How to Find a Precambrian Paleosol

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Abstract

Precambrian paleosols have been discovered in unprecedented numbers in recent years and it is likely that many remain unrecognized for what they are. This essay attempts to aid the search by outlining diagnostic versus common but not necessarily distinctive features of Precambrian paleosols. Three general classes of criteria can be used to recognize Precambrian paleosols: biological traces, soil horizons, and soil structures.

Precambrian paleosols do not contain the root traces that are so helpful in identifying Phanerozoic paleosols. Moreover, the microbial traces in paleosols, such as ministromatolites, rock varnish, and microbial borings, are not yet distinguishable from similar phenomena on the floors of shallow lakes or marine shelves. The most diagnostic microbial indicator of Precambrian paleosols yet reported is the upward lightening of isotopic ratios of ^{18}O vs. ^{16}O and ^{13}C vs. ^{12}C in weathered carbonate rocks.

Soil horizons of Precambrian paleosols, like those of the Phanerozoic, ideally exhibit gradational changes in color, texture, mineralogy, and chemical composition downward from a sharp upper contact. Subsurface horizons enriched in quartz, clay, sesquioxides, or humus have not yet been demonstrated in paleosols older than the Devonian advent of trees. There are, however, Precambrian examples of surface horizons enriched in clay and of subsurface horizons enriched in carbonate minerals, gypsum, amorphous silica, or manganese. There also are Precambrian examples of bauxitic, kaolinitic, and iron-rich deep weathering.

Some soil structures are diagnostic in that they demonstrate multiple episodes of breaking, infilling, and closure due to randomly oriented, small-scale extensional and compressional forces at atmospheric temperatures and pressures. Examples include subangular near-equant soil clods (blocky peds), veins of clay washed down soil cracks (clay skins or illuviation argillans), tepee or pseudoanticline (mukkara) structures of swelling-clay soils, sand wedges and other features of periglacial soils, and certain kinds of microfabric in which birefringent clays are arranged as random streaks in a less-oriented matrix (insepic, mosepic, and masepic plasmic fabric).

Most Precambrian paleosols are deformed, metamorphosed, and poorly exposed, so can be confused with upward-fining fluvial sequences, mudflow deposits, ash beds, marine hardgrounds, fault mylonites, or zones altered by either groundwater or hydrothermal fluids. Argillic hydrothermal alteration is particularly troublesome because its effects cannot yet be distinguished from the wide range of known soil-forming conditions on compositional criteria alone. To distinguish a paleosol from hydrothermally altered rock, evidence is best sought from horizonation, soil structure, and geologic context.

1 Introduction

Are Precambrian paleosols rare? Compared to Precambrian komatiites, placer uraninites, stromatolites, microfossils, or fluvial rocks, the number of known Precambrian paleosols remains meager (Sokolov and Heiskanen 1984; Pinto and Holland 1988). However, Precambrian paleosols have been discovered in unprecedented numbers in recent years and it is likely that many remain unrecognized for what they are. Few geologists have extensive knowledge of soil science, and few soil scientists have carefully studied Precambrian rocks. It is only in the past two decades that the expertise of soil science has been diligently applied to Phanerozoic fluvial rocks. As a result, literally thousands of paleosols have been discovered (Bown and Kraus 1981; Retallack 1983; Allen 1986).

Paleosol studies have not been as successful in the Precambrian as in the Phanerozoic for at least two reasons. Precambrian paleosols lack obvious and diagnostic biologic features, such as root traces (Retallack 1988 a) and many Precambrian paleosols have

M. Schidlowski et al. (Eds.)
Early Organic Evolution: Implications for Mineral and Energy Resources

been so obscured by metamorphic alteration that they are difficult to distinguish from hydrothermally altered rocks (Lowe et al. 1985; Duchac and Hanor 1987). Despite these problems, considerable progress has been made recently on where and how to look for Precambrian paleosols. This essay outlines diagnostic versus common but not necessarily distinctive features of Precambrian paleosols. In looking for Precambrian paleosols, as in all science, fortune favors the prepared mind.

Given that it is premature to judge how common Precambrian paleosols may be, it is difficult to predict their significance for reconstructing Precambrian life and landscapes. Precambrian, like Phanerozoic, paleosols potentially may provide evidence of past climates, organisms, topographic relief, parent materials, and duration of soil formation (Retallack 1986b). The history of Precambrian atmospheric oxygenation (Holland 1984; Pinto and Holland 1988) and the tectonic style of continental accretion and differentiation (Retallack 1990) are two perennial problems of Precambrian geology that are being profitably addressed through study of paleosols. A comparison of Precambrian paleosols with surficial alteration of equivalent or greater age on the Moon, Mars, Venus, and other planetary bodies may elucidate the early history of our Solar System (Retallack 1988b, 1990). Paleosols also provide evidence for the origin and early evolution of life on land (Retallack 1986a, 1990).

The discovery of additional Precambrian paleosols would enhance these studies. Three aspects of Precambrian paleopedology are emphasized herein to facilitate such discoveries: where to look for Precambrian paleosols, what features are diagnostic of them, and how they are distinct from other geologic phenomena.

2 Where to Look for Precambrian Paleosols

Perhaps the most direct way to find Precambrian paleosols is to advertise among geologists. Holland (1988) has tried this, with some success. A less direct, but perhaps more promising, approach involves "reading between the lines" of Precambrian studies. As in Phanerozoic sequences, Precambrian paleosols may be expected along major unconformities and within nonmarine sedimentary or volcanic strata.

Paleosol reconnaissance should start with a geological map. Unconformities overlain by fluvial rocks are especially promising candidates for paleosols (Gay and Grandstaff 1980; Farrow and Mossman 1988). Some rivers erode canyons deep into fresh bedrock but others bury weathered hillsides or plains under alluvium (Fig. 1). Colluvium also may bury soil. Irregularity of a mapped geological contact occasionally records paleotopography rather than tectonism (Williams 1969). Highly irregular contacts characterize some carbonate sequences and some of these irregularities have proven to be ancient karst (Button and Tyler 1981; Schau and Henderson 1983; James and Choquette 1987).

