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UNTANGLING THE EFFECTS OF BURIAL ALTERATION AND ANCIENT SOIL FORMATION

G. J. Retallack

Department of Geological Sciences, University of Oregon, Eugene,
Oregon 97403

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Introduction

The interpretation of buried soils within rock sequences has become an increasingly popular field of study in recent years but is certainly not new. James Hutton (1795), in a treatise widely regarded as marking the birth of modern geology, recognized that red, rubbly rock, dividing Devonian sandstones from steeply dipping Silurian slates in southeastern Scotland, was a buried soil. For him, this paleosol represented a long period of time during which the slate was uplifted and eroded. Webster (1826) and Buckland (1837) realized that brown clayey beds among the latest Jurassic limestones of the Isle of Portland in southern England were also once soils. They called them "dirt beds" and offered flawed, but charming, experimental evidence for their origin as soils: A variety of plants could be grown in the brown clay! More convincing evidence of their origin came from three general observations: (*a*) They supported fossil stumps and root traces, (*b*) they had a sharp upper contact and gradational contact downward into their parent material, and (*c*) they had the complexly jointed appearance of soils, rather than the bedded appearance of sediments (Francis 1986).

An impressive array of fossil root traces, soil horizons, and soil structure is now known from paleosols (Retallack 1988). It is natural to emphasize soil features in the interpretation of past environments from paleosols. This approach has been so productive because it draws upon an enormous pool of data and theory from modern soil science in interpreting the fossil

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record of soils (Retallack 1990). Nevertheless, many buried soils are not very similar to surface soils, and this dissimilarity may be one of the reasons why paleosols have not been recognized in many sedimentary formations until the past few decades.

In some cases, the distinctive nature of paleosols is due to soil-forming conditions that no longer exist. Precambrian paleosols, for example, are commonly drab colored, unoxidized, and depleted in iron, even though they are deeply weathered and fissured, as if formerly well drained (Retallack 1986a,b, Farrow & Mossman 1988). These ancient soils probably formed under conditions of much lower amounts of oxygen in the atmosphere than now present (Holland 1984).

In other cases, features of paleosols not found in surface soils are more likely due to alteration of paleosols after burial. The variety and scope of alteration possible during burial and metamorphism are so great that they may seem at times like a counsel of despair for the interpretation of paleosols. In attempting to compare analogous surface soils with hundreds of paleosols, I have found that burial alteration may be profound but tends to fall into a limited number of categories. Common and widespread alterations after burial, which can be expected for most paleosols older than Pleistocene, are the main subject of this review.

The effects of extreme metamorphic alteration, perhaps surprisingly, are often clear from the development of such structures as schistosity and high-temperature minerals such as corundum and sillimanite. More troublesome in the interpretation of paleosols are modifications during burial that fall short of metamorphism—that is, diagenetic changes. Especially problematic are changes that occur soon after burial, when the paleosol is still within the reach of surficial processes. These modifications in aqueous solutions, with the aid of microbes and under low temperatures and pressures, are not always distinct from soil formation. Indeed, if diagenesis is defined as alteration of sediments after deposition, then diagenesis includes both soil formation on sediments and their alteration after burial.

Burial Decomposition of Organic Matter

Studies of Quaternary paleosols and equivalent surface soils in central North America have shown that soon after burial, paleosols lose up to an order of magnitude of organic carbon as determined by the Walkley-Black technique, but that they may preserve the general trend of organic matter abundance with depth (Stevenson 1969). This loss of organic matter was found only in paleosols similar to well-drained surface soils, not in paleosols that were peat rich and waterlogged (Figure 1). The lost organic matter may have been metabolized by aerobic microbial decomposers that were

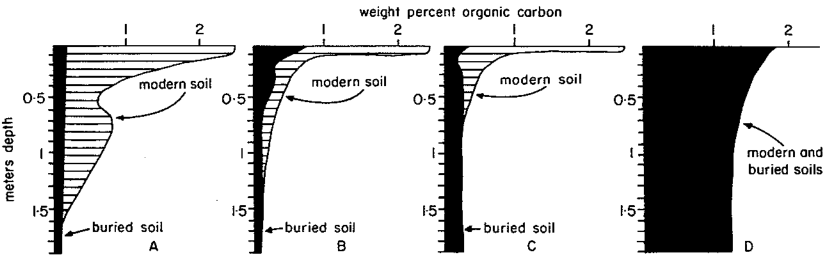

Figure 1 Comparison of organic carbon content, as determined by Walkley-Black titration, in some paleosols and comparable surface soils of the midwestern US: *A*, Yarmouthian (10^6 yr old) paleosol and surface (Holocene) Albaqualf; *B*, Yarmouthian and surface Hapludalf; *C*, Farmdalian (30,000 yr old) paleosol and surface Hapludalf; *D*, Wisconsinan (14,000 yr old) paleosol and surface Udifluvent [modified from Stevenson (1969), with permission of Williams-Wilkins Co.].

part of the ecosystem forming soil on the sediment that buried the paleosol. Additional loss of organic matter may occur during deep burial and the generation of oil and gas (Surdam & Crossey 1987).

Many paleosols contain much less organic carbon than one would expect considering their fine soil structure and abundant root traces and burrows, which indicate former biological productivity. For example, Miocene paleosols from Kenya and Pakistan (Retallack et al 1990, Retallack 1991) have very low analytical values of Walkley-Black organic matter (no more than 0.65 wt%). Generally comparable surface soils of the Serengeti Plain of Tanzania have organic carbon contents of up to 9% (de Wit 1978, Jager 1982). Similar surface soils of the Songhor area of Kenya have organic carbon values of up to 5.64% (Thorp et al 1960), and on the Indo-Gangetic Plains of India and Pakistan organic carbon is up to 2.68% (Murthy et al 1982).

The loss of organic matter in paleosols may change their color so that they now appear less dirty (higher Munsell chroma). Such organic matter loss also would make the paleosols more prone to badlands weathering in the modern outcrop compared with original soils with clays stabilized by roots and organic coatings around soil clods.

