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3

FOSSIL SOILS: INDICATORS OF ANCIENT TERRESTRIAL ENVIRONMENTS

Greg Retallack

INTRODUCTION

The study of fossil soils, particularly those older than Pleistocene, is now gaining widespread recognition as an emerging area of earth science. Such studies have great potential for reconstructing ancient environments. Fossil soils can provide evidence for the size and structure of ancient vegetation and its location relative to sedimentary environments, for paleotopography, for depth to the water table and its chemistry, for rates of relative sedimentation, subsidence and uplift, and for paleoclimate. Fossil soils are proving abundant and widespread in nonmarine rocks of all ages and are the only aspect of the geology of many areas remaining completely undescribed. Studies of fossil soils may commonly be combined with existing studies of fossil plants and vertebrates, sedimentary environments, paleocurrents, and paleoclimates to give a detailed impression of ancient environments and ecosystems.

As research on older fossil soils gains impetus, there is also the prospect of documenting the diversification of the world's soils through geological time. Studies of fossil soils may provide critical evidence for understanding the development of the modern atmosphere during the Precambrian and early Paleozoic, the origin of terrestrial organisms during the late Precambrian and early Paleozoic, the first vascular land plants and the first forests of the mid-Paleozoic, the emergence and spread of savanna and grassland during the Tertiary, and the recent impact of human beings on the land surfaces of the earth.

Compared to other branches of earth sciences so little is known of fossil soils, particularly those older than Pleistocene, that a review of them does not cover an impossibly large literature. This review is biased toward my personal conviction that interpretation of ancient environments is the most promising future direction for studies on fossil soils. Pleistocene buried soils and altered rocks at major geological unconformities are the best known fossil soils, but they are discussed here in less detail than well-preserved fossil soils within thick sedimentary sequences that also preserve other kinds of fossils. Because of difficulties peculiar to Pleistocene fossil soils and geological unconformities, the study of older fossil soils has often been approached timidly or avoided in the past. Caution is certainly needed, especially in unraveling the complex effects of sedimentation, volcanism, diagenesis, and metamorphism from ancient soil-forming processes. However, more and more studies of older fossil soils are overcoming these obstacles. It is now apparent that careful studies of fossil soils can provide evidence for many features of ancient environments that were previously indeterminable and can also support and integrate conclusions from a number of other earth sciences into surprisingly detailed reconstructions of the past. It is also becoming apparent that fossil soils are much more abundant in nonmarine rocks than generally realized. Many enigmatic kinds of rocks, masquerading under a variety of uninformative names are now turning out to be (at least partly) fossil soils. Among these are redbeds, variegated beds, badlands, cornstone, ganister, tonstein, underclay, and fireclay (Williamson 1967; Steel 1974; McBride 1974; Retallack 1977a, 1979).

What are fossil soils and how are they recognized? The most practical definition of a fossil soil (also called a paleosol) is a former soil buried by later deposits. The main difficulty with this definition arises, not from its fossil nature, but from the concept of modern soil, which is different for engineers, agriculturists, geologists, and soil scientists (Ruhe 1965; Hunt 1972). I prefer to broadly define soil as material on the surface of a planet altered by physical or chemical weathering, the action of organisms, or all of these. Fossil soils can be recognized by any of the features of modern soils. For older fossil soils the most diagnostic feature is the remains of fossil roots preserved in growth position. Other features include leached or reddened, massive-looking and clay-rich layers, prismatic or blocky jointed layers, and a variety of trace fossils, mottles, nodules, and concretions. The micromorphology of the fossil soil in thin section and its clay mineralogy and geochemistry are also useful in the study of older fossil soils, if interpreted with care.