Fig. 1. Sheigra clay paleosol beneath Torridonian (1 Ga) alluvial fan deposits in sea cliffs near Sheigra, NW Scotland. (Retallack 1986a)

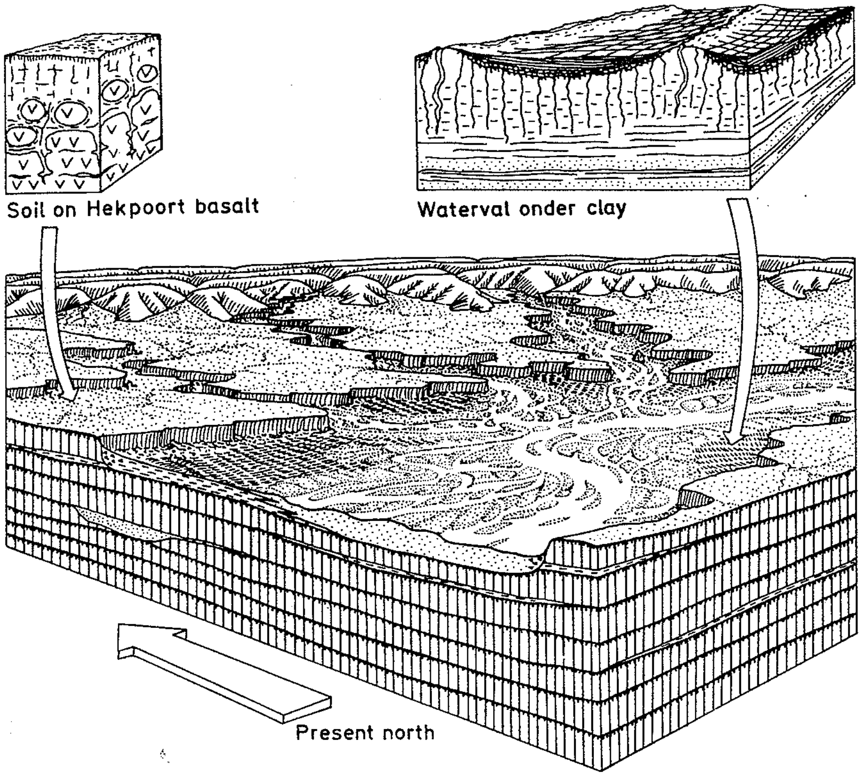

Fig. 2. Reconstructed sedimentary environment of the Waterval Onder clay (in basal Dwaal Heuvel Formation) and of undescribed paleosol on the 2.2 Ga Hekpoort Basalt in eastern Transvaal, South Africa. (Retallack 1986b)

Reconnaissance should also extend to mining records and publications. Some Precambrian aluminum, iron, manganese, and uranium ores formed as or near paleosols. This is not to say that they all are necessarily paleosols but rather that they deserve reexamination from this perspective. Crustal igneous rocks average about 15.6% Al_2O_3 but enrichments greater than 45% are found in pedogenic bauxite ores. Gibbsite, boehmite, and diaspore are characteristic of these paleosols. These three minerals become metamorphosed to kyanite, sillimanite, or corundum (Reimer 1986). Comparable degrees of iron enrichment are found in some Oxisols and paleosols (McFarlane 1976). Hematite in these soils and paleosols is remarkably stable up to greenschist-facies metamorphism. It alters to magnetite within the amphibolite and granulite facies (Thompson 1972). Manganese ores have been found in some paleokarst (Button and Tyler 1981; Schau and Henderson 1983) and some uranium ores are associated with major unconformities (Meyer 1985). Beware that Precambrian ores of sedimentary, volcanogenic or hydrothermal origin fortuitously also occur at major unconformities. In addition, the long periods of geological time represented at major uncormities means that they may preserve confusingly overlapping evidence of successive weathering episodes. Nevertheless, mining records for unconformity-associated ore deposits may well lead to the discovery of a Precambrian paleosol.

Studies of Precambrian sedimentary environments (Fig. 2) are also potential souces of paleosol information. Paleosols are abundant in Phanerozoic floodplain, eolian, alluvial-fan, and deltaic facies (Retallack 1986a). However, paleosols are difficult to distinguish in nonmarine Precambrian sequences because these rocks lack the fossil root traces which distinguish Phanerozoic paleosols. Useful criteria in Precambrian sequences mostly reflect climatic extremes. Ice wedges, sand wedges, pingos, and other periglacial features characterize soils which are classified as Cryorthents and Cryopsamments in the U.S. (Soil Survey Staff 1975). There is a long Precambrian fossil record of these soils of frigid climates (Williams 1986). The formation of salt crusts in deserts, playas, and coastal sabkhas is also a soil-forming process, resulting in soils identified as Salorthids in the U.S. taxonomy. Precambrian paleosols of this kind are among the most ancient known (Lowe 1983). Micritization, boring, and nodule formation in limestone and coastal beachrock also are soil-forming processes. The resulting soils are classified as Orthents and Calciorthids. Precambrian examples occur in shallow-marine and intertidal carbonates (Bertrand-Sarfati and Moussine-Pouchkine 1983; Grotzinger 1986). These distinctive alterations in place are the most obvious of a variety of subtle weathering effects which are detectable in Precambrian sedimentary rocks.

Studies of Precambrian volcanic rocks offer yet another potential source of information about Precambrian paleosols. Sequences of pillow lavas are not a promising place to look for paleosols because pillows form subaqueously. However, terrestrial sequences of lava flows may include paleosols, as well documented in the Irish Tertiary (Eyles 1952). Lahars, ignimbrites, and ash beds also are separated by paleosols in Phanerozoic volcaniclastic sequences (Retallack 1983) and a few Precambrian examples are known (Blades and Bickford 1976). Surficial weathering is not the only cause of stratiform alteration within volcanic sequences. Alternative methods of alteration include groundwater moving through porous pyroclastics after burial, hydrothermal alteration driven by volcanic heat, and downward baking by directly overlying lava (Fisher and Schminke 1984). None of these alternatives generally explains thick clayey horizons between Phanerozoic flows because the clays are less permeable than the columnar and tabular joints, because the mixing of hot lava and cool water is more likely to produce breccia, scoria, or lava spines than extensive interflow alteration, and because anhydrous volcanic rocks are such good insulators that baking generally is limited to just a few centimeters. Nevertheless, the foregoing nonpedogenic alternatives should be considered when evaluating possible paleosols in Precambrian volcanic sequences (Smith et al. 1982; Lowe et al. 1985; Duchac and Hanor 1987).

3 How to Look for Precambrian Paleosols

Three features of Phanerozoic paleosols are most useful for distinguishing them from enclosing rocks: root traces, soil horizons, and soil structure. Root traces are not found in Precambrian paleosols because they predate the advent of multicellular land plants. Nevertheless, other structures and effects of nonmarine microbes can be found. Similarly, the soil horizons and structures of Precambrian paleosols may not be as obvious or diverse as those in Phanerozoic paleosols, but comparable features have been found.

3.1 Microbial Traces

Even among soil scientists there is some disagreement about the exact definition of soil, but almost all definitions include material which supports the growth of vascular plants. The most diagnostic features of Phanerozoic paleosols are fossil roots or their clay or crystal-lined holes. The fossil record of roots is securely known only back to the Devonian (Retallack 1985, 1986a). Silurian paleosols contain some possible rhizome traces (Schopf et al. 1966; Retallack 1985). Paleosols as old as Late Ordovician contain burrows of animals so large that they probably fed on multicellular plants. They also contain erosional scours around reduction spots that may represent clumps of plants (Dewey, in Boucot et al. 1974; Retallack and Feakes 1987). The Late Ordovician is the time of the earliest fossil spores of land plants (Gray 1985). Prior to Late Ordovician, the land may have been covered with unicellular algae, bacteria, and other microbes (Wright 1985).