Burial Gleization Around Organic Matter

A distinctive feature of many paleosols is a blue-gray or green-gray color of horizons, of mottles, or of haloes around root traces (Figure 2). These gley colors provide a striking contrast to red parts of the profile.

In some paleosols, the light-colored surface horizon and root traces

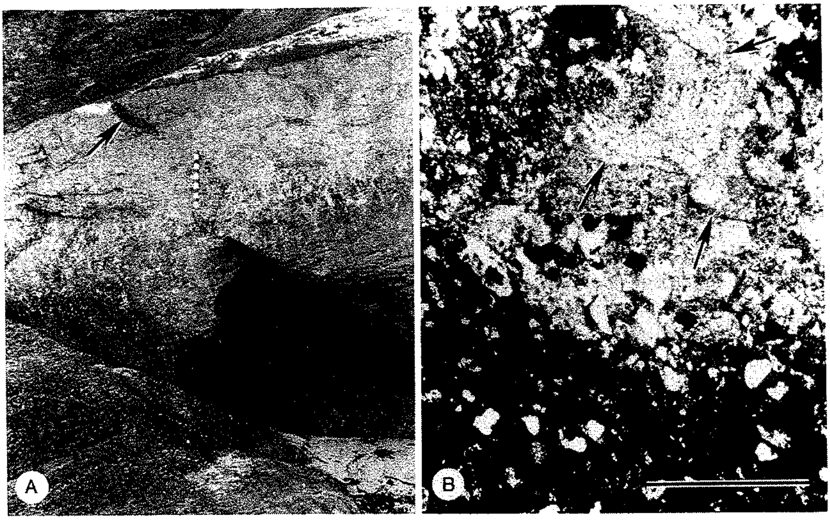

Figure 2 (*A*) Paleosol with a sharp top and dark burrow fill (at *arrow*) and a gray-green surface horizon with numerous drab-haloed root traces penetrating a dark red subsurface horizon. (*B*) Petrographic thin section under plane light and cut horizontal to the former land surface of a drab-haloed root trace, showing concentric clayey fill of hole of large root (*inside arrows*) and a small rootlet (*lowermost arrow*). Both samples are from the Long Reef clay paleosol, a Hapludult of Early Triassic age in the Bald Hill Claystone, near Sydney, Australia. Field scale is in inches; bar scale for micrograph is 1 mm. This paleosol is discussed by Retallack (1976).

extending down into a reddish subsurface horizon are superficially like the arrangement in some forested soils (Alfisols, Spodosols, and Ultisols), in which a sandy near-surface horizon (eluvial or albic horizon) has been leached of brown, oxidized clay into a subsurface horizon (illuvial or argillic horizon) that is clayey or brown. Root channels and burrows form natural conduits for such leaching and for the penetration of the light-colored sandy material deep within the profile, in the form of distinctive structures called glossic features (Soil Survey Staff 1975) or krotovinas (Buol et al 1980). A few paleosols of this kind are known (Retallack 1976).

Bluish and greenish colors are found in the clayey subsoils of peaty waterlogged soils (Ho & Coleman 1969, Landuyt 1990) and are a conspicuous feature of groundwater gley. These drab colors also occur in clayey soils on which stagnant water ponds for a good part of the year, as features of surface-water gley (Retallack 1990). In both cases, anaerobic microbes in poorly oxygenated water have chemically reduced brown and red iron oxides and hydroxides from the ferric state to a drab-colored ferrous state. Because ferrous iron is much more soluble than ferric iron, such gleyed layers or spots in soils commonly are strongly depleted in iron compared with other parts of the soil. Gleyed soils commonly also show other evidence of iron remobilization, such as nodules or layers of vivianite, pyrite, siderite, and amorphous iron-manganese. Many gleyed soils have peat-rich surface horizons and well-preserved tabular root systems and leaf litters. Such drab paleosols rich in fossil plants are widely known, especially in coal measures (Retallack & Dilcher 1981, Retallack et al 1987, Gardner et al 1988).

Drab horizons and mottles also can form by burial gleization—the chemical reduction of iron hydroxides and oxides by anaerobic bacteria consuming organic matter buried with the soil below or near the water table. Reduction haloes around buried organic matter can form in only a few thousand years (Allen 1986). Burial gley is an especially appealing explanation for drab-colored surface horizons and drab-haloed, deeply penetrating root traces in clayey paleosols with red oxidized subsurface horizons that lack textural differentiation of horizons or evidence of former waterlogging (Figure 2). There are many such paleosols (Retallack 1976, 1977, 1983, 1985, 1991, Retallack & Dilcher 1981, Bown & Kraus 1981, 1987, Blodgett 1988). Preliminary geochemical studies reveal that in most cases total iron is depleted in drab parts compared with red parts of the profile, but that substantial amounts of chemically reduced iron remain in drab areas (Retallack 1976, 1983, Braunagel & Stanley 1977), as might be expected during alteration in a slowly permeable system soon after burial. It is likely that a variety of anaerobic microbes common in waterlogged soils play a role in early burial gleization. The inorganic chemical reduction of iron oxides is sluggish under burial conditions milder than amphibolite-grade metamorphism (Thompson 1972), as shown by red- and purple-colored and red-green mottled paleosols of the greenschist facies of regional metamorphism (Retallack 1985, Feakes & Retallack 1988). The original color of parts of paleosols chemically reduced by burial gley was probably brown, orange, or yellow, darkened with organic matter, as is usual for the surface horizons of well-drained soils.

Some paleosols lack burial gley features. Among Miocene paleosols of Kenya, those formed in alluvial lowlands around the base of carbonatite-nephelinite cinder cones and stratovolcanoes (for example, at Songhor) have abundant evidence of burial gleization, but those formed on nearby volcanic footslopes (for example, at Koru and Fort Ternan) do not show burial gleization of either root traces or horizons (Retallack 1990, 1991, Retallack et al 1990). Such observations support the idea that subsidence below the water table is necessary for active microbial reduction of buried organic matter. In contrast, paleosols buried and remaining above the water table may have lost organic matter by aerobic decomposition.

