Pilbeam, 1984). The use of fire, permanent camps and organized hunting are ways in which even primitive peoples may alter their environment. In some ways their effects may mimic climatic drying. This has been a persistent problem in untangling human and climatic effects in the origin of deserts, such as the Sahara (Huzayyin, 1956; Williams, 1979; Alimen, 1982). Perhaps the paleosols of Olduvai Gorge or other thick sequences of Pleistocene paleosols (Kukla, 1977; Firman, 1979) will some day

yield evidence on the antiquity and extent of early human dominion over the landscape.

CONCLUSIONS

In this review of the diversification of soils through geological time, I have stressed not only what is known of the record of fossil soils, but also the record of fossils in them. Many non-biotic factors, such as climate, parent material, geographic setting and time, also play a role in the formation of soils, but it is organisms which give soils much of their distinctive character. The diversification of life on Earth is reflected in a diversification of soils and soil features with time (Figure 15). In some cases the link between organisms and environment was indirect and is difficult to assess, as in the likely oxygenation of Precambrian atmospheres by photosynthetic microbes and human desertification of our present environment. Some major events in the history of life on land such as the mid-Cretaceous dispersal and rise to dominance of angiosperms and the terminal Cretaceous extinction of dinosaurs do not yet appear to have had especially significant ramifications for the geological history of soils. Nevertheless, it is apparent that soils and ecosystems have developed in tandem and much can be learned about each by studying them together.

Paleosols are especially valuable in providing evidence of life in

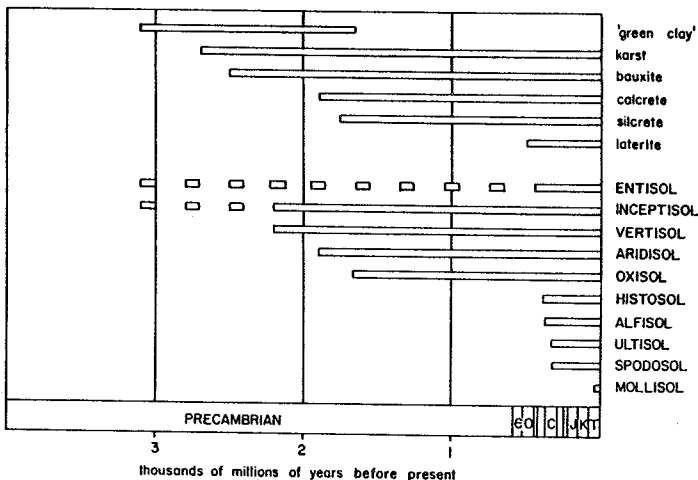


Figure 15. Geological range of soil features and USDA soil orders (of Soil Survey Staff 1975). Entisols and Inceptisols are assumed as precursors of other Precambrian paleosols, even though difficult to recognize in Precambrian rocks.

environments where fossils may not be preserved. Even though bones are not preserved in acidic paleosols, nor plants in oxidized paleosols (Retallack, 1984a), ecosystems of considerable biomass, such as tropical rainforest, commonly grow in such non-calcareous, red soils. Information from paleosols may provide clues to taphonomic biases in preserved fossil assemblages. Detailed evidence of behaviour and kinds of organisms may be obtained from the study of trace fossils in paleosols. In a sense paleosols themselves are trace fossils of ecosystems. Like other trace fossils they can be assumed to be in place and related to associated sedimentary environments. This is of value for assessing the original distribution of scattered remains of plants and animals in non-marine sediments. Fossil plants associated with paleosols may be more informative of past vegetation than mixed remains found in deposits of lakes and lagoons, although widely dispersed fossils may provide indications of vegetation not preserved in well drained paleosols (Retallack and Dilcher, 1981a). Similarly, bones associated with distinct kinds of paleosols are likely to reflect original vertebrate communities more accurately than transported remains found in fluvial palaeochannels (Bown and Kraus, 1981b; Behrensmeyer, 1982; Winkler, 1983).

Paleosols also provide evidence of former ecosystems independent of fossils. Interpretations based on paleosols can be used to check or enlarge reconstructions of the past based on fossil animals and plants. Many reconstructions using fossils alone have been based on the assumption that adaptive features of organisms were optimally suited to their environment (Gould and Lewontin, 1979). For example, open grassland environments have been inferred from assemblages of mammals with high crowned teeth and slender limbs. Using paleosols as independent evidence of environments, it is possible to recognize those evolutionarily interesting times when vegetation and animals were not well co-adapted (Retallack, 1983a,b). Degree of integration and coevolution within terrestrial ecosystems can thus be probed. The tentative early stages in the evolution of major new kinds of ecosystems are of special interest in allowing us to understand how presently complex, natural systems began.

Although research on fossil soils older than Quaternary is now accelerating, there is much to be done. The present inventory of described paleosols is especially inadequate to assess the earliest representative of various soil types and features. The author's tentative compilation (Figure 15) is based as far as possible on available data, but is also compatible with current concepts of soil forming processes and of changes in life and environments on land with time. The author is very conscious of the risk that such concepts may bias the way in

which paleosols are interpreted. As more paleosols become known, they will be better assessed on their own terms. Perhaps then a history of soils on Earth may be constructed 'from the bottom up,' rather than 'from the top down.' At that time, studies of paleosols can be expected not only to illuminate specific events in the long history of life on land, but to contribute to general theories of soil formation and organic evolution.

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