No indisputable microfossils have yet been found in a Precambrian paleosol despite several tantalizing possibilities (Retallack 1986a). Encouragement for the search comes from the discovery of traces of supratidal microbes in Cambrian phoscretes (Southgate 1986). These microbial traces can be distinctive but are very similar in soils and shallow marine or lacustrine deposits (Siegel 1977; Golubic and Campbell 1979). For example, some Precambrian paleosols contain structures like microbially produced rock varnish (Fig. 3). Modern iron-manganese varnish occurs on the upper but not lower sides of exposed pebbles within soils, streams, and lakes (Dorn and Oberlander 1982).

Modern carbonate crusts on cliffs and in caves (Reams 1989) and opaque rock varnishes on exposed rocks (Perry and Adams 1978) are microbially laminated and mounded (Fig. 4). These resemble the widespread Precambrian trace fossils called ministromatolites (Hofmann and Jackson 1987). Ministromatolites are not diagnostic of paleosols but neither should they be assumed to be exclusively aquatic. Possible Precambrian pedogenic ministromatolites include the laminated andalusite caps on quartz grains illustrated by Grandstaff et al. (1986).

Like ministromatolites, traces of endolithic microbes are similar in subaerial and aquatic environments. The bulbous or vase-shaped hollows created by endolithic algae in carbonate rocks occur in both marine (Knoll et al. 1986) and nonmarine settings (Folk et al. 1973). Subaerial endolithic algae produce spongy dark limestone which is known as phytokarst. Possible fossil examples are as old as Devonian (Folk and McBride 1976). Endolithic algae also are thought responsible for networks of surficial grooves on both silicate and carbonate grains in Phanerozoic marine and nonmarine sandstones (Ross and Fisher 1986). Endolithic microbes presently produce thin leached or ferruginized zones just below exposed rock surfaces

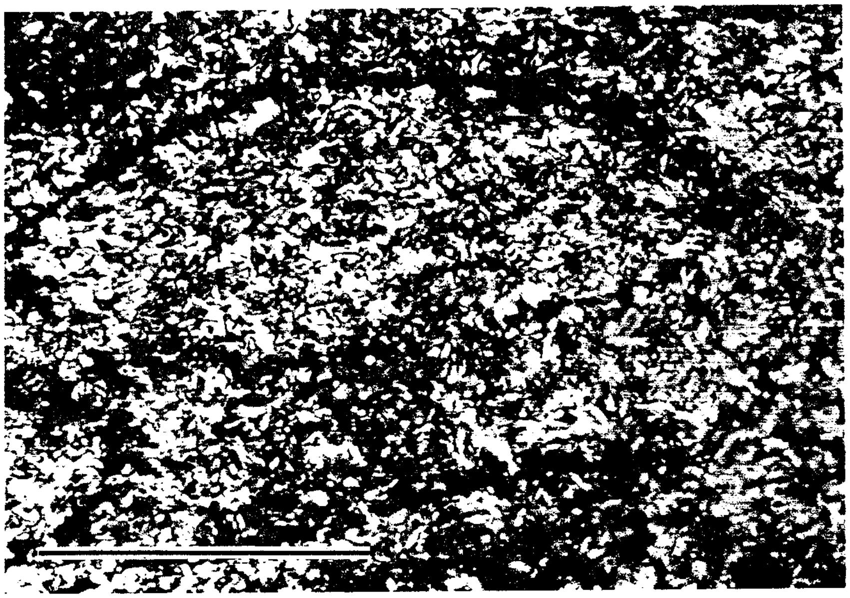

Fig. 3. Possible recrystallized rock varnish on clay granule in a petrographic thin section of surface (*A*) horizon of Waterval Onder clay paleosol of Fig. 2. *Bar* 0.1 mm. (Retallack 1986b)

Fig. 4. Ministromatolitic rock varnish in thin section from exposure near Phoenix, Arizona. Field is 0.15 mm wide. (Perry and Adams 1978)

(Friedman and Weed 1987) and similar submarine processes may be responsible for the discoloration of hardgrounds and boulders (Bathurst 1975; Bromley and Hanken 1981). The search for these kinds of microbial traces in Precambrian paleosols has just begun.

An isotopic signature may be diagnostic for pedogenic microbial colonization of subaerially exposed rock and sediment. An upward increase in lighter stable isotopes of both oxygen ($^{18}O/^{16}O$) and carbon ($^{13}C/^{12}C$) characterizes micritized limestone bedrock beneath modern Vertisols in Bermuda (Vernon and Carroll 1965; Allan and Matthews 1982) and paleosols which have developed on carbonate rocks ranging in age back to 1.2 Ga (Beeunas and Knauth 1985). Enhancement of the light isotopes is attributed to photosynthetic organisms but such isotopic fractionation may be compromised by burial alteration and metamorphism (Vahrenkamp et al. 1987). Precambrian occurrences are best evaluated by concurrent study of such indicators of alteration as illite crystallinity, kerogen maturation, and metamorphic petrography.

Paleosols may exhibit a surficial enrichment in either organic carbon or the trace elements which normally become complexed with organic matter, such as P, Cu, Ni and Y (Retallack 1986b). However, these elements also are concentrated in phoscrete and marine hardgrounds. Although minor-element distributions and microbial traces are not diagnostic of paleosols, they are important evidence of life in a paleosol which has been recognized by other criteria.