Burial Reddening of Iron Oxides and Hydroxides

Many paleosols are a brick-red color (Munsell hue 5YR or redder). This color may not seem surprising because some surface soils also are so brilliantly red. The red color of soils is mainly due to the nature and grain size of iron oxide and hydroxide minerals (Morris et al 1985, Torrent & Schwertmann 1987) formed by oxidation during weathering of iron-bearing minerals in the parent material. Brown hydroxides such as goethite form at first and then dehydrate and recrystallize to coarser grain size with further weathering to the red iron oxide, hematite. Red soils are found mainly in tropical regions and on land surfaces hundreds of thousands to millions of years old (Birkeland 1984).

There are many red paleosols that appear neither to have formed under a tropical climate nor to have been so strongly developed (Retallack 1976, 1977). For these paleosols, a more likely explanation of their red color is burial dehydration and coarsening of crystal size of iron hydroxides and oxides, resulting in the formation of hematite during burial (Blodgett 1988). The results of this burial process are petrographically indistinguishable from those of the formation and maturation of iron hydroxide minerals in surface soils. Iron-bearing minerals are mantled with weathering rinds and wisps of opaque red hematite crystals that extend into the matrix. In a few Quaternary paleosols and comparable surface soils of central North America for which this change has been assessed, the change in color during shallow burial for only a few tens to hundreds of thousands of years amounted to 2 to 3 Munsell hue units, from 10YR to 7.5YR or from 7.5YR to 5YR. No change in hue was noticed in drab-colored (5Y) paleosols compared with similar waterlogged surface soils (Simonson 1941, Ruhe 1969).

Such changes in hue are especially evident from paleomagnetic studies of red paleosols of Miocene age from Pakistan (Retallack 1991). These paleosols show at least three distinct components of magnetization that can be isolated by different temperatures of demagnetization: (*a*) a low-temperature (less than 300°C) pigmentary iron stain magnetized in the general direction of the modern magnetic field; (*b*) a medium-temperature (300–650°C) pigmentary hematite magnetized in the direction thought to have prevailed during soil formation; and (*c*) a high-temperature (650–685°C) specular hematite, randomized in its magnetic direction within pedogenically altered zones but more consistently oriented in paleosol C horizons and sediments, and taken to reflect the magnetic field prevailing during deposition of the sediments (Tauxe & Badgley 1988). In some cases, two separate magnetization directions can be recognized at medium temperatures of sample preparation (Tauxe & Badgley 1984). In all cases, the red paleosols contained a more complex mix of magnetization directions than associated red sediments or gray paleosols and sediments, which were not so altered by oxidizing soil formation. Thus, the high-temperature component is thought to have formed during deposition of the alluvial parent material of the soils. This view is also supported by a conglomerate test, among other observations. The alignment of all but the low-temperature components with the ancient magnetic field, even within several paleosols in a sequence deposited over the time of field reversal temporally constrained to less than a few thousand years, is an indication that the medium-temperature components were acquired during formation of well-drained soils or early after burial (Tauxe & Badgley 1988). The duration of soil formation indicated by persistence of relict bedding and paleomagnetic studies (thousands of years) is less than that required (hundreds of thousands of years) for the aging of iron hydroxides to hematite in modern soils (Birkeland 1984), so it is likely that much of the medium-temperature component was produced during early burial reddening.

Surface soils as red and rich in hematite as the Pakistani paleosols are known in India, but they are mainly found on very old land surfaces of Precambrian shield areas. Surface soils of the Indo-Gangetic Plains are either drab colored when waterlogged or brown [10YR to 7.5YR (Murthy et al 1982)]. Although these alluvial soils are siltier and are forming within a less vegetated agricultural landscape than existed during Miocene time, they are similar in overall profile form and soil structures to the paleosols. Such brown hues are likely original colors for the Miocene paleosols. Their current red hue (7.5YR to 5YR) is best attributed to burial dehydration and recrystallization of iron hydroxide and oxide minerals.

It is also conceivable that red paleosols and paleochannels have been oxidized by groundwater during burial from an originally drab color (Walker et al 1981). Oxidizing groundwaters are not common and are largely found within actively recharged, sandy aquifers. Most groundwater is chemically reducing because of oxygen scavenging by microbes and by abundant mafic minerals, such as the schist and hornblende found in the Pakistani paleosols discussed above. These Miocene paleosols also are too clayey to have permitted extensive groundwater flushing. Many of these clayey paleosols are a brilliant red, but interbedded sandstones, including parent materials to red paleosols, are tan and blue-gray in color. For these paleosols, as well as for other sequences of interbedded red clayey paleosols and green-gray sandstones (Retallack 1976, Brown & Kraus 1981, 1987, Lehman 1989), burial reddening by dehydration and recrystallization of iron hydroxides is likely, but not oxidation of the paleosols during burial.

Cementation of Primary Porosity

The transformation of soft friable soil into an indurated paleosol within sedimentary rocks is in part due to the precipitation of cements of calcite, gypsum, hematite, and silica. Cementation is also a widespread process in soils at the surface, giving rise to petrocalcic, petrogypsic, and petroferric horizons and duripans (of Soil Survey Staff 1975), respectively, for the cements mentioned. It may not be easy to determine whether a particular cement formed during soil formation or burial.

A common kind of soil cement is micritic, low-magnesium calcite that can be seen to replace preexisting grains of the soil, so that the grains are deeply embayed (caries texture) and appear to float within a micritic matrix (Esteban & Klappa 1983). Zones of micritization in sequences of shallow marine limestones have been interpreted as paleosols and as evidence of subaerial exposure (Wright 1986, Searl 1989). Unlike these replacive fabrics, displacive calcite fills cavities opened by the expansion or rotation of large clods of soil or the cracking out of clods or grains (Figure 3). Displacive fabrics are characteristic of soils because such rotations and expansions of volume are unlikely under confining pressures of deep burial

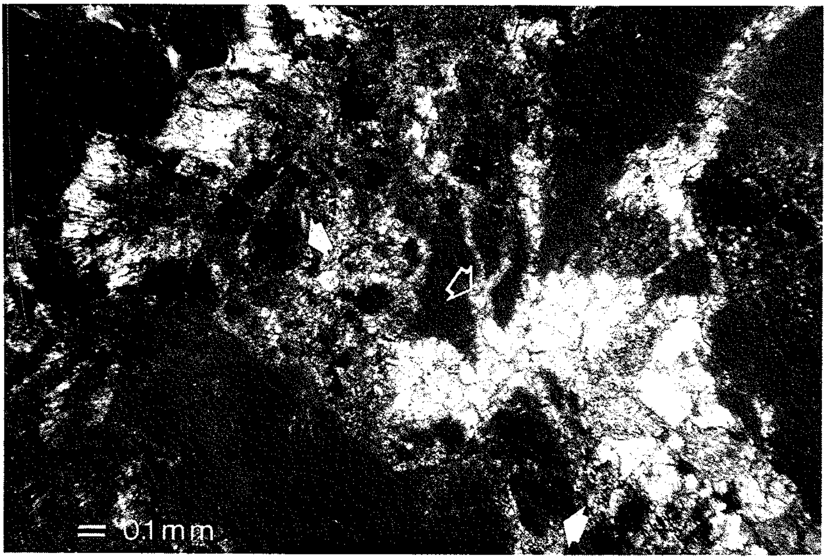

Figure 3 Replacive micrite embaying fragments of claystone (at *white filled arrows*) and displacive sparry calcite between rotated, angular clasts of claystone (at *white open arrow*) in the calcic horizon of a Natrustoll paleosol in the late Pliocene Hadar Formation at fossil site AL333 along the Awash River, Ethiopia. Scale bar is 0.1 mm. Paleosol discussed by Retallack (1990) and in work in progress with S. Radosevich and M. Taieb.

environments (Braithwaite 1989). Another microfabric characteristic of soils and paleosols is a loose aggregate of randomly arranged calcite needles, thought to be produced in soils by the activity of fungi in vugs (Wright 1986).

Displacive, replacive, and needle-fiber fabrics are found mostly in calcareous nodules of soils. Within paleosols, such nodules can be seen to be original parts of the soil by a variety of lines of evidence. Soil nodules may be resorted into overlying paleochannel sandstones or into burrows within the paleosol. Soil nodules may be sidled or avoided by root traces and burrows that approach them from the soil matrix. Some nodules form around root traces and burrows. Soil nodules may contain easily weathered minerals, such as pyroxene, that were totally destroyed by weathering in the soil matrix. Soil nodules may preserve gastropods, turtle shells, or skulls less compacted than in the clayey matrix beyond the nodule.

In contrast, calcite cements formed during burial or in groundwater tend to be sparry or bladed, cavity-lining aggregations of crystals (Searl 1989). They also tend to be concentrated in coarse-grained parts of sandy to gravelly paleosols or within open cavities such as cracks and root channels. In general, they form solid cemented beds or large nodules with smooth margins that are much less irregular or complexly mineralized than pedogenic nodules (Retallack 1991).

This account of pedogenic versus groundwater and burial calcite cements presents useful caricatures, but real situations may be complex. For example, Late Cretaceous and early Paleocene paleosols in eastern Montana, USA (Retallack et al 1987), include subsurface horizons and diffuse nodules with micritic carbonate both replacing clay skins and grains and filling etch pits in grains (as in Figure 6 of McSweeney & Fastovsky 1990). In addition to these pedogenic cements, there also are multiple generations of sparry, pore-filling cements (Figure 5 of McSweeney & Fastovsky 1990), especially in paleochannel sandstones interbedded with clayey paleosol-rich sequences. Some of the cemented areas of the sandstone take the form of large, log-shaped nodules whose orientation reflects regional groundwater flow (Jacob 1973). Also obvious in paleochannel sandstone are detrital grains of limestone and resorted calcareous nodules. These clasts may be poorly preserved in clayey paleosols and can be difficult to distinguish from pedogenic nodules. Such a complex mix of detrital, pedogenic, and groundwater carbonates is common in surface soils of seasonally dry, riverine plains, such as the Indo-Gangetic alluvium of northern Pakistan and India (Sehgal & Stoops 1972, Courty & Federoff 1985, Courty 1990).

Fortunately, there are other techniques that can be brought to bear on this problem. The carbon and oxygen isotopic compositions of surface waters are different from those of groundwaters (usually heavier), so that the isotopic composition of carbonate cements can be used to establish their relative age (Beeunas & Knauth 1985, Vahrenkamp et al 1987, Searl 1989). Compared with rain and surface waters, groundwaters tend to be reducing and offer more iron than manganese to carbonate cements. A low iron/manganese ratio within calcite cement can cause it to fluoresce brightly under cathode rays (Marshall 1988). Such cathodoluminescence studies can be used to establish the relative ages of different cements from Eh-related variations in fluorescence. These kinds of studies have so far been undertaken primarily on marine limestones, but they are now finding application to calcareous paleosols (Searl 1989).

Cements of other carbonates, such as siderite and dolomite, are known from very few soils. Siderite occurs as nodules and small spherules in underclays to peaty soils (Ho & Coleman 1969, Landuyt 1990). Dolomitic crystals, nodules, and layers are common in highly alkaline clays of dry, salty soils (Cline 1955, Botha 1988). Both siderite and dolomite are commonly interpreted as diagenetic minerals, especially when filling secondary porosity created by acidic fluids associated with kerogen maturation and cracking during deep burial (Surdam & Crossey 1987).

Silica is a common cement in desert soils and playa lakes of high pH—conditions that markedly enhance silica solubility for the short periods when water is available in such environments (Summerfield 1983). Remobilization of silica under highly alkaline conditions is likely for small silica nodules or for silica pseudomorphs of gypsum roses or other desert salts (Retallack 1983). Voluminous cements of silica in paleosols more likely form during deep burial, in part as a consequence of intergranular pressure solution (Houseknecht 1988) and of dewatering and illitization of smectitic clays (Weaver 1989). Strongly silicified surface horizons of paleosols could have derived much silica from plant opal, which is an easily mobilized form of silica compared with quartz dissolution or clay illitization (Retallack 1976). This remobilization of plant opal must also have been a phenomenon of burial because such silicified surface horizons, or ganisters as they are widely called (Percival 1983), preserve fossil root traces that were unimpeded by cementation.