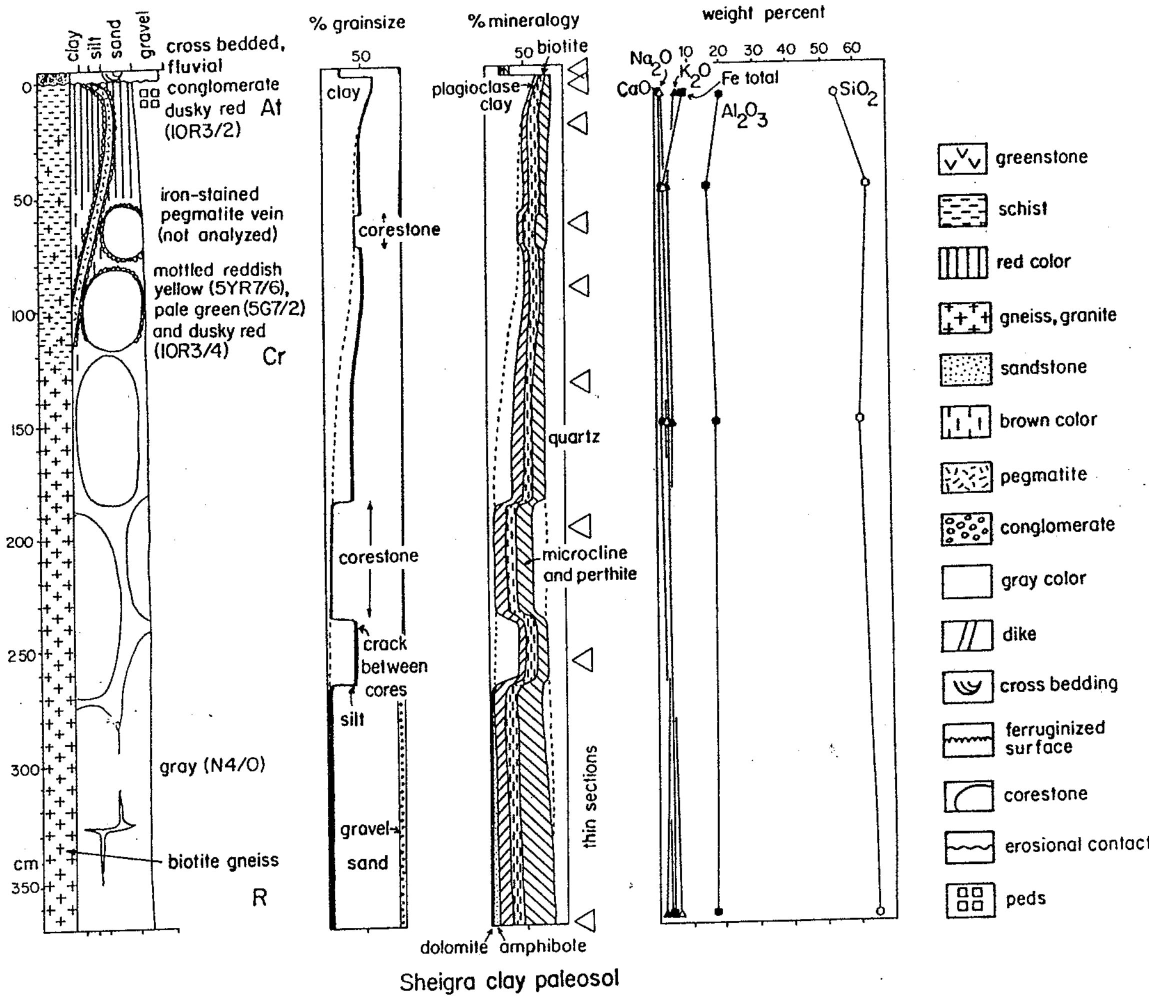

Fig. 5. Field section (by G. J. Retallack), grain-size distribution and modal analysis (by G. S. Smith) and chemical analysis (by G. E. Williams) of the Sheigra clay paleosol of Fig. 1. (Retallack 1990)

3.2 Soil Horizons

Soil formation essentially is alteration downward from a land surface. Ideally, it results in gradational changes downward through one or more distinct kinds of soil toward a little-altered parent material. Paleosols generally exhibit a sharp contact with overlying strata due to erosion of the uppermost soil. Boundaries between horizons within soils and paleosols are gradational over centimeters to meters (Figs. 1, 5). Vertical profiles of color, clay content, or elemental abundance through a typical paleosol exhibit only minor irregularities around nodules or less-weathered corestones. In contrast, sedimentary beds, hydrothermal vein networks, and igneous intrusions typically exhibit either sharp contacts or erratically zig-zagging compositional profiles.

One exception to the norm of a sharp upper contact and gradational boundaries is a floodplain soil subjected to increased frequency of flooding. For interpreting paleoenvironments, it is important to differentiate the paleosol from overlying transported soil material. For example, the NAN profile of the Elliot Lake paleosol (Mossman and Farrow, this Vol.) appears to be capped by transported soil (above 2.3 m).

Other exceptions are lithological differences in the parent material, which become inherited as sharp boundaries, color contrasts, compositional anomalies, or stone lines. One example is the Waterval Onder paleosol of the basal Dwaal Heuvel Formation in the Transvaal, South Africa. It was initially interpreted as a single thick paleosol developed on basalt (Button and Tyler 1981). However, the mineralogical and

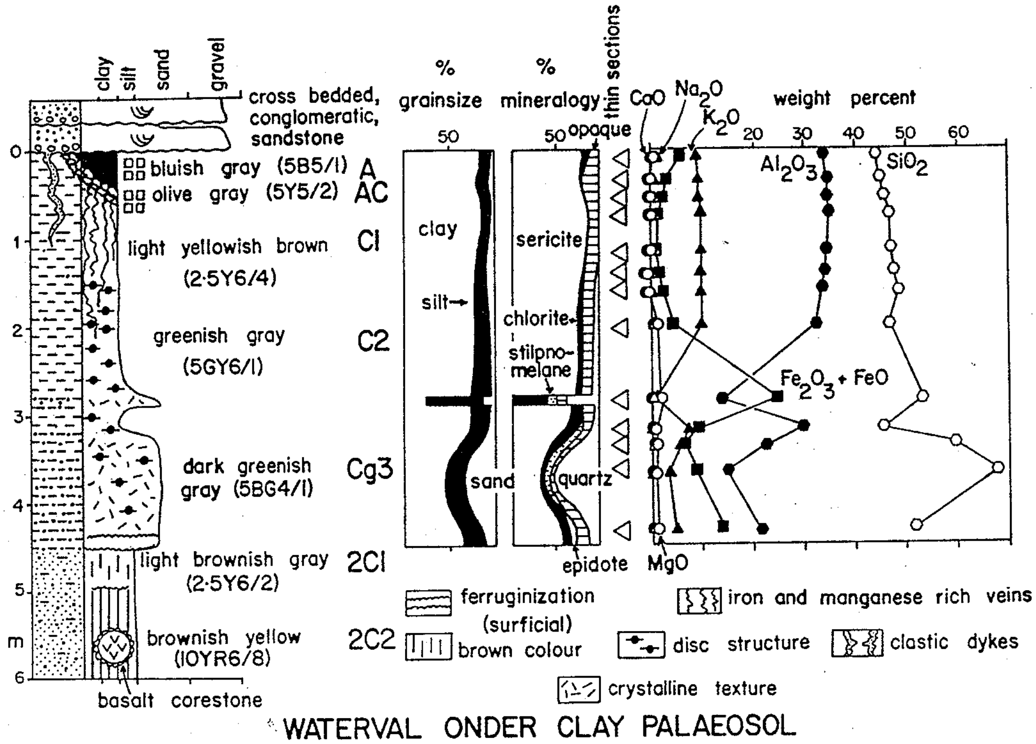

Fig. 6. Field section and grain-size distribution (by G. J. Retallack), petrographic composition and chemical analysis (by A. Button) of the Waterval Onder paleosol of Fig. 2. (Retallack 1986b)

chemical profile is erratic below 2.5 m (Fig. 6). Examination by the author has revealed relict beds and sedimentary structures of a fining- and thinning-upwards sequence of metamorphosed sandstone and shale overlying the basalt (Retallack 1986b). In contrast, the upper portion of this profile (above 2.5 m) shows the smooth chemical and petrographic variation expected in a paleosol.