Hematite, goethite, and other iron oxides and hydroxides are common cements in soils and are particularly abundant in the reddish subsurface horizons of acidic sandy soils (the spodic horizon of Spodosols in the terminology of Soil Survey Staff 1975). These soil horizons have a characteristic micromorphology of concentric layers of iron stain around the grains, with wide radiating desiccation cracks (de Coninck et al 1974). The whole horizon may be deflected around tree stumps (a basket podzol) or redeposited within deep burrows or root holes (a krotovina): These features are also known from paleosols (Retallack 1976, 1985). Arguments have been advanced (by Walker 1967, Walker et al 1967, 1981) for extensive cementation of red beds by oxidation of iron-bearing minerals during burial, presumably by oxidized groundwater. Most of these arguments are based on the distribution and nature of iron-bearing minerals in sediments. They carry little weight for soils and paleosols, which may show a very wide variation in degree of ferruginization. For the red clayey paleosols of my experience (Retallack 1976, 1983, 1985, 1991, Bown & Kraus 1981, 1987), burial reddening of ferric hydroxides is likely, but not burial oxidation, because of associated unoxidized paleochannel sandstones, among other reasons already discussed.

Gypsum is a common cement in desert soils (Porta & Herrero 1990) and evaporite sediments (Warren 1989). Many so-called "sediments" of playa lakes and coastal sabkhas are formed by the in-place growth of gypsum and other salts that is the hallmark of soil formation (as in Salorthids of Soil Survey Staff 1975), rather than the accumulation of a precipitate that would be better regarded as a chemical sediment. Dissolution of salts on burial appears to have been more widespread than their precipitation. Many evaporite beds and salty paleosols or salt crystals in paleosols have been dissolved into brecciated layers (Bowles & Braddock 1963) or replaced with other minerals during burial (Barley et al 1979).

Lithostatic Compaction

Burial of paleosols results in compaction as the void spaces, organisms, and water are crushed by the weight of overburden. The compaction of originally loose soil clods against one another could create a complex pattern of slickensides superficially similar to the more regular slickensides of swelling-clay soils of seasonally dry climates (Paton 1974, Gray & Nickelsen 1989). Changes in thickness due to compaction can also compromise interpretations of former rainfall based on depth within a paleosol to the horizon of calcareous nodules (Jenny 1941, Retallack 1991). Such studies of paleosols may depend on accurate estimates of compaction.

The best indicators for the degree of compaction of paleosols are clastic dikes—cracks in the soil that were more or less vertical and subsequently were filled with contrasting material. With compaction, clastic dikes are deformed ptygmatically, so that the amount of compaction can be calculated by the ratio of the vertical height of a clastic dike and the distance around undulations of the dike. Unfortunately, such features are found in only a few kinds of soils and paleosols: the swelling-clay soils, or Vertisols, and their buried equivalents (Gray & Nickelsen 1989). When present, clastic dikes can be used not only to reconstruct the thickness of the original profile, but also to recalculate the original chemical composition of a paleosol (Retallack 1986b).

Other features useful for estimating compaction include fossil skulls, turtle shells, snails, logs, burrows, and the spacing of uncompacted laminae within nodules. Unfortunately, these various kinds of fossils can collapse before burial, because of decay in and on a soil. There are similar problems in calculating the compaction of paleosols from their bulk density, usually estimated by the clod method (of Blake & Hartge 1986), compared with the bulk density of comparable surface soils. Bulk density varies considerably within modern surface soils. Another approach is to point count the number of planar, concavo-convex, and sutured contacts between grains in sandstones, because sutured contacts become more abundant at greater depths (Taylor 1950). This kind of analysis must be restricted to a particular mineral (quartz is usually best) because different kinds of grains deform differently under compaction.

For most paleosols, compaction is best estimated by comparison with standard compaction curves now available from the statistical study of large numbers of boreholes in sandstones, claystones, and coals (Ryer & Langer 1981, Elliott 1985, Baldwin & Butler 1985). The standard error on these curves is large for a variety of reasons. Quartz sandstone is less readily compacted than lithic sandstone, calcareous or pyritic shale less than carbonaceous shale, and peat formed from wood (coal rich in vitrinite) less than peat formed from mosses or leaves (coal rich in exinite). In peaty soils, there can be considerable compaction at the base of the organic horizon, even before burial (Elliott 1985). Similarly, dry footworn soils may have been compacted before burial to a greater extent than marine sediments and lowland soils, which formed most of the rock sequences used to construct compaction curves.

One way of using these curves is to measure the porosity of the sample, or preferably the percentage of solid grains (Baldwin & Butler 1985), and use this figure as an index of compaction to establish a depth of burial. Considering the magnitude of likely errors and the way compaction curves steepen with depth, this approach is of limited usefulness. More consistent results can be gained by estimating the depth of burial by geological methods and then estimating likely compaction from the standard curves. For Miocene paleosols from Pakistan (Retallack 1991), the depth of burial could be calculated in several ways: by assuming a monotonic thickness decrease from north to south and east to west from sections where both tops and bottoms of overlying formations are exposed, or by extrapolating the paleomagnetically determined rate of sediment accumulation from the age of the paleosols to the time of initial uplift of the region. Both the paleomagnetic and geological estimates were compatible with occasional concavo-convex contacts between quartz grains in thin sections of sandstones associated with the paleosols and with complete lack of sutured contacts. If burial by roughly 2000 m is used with standard compaction curves (Baldwin & Butler 1985), the clayey parts of the paleosols should be about 58% of their former thickness and the sandy parts about 70%.