Paleosol horizons can be specified using widely accepted abbreviations and concepts for soil horizons (Guthrie and Witty 1982). These include surface horizons of peat (organic or O) and near-surface horizons enriched in quartz (eluvial or E). Subsurface horizons may be enriched in clay (argillic or Bt), in humus and sesquioxides of iron and aluminum (spodic or Bs), or in carbonate (calcic or Bk).

Clayey soils of peaty swamps and marshes have become the rooted underclays of Phanerozoic coal seams (Gardner et al. 1988). Precambrian organic soils could potentially yield comparable evidence of ministromatolites, mudcracks, clay-flake breccias, and gradations of microbial mats into underlying shale.

Phanerozoic ganister paleosols with near-surface quartzose (E) horizons are found in Euramerican coal measures (Percival 1986). Similar effects are difficult to detect in cherty Precambrian paleosols because abundant silica has been supplied by the illitization and sericitization of clay minerals during late diagenesis (Curtis 1985) and hydrothermal alteration (Rinehart 1980). In thin section, near-surface quartzose (E) horizons of paleosols may exhibit evidence of hydrolytic weathering in a loose medium, e.g., abundant quartz and K-feldspar grains with subordinate etched mafic grains, all surrounded by thin clay skins (Dumanski and St. Arnaud 1966).

Pedogenic clay in subsurface clayey (Bt) horizons is characteristically laminated along former cracks in a paleosol (Brewer 1976). Pedogenic iron coatings around mineral grains in subsurface spodic (Bs) horizons exhibit internal lamination which is interrupted by deep radial cracks that extend inward from the surface of the coating (de Coninck et al. 1974). These petrographic criteria may prove useful in recognizing organic, eluvial, argillic, and spodic horizons in Precambrian paleosols, but they have not yet been found in paleosols older than Devonian (Retallack 1986a). Perhaps they do not occur in Precambrian paleosols because their formation is promoted by large plant roots which influence soil formation more than is possible under microbial ecosystems.

A common kind of surficial horizon in Precambrian paleosols is one enriched in clay (horizon At in Fig. 5). In some Early Precambrian paleosols, this horizon is a distinctive lime green color and is chemically reduced (high Fe^{2+}/Fe^{3+}), with a thick subsurface horizon (saprolite or C horizon) which exhibits the crystalline, metamorphic, or sedimentary structures of its parent material. These reduced but formerly well-drained paleosols are of great interest as evidence for a very weakly oxidizing primeval atmosphere (Holland 1984).

Some subsurface horizons in Precambrian paleosols are enriched in carbonate minerals (calcic or Bk). The carbonate may resemble shallow-marine carbonates in being nodular or pisolitic. A characteristic feature of pedogenic carbonate nodules is displacive fabric, e.g., spar-filled cracks or wedge-shaped cavities which surround the nodules or clasts due to their movement in loose soil. Replacive fabric is common both in soil and marine nodules. In this fabric, embayed margins of grains are filled by and float in a micritic cement. Almost all Precambrian carbonates are recrystallized to some extent, and this may obscure original fabrics (Kalliokoski 1986). Isotopic (Frank et al. 1982) and cathodoluminescence studies (Dickson and Coleman 1980) may be useful in unraveling their burial alteration history. Dolomite is a common carbonate in Precambrian paleosols (Retallack 1986a) whereas low-magnesium calcite is characteristic of calcareous Phanerozoic paleosols (Doner and Lynn 1977). It remains unclear whether this mineralogical difference is a product of burial alteration or records secular variation in the chemical conditions of soil formation. Similarly, the primary or diagenetic origin of marine dolostones remains disputed (Zenger et al. 1980; Given and Wilkinson 1987).

Precambrian paleosol horizons include near-surface enrichments in gypsum (gypsic or By: Lowe 1983), silica (duripan or Bq: Ross and Chiarenzelli 1985), and manganese (placic: Button and Tyler 1981). The classification of thick bauxitic, kaolinitic, and ferruginous horizons is controversial (Goudie 1973) but here they are interpreted to be deep weathering zones (Co, Cv and Ct, respectively). Bauxite, laterite, and china clay have all been reported in Precambrian rocks (Button and Tyler 1981; Bardossy 1982; Morris 1985).

3.3 Soil Structure

Soil structures progressively overwhelm the structures of the parent material, such as sedimentary bedding, metamorphic foliation, or igneous crystal outlines. These soil structures may prove critical for recognizing paleosols among their enclosing rocks. Nontechnical terms often used to describe soil structures in paleosols include "massive," "structureless," "jointy," "slickensided," "veined," "mottled," and "nodular". There is a much wider array of technical terms for soil structures (Brewer 1976) but this account will focus on peds, cutans, mukkara, periglacial patterned ground, and sepic plasmic fabric. These can be used to characterize paleosols even as ancient as Precambrian.

Peds are the natural aggregates (clods) of soil material that tumble loose when one digs in soil. Cutans (clod skins) are the naturally altered surfaces of peds, such as films of clay washed down cracks, or the rims of peds stained by iron oxides. Peds also are defined by cracks, pores, vugs, vesicles, and other open spaces, but these almost always collapse during burial of clayey paleosols. Former open spaces filled with later cements may be found in paleosols which developed on vesicular lava, volcanic ash, or limestone. In most cases, however, peds are recognized in paleosols by means of their bounding cutans. Peds are not necessarily the pieces of rock that come loose from a paleosol upon impact with a hammer. These fragments more often reflect jointing, foliation, and other nonpedogenic features.