Illitization of Smectite

Many clayey paleosols are predominantly illitic in composition, and in many soils illite or interlayered illite-smectite is the main clay mineral. This is especially true in young soils of desert regions with strong wet-dry seasonality (Robinson & Wright 1987, Weaver 1989). Considering the likely low demand for potassium of microbes and early land plants, unlike that of forests, illite-rich soils could have been much more widespread in the distant geological past than they are today (Weaver 1989). Although these explanations have some appeal, many illitic paleosols lack evidence of sodium or calcium enrichment within the salty or carbonate-rich horizons (natric or calcic horizons of Soil Survey Staff 1975) characteristic of desert soils. Nor do they show the cracking and undulation of horizons (mukkara structure of Paton 1974) found in soils of wet-dry seasonal climates (Grandstaff et al 1986, Farrow & Mossman 1988, Wiggering & Beukes 1990). Even paleosols of Paleozoic age with copious large root traces of the kind formed under forests are predominantly illitic (Retallack 1985).

An additional explanation for the illitic composition of many paleosols is the alteration of smectite to illite during burial, a process now widely documented from studies of boreholes. The overall chemical effects of this alteration are displacement with K^+ of Ca^{2+}, Mg^{2+}, and Na^+, which are lost with water of hydration, alumina, and silica (Bethke & Altaner 1986). This has been regarded as a solid-state replacement of interlayer cations with preservation in illite of most of the 2:1 layer structure of precursor smectite. However, X-ray diffraction and transmission electron microscopy studies show that the alteration proceeds by destruction of small crystals of smectite by interstitial fluids, followed by the neoformation and growth of increasingly large crystals of illite (Eberl et al 1990). The source of the potassium may be minerals such as microcline, to judge from concomitant illitization and loss of microcline in deep boreholes (Hower et al 1976). Microcline and other potash minerals may have been hydrolyzed by solutions rich in carbonic acid generated during kerogen maturation and cracking (Weaver 1989), but a case also has been made for the illitization of paleosols by groundwater containing potassium leached from surface soils (Nesbitt & Young 1989). Pervasive local potash enrichment also occurs around hydrothermally altered veins (Palmer et al

1989, Holland et al 1989a, Eberl et al 1990) and certain kinds of igneous intrusion (Le Bas 1977).

Unlike local hydrothermal and igneous potash enrichment, diagenetic illitization commonly is limited in its thoroughness. The ionic exchange between microcline-rich sandstones and smectitic claystones may be restricted to distances of only a few tens of centimeters. Gradients of illitization have been observed away from a likely source bed (Altaner et al 1984). Clayey paleosols may remain unillitized under pure quartz sandstone (Kimberley & Grandstaff 1986). The illitization seldom is complete under conditions of burial short of metamorphism. Some remnant smectite is common (Hower et al 1976). Illitization also may proceed during a specific window of burial conditions and then slow or stop, as indicated by radiometric dating of shales in boreholes (Morton 1985). The slowing of the process may be due to dewatering during burial and increased scarcity of readily altered potassium bearing minerals.

Illitization of shales becomes readily discernible when shales are buried at depths between 1.2 and 2.3 km and reach burial temperatures of from 55 to 100°C, judging from other indications of burial conditions in borehole studies (Weaver 1989). Theoretically, however, the transformation of smectite to illite could occur at much lower temperatures over very long periods of geological time. For example, Ordovician shales may be illitized even if only buried by a few hundred meters (Bethke & Altaner 1986).

One way to demonstrate illitization is through bulk chemical analysis of paleosols. Those paleosols showing a surficial increase in the amounts of potash, without a concomitant surficial increase in soda or local enrichments in lime found in desert soils, probably have been illitized (Retallack 1986b). This anomalous combination of chemical depth functions can be displayed graphically in a triangular plot of carbonate-free recalculated molar proportions of lime and soda vs alumina vs potash (Figure 4). Unillitized profiles on such a plot should fall on a line heading toward the alumina corner with increased weathering or proximity to the surface of the profile, whereas illitized parts of the profile will deviate out toward the potash corner (Nesbitt & Young 1989, Rainbird et al 1990). Another approach is to take the chemical analyses to calculate the potash/alumina stoichiometry of the clay fraction, which can in some cases approach the composition of muscovite rather than the usual composition of illites found in soils (Feakes & Retallack 1988). Another approach is to calculate the normative mineralogical composition of a paleosol from chemical analysis, using a large spread sheet and minerals judged likely to have been in the paleosol from petrographic and other observations (Garrels & Mackenzie 1971). For illitized paleosols it may be necessary to use illite

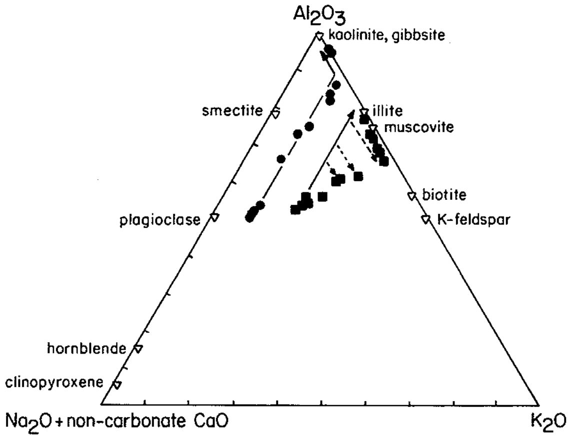

Figure 4 Chemical analytical data for a surface soil on the Toorongo Granodiorite, Mt. Baw Baw, Victoria, Australia (*closed circles*) and for a pre-Huronian (more than 2300 Ma) paleosol on the Ville Marie Granite near Lake Timiskaming, Quebec, Canada (*closed squares*), plotted with the composition of selected minerals (*open triangles*) on an A-CN-K diagram. Figure from Rainbird et al (1990), with permission of the authors and the *Journal of Geology*.

compositions unrealistically close to those of muscovite in order to solve the equations (Retallack 1986b).