The clearest known remnant of peds and cutans in a Precambrian paleosol occurs in the subsurface (AC in soil-science notation) horizon of the Waterval Onder clay paleosol, Transvaal, South Africa (Retallack 1986b). Here, blocky angular peds are defined by thick dark cutans of washed-in clay with opaque oxide grains of amorphous iron and manganese minerals (Fig. 7). On first inspection, one might be tempted to dismiss this structure as jointing, hydrothermal veining, secondary porosity, or tectonic breccia. However, the individual peds have healed interior faces of former cutans and boundaries that are either gradational or irregular with wispy flakes. The dark clayey cutans (clay skins or illuviation argillans) are internally banded, laterally continuous, and show local wedge-shaped cross-sections, as if the peds had been loose enough to rotate slightly. Furthermore, the cutan material is identical to the overlying surface (A) horizon from which it was washed downward. The ped material, on the other hand, is identical to the subsurface (C) horizon from which it was loosened. These observations indicate a complex history of multiple episodes of breaking, infilling, and closure under randomly oriented, extensional, and compressional stresses at moderate temperatures and negligible pressures. Such conditions are unlikely in any purely sedimentary, volcanic, igneous, or metamorphic environment. Other possible blocky peds in

Fig. 7. Angular blocky peds in polished slabs cut horizontally (*above*) and vertically (*below*) in subsurface (AC) horizon of Waterval Onder clay paleosol of Fig. 2. *Scales* graduated in cm. (Retallack 1986b)

Precambrian paleosols worthy of further examination are the clay-lined units of soil material illustrated by Gay and Grandstaff (1980) and the so-called "zebraic fracturing" of Lowe (1983), Lowe et al. (1985), and Duchac and Hanor (1987).

Even more diagnostic of paleosols than blocky peds are the pelletlike granular peds which characterize grassland soils (Retallack 1988a). However, other kinds of peds are not much more than mudcracks and may be indistinguishable from subaqueous synaeresis cracks or hydrothermal veins in deformed rocks. Among cutans, those formed by successive lining of cracks (illuviation cutans) are most diagnostic of paleosols (Fig. 7). Cutans which have formed by shear in soils (stress cutans) may show slickensides in a distinctively scattered orientation (Gray and Nickelsen 1989) but these are not markedly different from slickensides formed by by faulting, igneous injection, or compaction around coarse clasts. Cutans formed by chemical modification inward from a surface (diffusion cutans) also may not be appreciably different from features formed by development of secondary porosity during deep burial (Schmidt and McDonald 1979) or by hydrothermal alteration of porous rock (Edwards and Atkinson 1986). Putative peds and cutans therefore should be studied carefully and in context to determine their true nature.

Mukkara structure is a diagnostic feature of some paleosols (Paton 1974) and some modern swelling-clay soils (Vertisols of Soil Survey Staff 1975). Such features commonly are reported as "tepee structures" or "pseudo-anticlines" (Allen 1986). They may be differentiated from otherwise similar stratabound contorted laminations in that they have less conspicuous internal bedding and are more symmetrical and angular. The term mukkara is Australian aboriginal for "finger" and refers to the ridge of subsoil that is exposed between depressions floored by the disrupted surface horizon. This surface undulation is called gilgai microrelief. Gilgai is the aboriginal word for the small waterholes which abound on these undulating clayey soils after rainfall. Mukkara and gilgai are found in clayey soils, many of them rich in smectite, in subhumid to semi-arid climates with a strongly seasonal distribution of rainfall. The swelling of wet clay creates the large-scale structure as well as the radiating fans of intersecting shear planes that define slickensided lentil peds in some of these soils (Krishna and Perumal 1948). Numerous examples of these structures have been reported from Phanerozoic paleosols (Allen 1986) and the Waterval Onder clay is a well-known Precambrian example (Fig. 8).

Cryogenic features of paleosols include ice wedges, sand wedges, and patterned ground. Their record extends back to 2.3 Ga (Williams 1986). Sand wedges have vertically layered sandy fill and ice wedges are filled with horizontally bedded sediments. Both taper more strongly than cracks of mukkara structure. Periglacial patterned ground develops at or near the land surface during annual freeze-thaw cycles. Some of the patterns are restricted to specific climatic regimes (Washburn 1980).

Nodules and crystals are obvious features of many paleosols in outcrop but similar features can form during sedimentation, diagenesis, metamorphism, and cooling of lava. Like most nodules in sedimentary rocks, paleosol nodules are composed of minerals which form at low temperature and pressure. Diagnostic features of soil nodules and crystals include cement-filled voids which surround the nodules in a pattern which indicates that they could rotate slightly (displacive fabric). Paleosol nodules also exhibit good preservation of easily weathered minerals versus poor preservation (etching and pitting) of such minerals outside the nodules. Nodule-bearing paleosols may be overlain by fluvial paleochannels which contain similar nodules as pebbles. Paleosol nodules commonly congregate in horizons at a fixed depth below an

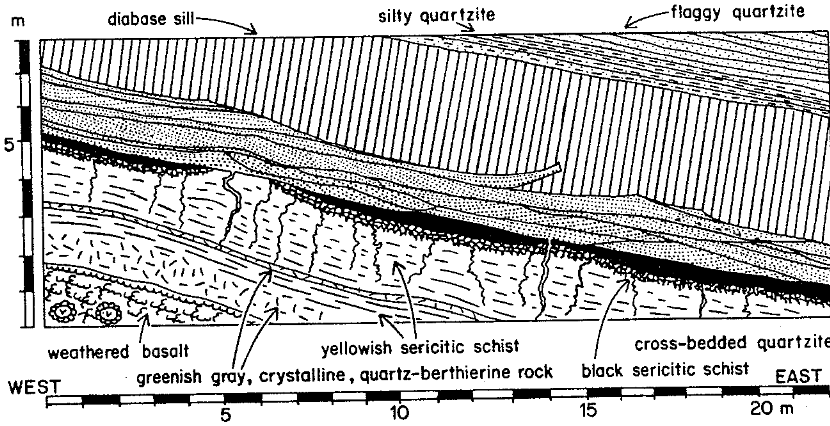

Fig. 8. Field sketch of mukkara structure in roadcut exposure of the Waterval Onder clay paleosol reconstructed in Fig. 2. (Retallack 1986b)

ancient erosional surface. In paleosols developed on sedimentary rocks, nodules may exhibit particularly well-preserved fossils, stromatolites, or sedimentary lamination inherited from the parent rock.

Among microscopic soil structures, the most clearly diagnostic for paleosols are certain kinds of bright-clay microfabric (sepic plasmic fabric). This is a pattern of randomly arranged streaks of highly birefringent clay in a less birefringent matrix (Brewer 1976). Such fabrics are observed in thin section at intermediate magnification (100 to 300 power) under cross-polarized light. The section should be cut perpendicular to the ancient land surface. Sepic plasmic fabric is a microscopic expression of peds and cutans and is therefore best expressed in strongly developed, well-drained, clayey soils and the corresponding paleosols (Brewer and Sleeman 1969).

Not all kinds of sepic fabric are diagnostic of paleosols. Grain-coating bright clay (skelsepic plasmic fabric of Brewer 1976) can be formed by the rolling of grains during sedimentation, by compaction of a clayey matrix during burial around hard mineral grains, and by grain-margin stress generated during shrinking and swelling of a soil. Sedimentary bright-clay rinds generally have smooth outer margins, and compactional clay rinds generally are thickest in the direction of principal strain. In contrast, bright clay around soil grains generally is wispy and randomly oriented. However, such distinctions are difficult to make in many cases. Cavity-lining bright clay (vosepic plasmic fabric) can be formed by infilling of secondary porosity generated by acidic groundwater during deep burial, as well as by water flow in soil cracks. Deep burial vosepic fabric may remain uncrushed or show highly discontinuous lamination unlike pedogenic examples, but these criteria also are difficult to use in practice.