Yet another approach is to evaluate illite "crystallinity" from the sharpness of the 10-Å peak on X-ray diffractometer traces. Illites formed during burial have sharp narrow peaks, whereas those formed in soils have broad low peaks. A number of indices of peak sharpness can be used (Frey 1987). For example, the Weaver index is the peak height above background at 10 Å divided by the peak height at 10.5 Å, and the Kubler index is the width in degrees 2θ of the 10-Å peak at half the level of its height. The onset of illitization is marked by an increased proportion (more than 4 : 1) of illite interlayers within smectites but only a slight improvement of illite crystallinity. Illitization is very marked, and little smectite remains at what has been called the transition from diagenesis into the anchizone of metamorphism, which corresponds roughly to a burial depth of about 5 to 7 km, or temperature of about 200°C, and to a Weaver index in excess of 2.3 or a Kubler index less than $0.42°\Delta 2\theta$.

Coalification of Peat

Very few peats accumulate from rafted plant debris. Almost all coal seams are formed by alteration during burial of peats that accumulated under swamp, marsh, or other kinds of vegetation that can tolerate waterlogged ground. In soil science, peats formed in place are called histic epipedons and are the most characteristic feature of a Histosol soil (of Soil Survey Staff 1975). In a sense, then, most coal seams are paleosols, although their original profile form is altered considerably by compaction and chemical alteration of the peat during deep burial. The chemical alteration of peat during burial has been studied for many years (Stach et al 1975). It is primarily an enrichment in carbon with the loss of water, nitrogen, and other volatile materials. As a result, a soft, dull, brown peat is transformed into a brittle, shiny, black coal. The most commonly used chemical index of coalification is total organic carbon. The main physical index is the reflectance of a particular component of coal: the maceral vitrinite that is derived from wood fragments. The relationship between organic carbon content or vitrinite reflectance and depth or temperature of burial has been established in many deep sedimentary basins (Stach et al 1975).

Coalification also affects organic materials other than peats. The browning, and then blackening, of normally white to gray-colored conodonts with increased depth of burial is useful in estimating the depth of burial of the marine sediments containing them (Epstein et al 1977). Two techniques already used in this manner in nonmarine rocks are blackening of fossil pollen and spores (Wilson 1971) and the increased reflectance of isolated pieces of organic matter with increased burial (Tissot & Welte 1984). These techniques require local calibration for specific sedimentary basins, but the information they provide on burial conditions can be invaluable in sorting out the effects of other alterations after burial.

Kerogen Maturation and Cracking

A widespread chemical reaction during soil formation is the hydrolysis of aluminosilicate minerals to clay and cations in solution. The cations are displaced by hydronium in weak solutions of carbonic acid, created principally by respiration of carbon dioxide by soil organisms (Birkeland 1984). Similar destruction of grains also can occur during deep burial in acidic solutions generated during the decarboxylation of buried organic matter (kerogen), its maturation to oil, and its ultimate cracking to natural gas (Surdam & Crossey 1987).

Secondary and primary porosity can appear similar under the microscope, with deeply etched grains or hollow clay skins remaining from destruction of the grain within. Characteristic of secondary porosity are oversized vugs, which show pronounced elongation and alignment and cut sharply across other grain boundaries (Schmidt & McDonald 1979). These kinds of cavities are uncommon in secondary porosity but are very distinct from primary pores in soils, which do not tend to cut sharply across grain boundaries and are not so well aligned. Soil pores also tend to be altered around the margins with clay skins or oxidation rinds. In most paleosols, primary pores are crushed during burial compaction, so that abundant inflated vugs may indicate secondary porosity. Nevertheless, some root holes and vugs in soils fortified by cements during soil formation or early after burial can withstand compaction.

Secondary pores also may be distinguished from primary pores by cementation with minerals stable under Eh and pH conditions that do not make sense for the host paleosol (for example, siderite cement in a red, hematite-rich paleosol). Cathodoluminescence and isotopic studies of cements may also be useful for distinguishing between the primary pores of an ancient soil and secondary pores imposed during deep burial (Marshall 1988, Searl 1989).

For some paleosols, the influence of decarboxylation and kerogen maturation and cracking can be ruled out by evidence, such as low vitrinite reflectance or poor illite "crystallinity," for shallow burial. Abundant oil, wet gas, and organic acid is generated from organic source rocks at burial temperatures beyond those needed to create vitrinite reflectance values greater than 0.5%, at which point coals reach the bituminous grade. Kerogen cracking to dry natural gas and carbon dioxide is greatest when vitrinite reflectance exceeds 2%, as in higher rank bituminous coal and anthracite (Tissot & Welte 1984). This "oil window," or interval of catagenesis in burial history, is typically between burial temperatures of 50 and 150°C, or at depths of some 1 to 5 km, depending on local geological age, tectonic history, and geothermal gradient. Assessing whether a particular paleosol has been affected by fluids generated during kerogen maturation and cracking can be as complex as finding oil and relies on the same kinds of information.

Recrystallization of Carbonate

Many paleosols are strongly calcareous, and calcite tends to recrystallize to coarse grain size during burial (Bathurst 1975). Such recrystallized calcite may display distinctive features, such as equiangular crystal junctions or cross-cutting relationships, in thin section.

Recrystallization can be especially obvious when it affects fossil shells, such as those of snails, converting their original fibrous or prismatic microstructure to a sparry mosaic. Similarly, it also has been assumed that calcareous nodules in Paleozoic paleosols with scattered replaced and embayed grains were recrystallized to sparry calcite from the more common micritic texture of soil nodules (Retallack 1985). Isotopic and cathodoluminescence studies are now being applied to the study of such nodules (Searl 1989). Recrystallization may be part of the reason why needle-fiber calcite, which is common in arid-land soils (Courty 1990, Guo & Federoff 1990), has been found in only a few paleosols (Wright 1986).

Evaluation of burial conditions using such indices as vitrinite reflectance or illite "crystallinity" also may be useful. Recrystallization is generally not extensive under conditions of burial more shallow or cool than those required for illitization beyond the anchizone or coalification beyond bituminous grade.

Metamorphism

The various kinds of alteration already outlined pass into metamorphism beyond the point of illitization, kerogen cracking, and coalification to anthracite grade. This transition is passed at temperatures somewhere in excess of 200°C or depths greater than 7 km, whichever comes first. A precise boundary is difficult to specify. In practice, it is best to label a rock as being metamorphic if it has newly formed structures such as schistosity, or minerals such as clinozoisite.

Many metamorphosed paleosols have now been reported. Especially common minerals in metamorphosed paleosols are corundum, andalusite, and sillimanite, which are to be expected in metamorphic rocks derived from such aluminous protoliths (Grandstaff et al 1986, Reimer 1986, Dash et al 1987, Golani 1989, Bol et al 1989). Illitized clays in paleosols may be further altered during metamorphism to sericite, chlorite, paragonite, and muscovite (Retallack 1986b, Holland et al 1989b).

Even with extensive metamorphism, it is surprising how well the chemical signature of soil formation survives. Chemical depth functions similar to those of soils in oxides other than potash have been widely reported in paleosols metamorphosed to greenschist facies (Grandstaff et al 1986, Farrow & Mossman 1988, Zbinden et al 1988, Holland et al 1989b) and, in some cases, even to amphibolite facies (Barrientos & Selverstone 1987). Hematite, as well as red color, is preserved in paleosols metamorphosed to greenschist facies (Retallack 1985) and does not appear to be converted to gray magnetite until the amphibolite facies (Thompson 1972). Textural and chemical alteration owing to metamorphism is in most cases easily distinguished from that owing to soil formation. Highly metamorphosed paleosols, however, can be difficult to distinguish from hydrothermally altered rocks (Barrientos & Selverstone 1987, Williams et al 1988, Palmer et al 1989, Holland et al 1989a,b).

Conclusions

The great array of potential alterations during burial that could compromise interpretation of paleosols can be intimidating, but as research proceeds, common patterns of alteration are emerging. Burial decomposition, burial gleization, burial reddening, and cementation can significantly alter the appearance of paleosols very shortly after burial. This is not to deny the importance of alteration during deep burial, such as illitization, coalification, kerogen maturation, and recrystallization. Many physical features of paleosols, such as horizonation and root traces, persist through these deep burial alterations well into the greenschist facies of metamorphism. It is not until metamorphism to amphibolite facies that physical structure is completely obscured by recrystallization.

The effects of burial on soils are best appreciated by considering specific examples of soils and paleosols (Figure 5) that represent a common pattern of alteration of weakly to strongly developed soils under forest or woodland vegetation in well-drained parts of lowland landscapes and subhumid to humid, nonfrigid climates. The original soil may be clayey, brown, and friable, with a dark-gray surface (A horizon) containing decaying organic matter over a brown subsurface (Bt horizon) enriched in clay and penetrated by stout root traces. The effects of burial decomposition, burial gley, and burial reddening may convert this to a strange caricature of a soil: a claystone bed with very low amounts of organic carbon, with a

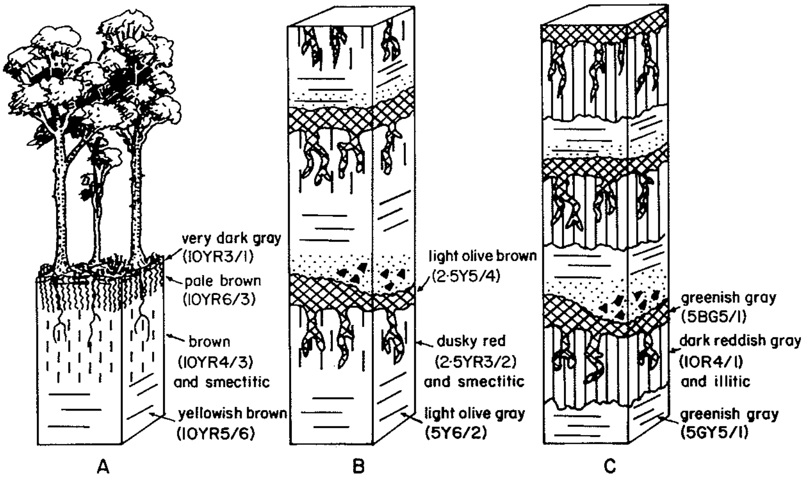

Figure 5 Hypothetical alteration of a woodland soil by burial gleization of organic matter, burial reddening of iron hydroxide and oxide minerals, and compaction and illitization of smectite. The profiles illustrated are (*A*) a surface soil in Iowa (#7 in appendix of Buol et al 1980), (*B*) early Eocene paleosols in the type area of the Sand Creek Facies of the Willwood Formation of Wyoming (Bown & Kraus 1981), (*C*) Late Devonian Peas Eddy clay paleosol of New York (Retallack 1985).

former surface horizon of green-gray claystone that penetrates as haloes around root traces into a brick-red clayey subsurface horizon. A subtly colored gray to brown soil has thus been converted to a green-red mottled bed as gaudy as a candy cane. There are many such paleosols. With additional burial and metamorphism to lower greenschist facies, such a profile may darken so that its surface horizon is dark grayish green and its subsurface purple, its clays may coarsen to silt size with the development of illite and chlorite, and its irregularly oriented soil structures may become overprinted by schistosity. Paleosols of this kind also are known.

Alteration of paleosols during burial is of interest in itself because soils are a distinctive class of materials that behave somewhat differently than sediments during burial. Untangling the alteration of paleosols after burial from alteration due to soil formation is also an important prerequisite for interpreting paleoenvironments from paleosols. The environmental significance of the amount of organic carbon, of the oxidation state, and of the clay mineral composition of soils is now well established. Studies of these aspects surface soils must be used with care to interpret paleoenvironments from paleosols, and in many cases they may not be at all useful. Fortunately, there remains an impressive residue of soil features in paleosols that can be used for environmental reconstruction. Physical features such as nodules, root traces, and burrows are particularly robust in the face of alteration after burial. These also deserve scrutiny to determine the extent to which paleosols as natural essays on landscapes of the past have been smudged and torn by alteration after burial.

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