In woven bright clay (omnisepic plasmic fabric), almost all the clay is bright and complexly interwoven. This can result from illitization and sericitization of clays during late diagenesis as well as through the accumulation of highly birefringent clay streaks over a long period of soil formation. Even so, when all the phyllosilicates are highly birefringent in metamorphosed paleosols, relative differences in brightness could represent original bright streaks of a less pervasively birefringent original fabric.

This leaves as diagnostic soil microfabrics incipient bright clay (insepic), streaky bright clay (mosepic), and intersecting bright clay (masepic). In these, the highly birefringent clay forms distinct wisps in a less birefringent matrix, indicative of a highly deviatoric, stratabound system of small-scale stresses that are difficult to envisage outside of soil environments. If the wisps within the highly birefringent matrix (omnisepic) of the thin section illustrated here (Fig. 9) can be considered to reflect an originally less pervasively sepic fabric, it would then be identified as clinobimasepic in the soil terminology of Brewer (1976).

Fig. 9. Cross-polarized thin section of the subsurface (C) horizon of the Waterval Onder clay paleosol of Fig. 2. Note woven bright clay (omnisepic plasmic) microfabric due to sericitization, with grayish streaks (third order purple birefringence) reminiscent of clinobimasepic plasmic fabric. Field is 0.8 mm wide

Paleosol microfabrics may include all the igneous, metamorphic, hydrothermal and sedimentary fabrics of their parent materials. These parent microfabrics are especially well preserved in paleosols that were young, waterlogged, or salty, and so less affected by biological activity or wetting and drying. Massive, opaque, isotropic or nearly isotropic (inundulic and undulic) fabrics formed by cementation or clay flocculation in soils also can form during sedimentation. Colloform cavity fills, cavity-lining crystal growths, grain overgrowths, grain etching or dissolution within clay skins, and complex vein networks formed during soil formation are not much different from those formed in the subsurface by groundwater, hydrothermal alteration, or igneous instrusion. Microfabrics produced during shallow burial are difficult to distinguish from those produced during soil formation even in paleosols as young as Quaternary. A soil origin should be considered whenever such fabrics are observed but a distinction between pedogenetic and diagenetic origins may require a careful evaluation of the paragenetic sequence of mineral growth, together with isotopic, cathodoluminescence, and geochemical-modeling studies.

4 What Else Could it Be?

Paleosols are not always easy to recognize and it is therefore useful to consider what else could resemble a paleosol. There are several possibilities. Some beds superficially similar to paleosols form by chemical alteration or biologic activity on the floors of oceans or lakes. During long periods between sedimentation events, burrowing by marine organisms or crystallization of salt may destroy primary bedding within sediment (Reineck and Singh 1975). Shallow tropical seafloors become cemented with aragonite and high-magnesium calcite and then become further modified (micritized) by boring algae. Deep marine hardgrounds may exhibit a complex history of dissolution and multiple encrustation of iron and manganese (Bathurst 1975). These marine and lacustrine hardgrounds and burrowed beds can be differentiated from paleosols by their sedimentary structures or fossil content.

Fluctuation of shorelines produces material gradational between sediment and soil. The modern intertidal zone does not have soil in the conventional sense except where vegetated by reeds, salt marsh, or mangroves. Fluctuation of a lacustrine shoreline in cycles of tens to thousands of years produces more complex situations (Freytet and Plaziat 1982). Fossil root traces are the best evidence for such Phanerozoic paleosols, and distinctive assemblages of stromatolites and microfossils may eventually prove useful guides to Precambrian examples (Knoll 1985a, b). Other useful indicators of paleosols in such a context include mudcracks, bright-clay (sepic plasmic) fabric, and clayey, sesquioxidic- or carbonate-rich soil horizons.

Massive or fining-upwards sedimentary beds in alluvial or colluvial deposits may present problems for paleosol recognition. For example, mudflows usually occur by slumping along a mud-lubricated slope. In a volcanic mudflow (lahar), the lubricating water may be hot and chemically aggressive, extensively altering both the muddy matrix and contained clasts as well as underlying materials (Fisher and Schmincke 1984). Mudflows, however, do not exhibit subsurface horizons of mineral enrichment, soil structures, or microbial trace fossils.

Zones of liquification during mass wasting may superficially resemble paleosols where there is contorted lamination and water-escape structures. However, paleosols tend to be more massive than contorted laminae and paleosol root traces or burrows tend to be more complex than water-escape structures. The latter typically branch upward at a narrow angle (Cloud and Lajoie 1980).

Upward-fining sequences deposited from waning flood flow (Reineck and Singh 1975) also have some of the general appearance of paleosols. They lack biological traces, subsurface horizons of mineral enrichment and etched minerals within the wisps of their clayey and sesquioxidic weathering products.

Faulting produces fine-grained and massive mylonite (Suppe 1985). However, mylonite generally is readily distinguishable from paleosol, even where developed along a bedding-parallel thrust fault because mylonite generally has uniformly directed penetrative deformation (Gray and Nickelsen 1989).

Igneous rocks are generally distinguishable from paleosol given their distinctive chemical composition, crystalline textures, typically sharp contacts with surrounding rocks, and cross-cutting relationships. However, sills and tuff beds may be deceptive if extensively altered by metamorphism and hydrothermal fluids (Fisher and Schmincke 1984). Sharp lower and upper boundaries, lack of biological traces, and the persistence of relict shards or crystalline textures in thin section commonly betray their origin.

Rock altered by the passage of groundwater during deep burial may resemble paleosols (Pavich and Obermeier 1985). This is particularly true along a major unconformity where a sharp contrast in permeability commonly occurs. Precambrian unconformities may have experienced a long history of

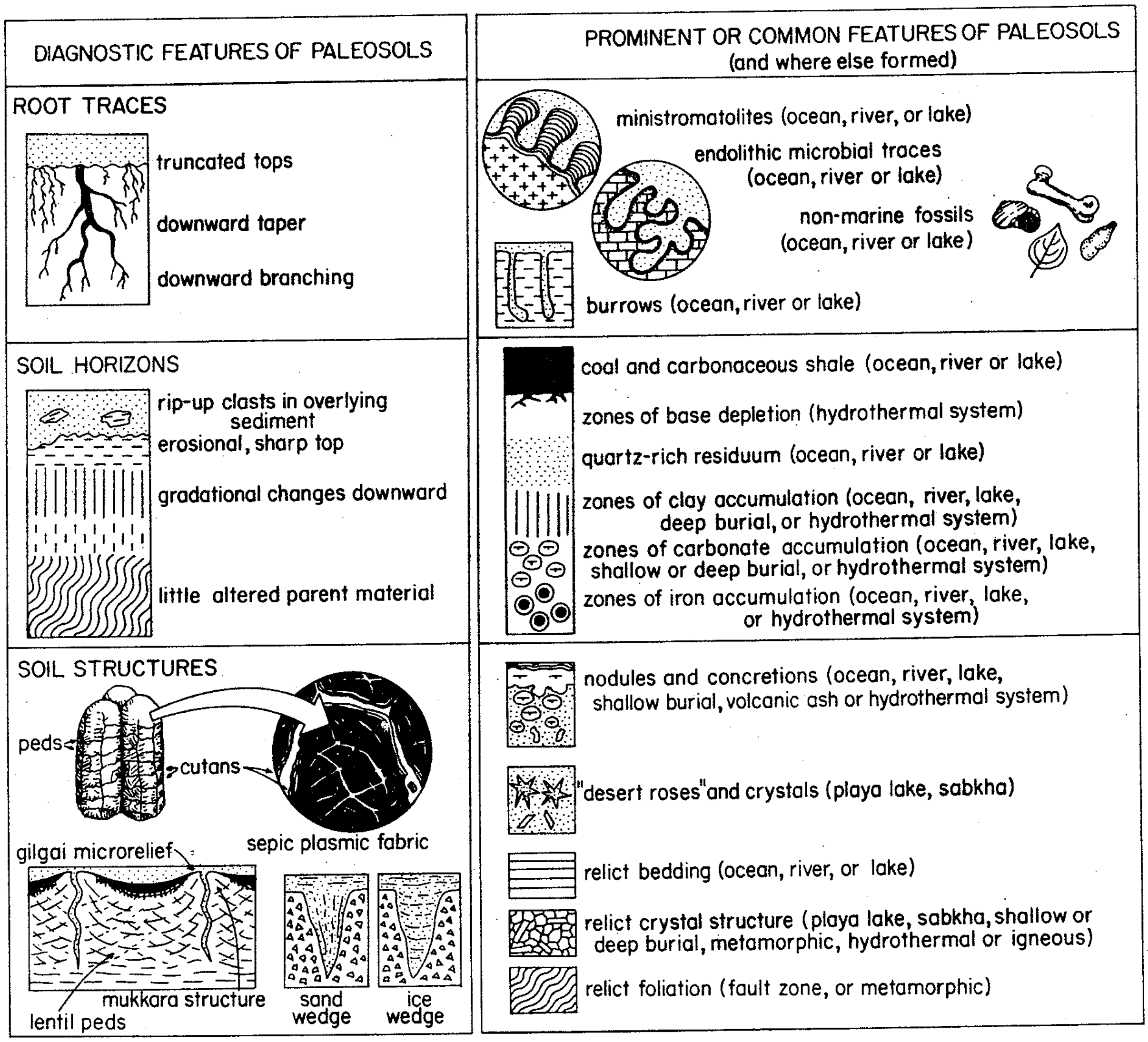

Fig. 10. Summary of diagnostic paleosol features and similar nonpedogenic features

repeated leaching and cementation. However, the resulting diagenetic structures generally are simpler than soil structures. Moreover, groundwater alteration rarely is oxidizing whereas weathering more commonly produces reddening from iron oxides.

Rock alteration by aggressive geothermal water is a special case of groundwater alteration (Rinehart 1980). Like weathering, this hydrothermal alteration may convert primary minerals to base-poor, alumina-rich clays (argillic alteration). In most cases, hydrothermal alteration is readily identifiable from the geologic context, such as pronounced lateral variation in alteration within a short distance, an association with volcanic rocks, and extension along a fault plane rather than along bedding.

Attempts have been made to distinguish paleosols from hydrothermal rocks on the basis of trace-element and alumina concentrations (Schreyer et al. 1981) and potassium enrichment (Duchac and Hanor 1987; Palmer et al. 1989). However, these criteria are not convincing, given the extreme alumina enrichment of bauxitic paleosols (Edwards and Atkinson 1986), the likelihood of even more aggressive weathering by Early Precambrian acid rain (Holland 1984), the alkali enrichment of many modern soils (Retallack 1986a), the diversity of trace-element abundances in soils (Kabata-Pendias and Pendias 1984), and the common late-diagenetic enrichment of potassium in paleosols (Curtis 1985). Chemical analyses commonly are insufficient to distinguish a paleosol from hydrothermally altered rocks.

Mineralogy is potentially more useful. Hydrothermal chlorite (propylitic alteration) and sericite (potassic alteration) are widespread whereas neither mineral typically results from weathering. The rupture of country rock by hydrothermal solutions may form a three-dimensional boxwork of veins that is superficially similar to soil structure. This may be distinguished from soil structure, however, because the minerals within the veins generally differ from typical soil minerals. Cracks in soil differ from hydrothermal boxwork by being filled more commonly with multiple generations of clay minerals. When exposed and weathered to a gossan, boxwork ores contain low-temperature minerals such as hydrous iron oxides but the primary ore minerals, such as sphalerite, form under conditions which are uncharacteristic of paleosols.

Zones of metamorphic alteration seldom are confused with paleosols because they contain distinctive minerals and cut across bedding (Kisch 1983). Superficial resemblance to paleosols may occur where metamorphic assemblages are juxtaposed due to metamorphism of contrasting protoliths or metamorphism along a fault zone. Schistosity and other metamorphic structures are quite dissimilar from biological traces and other soil structures. Furthermore, metamorphic alteration tends to be strongly reducing (Thompson 1972) whereas soil formation typically ranges from oxidizing to mildly reducing.

Many of the foregoing alternatives to interpreting a rock as a paleosol may seem so removed from soil formation as to be far-fetched. However, for small, tectonically complicated and metamorphosed outcrops, mistakes are on record in indentifying them as paleosols and in failing to identify paleosols that were present. Given that almost all Precambrian paleosols are metamorphically altered to some extent, the unraveling of metamorphic, hydrothermal, and soil components of alteration can prove challenging. In some cases, laboratory study is essential but the most tangible and direct evidence often is discovered by field observation. In searching for Precambrian paleosols, a clear concept of their diagnostic features is needed (Fig. 10). Paleopedology remains largely a field science, for which the most valuable instrument is a prepared mind.

Acknowledgments. This paper was written with fond memories of the splendid isolation, good company, and high science of the IGCP 157 meeting at Maria Laach, Germany, in 1988. I thank especially Manfred Schidlowski, who graciously facilitated my attendance. I also acknowledge more recent discussions with H. D. Holland and C. R. Feakes.

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