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In Quaternary Sediments

Fossil soils of Pleistocene and Recent age have been used for stratigraphical mapping in many parts of the world (Ruhe 1965; Gibbs 1971; Paepe 1971; Jessup and Norris 1971). The "Sangamon soil" of North America has been mapped from Ohio through Illinois to Texas over a distance of more than 3,000 km. This and other fossil soils are useful correlation surfaces in complex glacial and periglacial sediments. As mapping units these have usually been labeled "soil," in accordance with the American Code of Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature 1961). Suggested alternative names include "pedoderm" (Brewer, Crook, and Speight 1970), "geosol" (Morrison 1968), and "profile"; the latter specifically for deeply weathered rock units of uncertain relationships (Senior and Mabbutt 1979). The problems of mapping Quaternary paleosols are ably discussed by Ruhe (1965). Similar fossil soils may have formed under similar conditions and may not be the same age everywhere. Soil formation may be progressively or locally interrupted by sedimentation at different times and places. Particular care must be taken when following fossil soils along strike as they may change character where they formed in different parts of the landscape.

Quaternary paleosols have also been used to reconstruct ancient environments. Ruhe (1970) found that the evidence of fossil soils was in agreement with palynological and other evidence that much of North America now under prairie was wooded and received more rainfall during the last interglacial. Similarly, Ložek (1967) has reconstructed the changing Pleistocene climate and vegetation of Czechoslovakia, largely on the basis of fossil soils. One great advantage in interpreting younger fossil soils is that they can often be matched closely with modern soils.

The problems with interpreting ancient environments from Quaternary paleosols are due largely to their occurrence on stable cratonic areas. Such land surfaces may be so stable that the same surficial material is altered in different ways by successively changing climates and vegetation. Such soils, formed under conditions unlike those of today, are called relict soils. The different phases of weathering in them can be very difficult to disentangle. As many cratonic areas subside only slowly or may even be rising, paleosols may only be partly buried or even exhumed. Even soils that are completely buried may be covered by such a thin layer of sediment that they are subjected to additional weathering at depth. Quaternary paleosols are not easy to interpret, although they are geologically young. These various difficulties make it easy to become pessimistic about studies of older paleosols. These difficulties apply to all fossil soils, but fortunately they are not as severe in other geological settings.

At Major Unconformities

Fossil soils are also commonly recognized at major unconformities. The original geological unconformity, first recognized by Hutton in 1787, shows several features of Siluro-Devonian soil formation. The surface has topographical relief of at least 400 m and is reddened and fissured, with fissure fills of red, calcareous, sandy breccia (Friend, Harland, and Gilbert-Smith 1970). Other examples of paleosols at unconformities are the "lateritic" paleosols which have been widely recognized at the unconformable contact between Cretaceous and early Tertiary rocks in many parts of the western United States (Wanless 1923; Pettyjohn 1966; Abbott, Minch, and Peterson 1976; Thompson, Fields, and Alt 1977).

Although easy to recognize, because they indicate millions of years of erosion and nondeposition, this in itself is a problem for their interpretation. Many features of these paleosols may be relicts of soil formation under a climate and vegetation very different from those just before burial. A more serious problem with such paleosols arises from the way unconformities commonly juxtapose rocks of different porosity, mineralogy, and other characters. Unconformities are especially susceptible to later modification by hydrothermal alteration, diagenetic or metamorphic changes involving reaction between the contrasting materials, and leaching or precipitation of minerals by intrastratal solutions or groundwater. The difficulties of unequivocally distinguishing between later modification and original weathering are well illustrated by the disagreement of Lewan (1977) and Kalliokosi (1977) over the nature of altered rocks underlying the 1-billion-year-old Jacobsville Sandstone, north of Marquette, Michigan. The need for caution can also be seen from the following example. In the driftless area of southwestern Wisconsin, successive early Paleozoic sandstones may have strongly silicified and ferruginized crusts and mottled and pallid zones immediately below each capping carbonate unit. Evidence presented by Dury and Haberman (1978) indicates that these are not early Paleozoic paleosols, but were more likely produced by deep weathering during the late Cretaceous and early Tertiary.

In Thick Terrestrial Sedimentary Successions

The ideal situation for the preservation of fossil soils is in sedimentary basins subsiding at such a rate that soils are covered and sink below the water table shortly after reaching the greatest differentiation possible, given the parent material, vegetation, and climate at the time. As discussed by Allen (1974b), this may occur without episodic basin-wide subsidence and sedimentation. The covering of soils at long intervals is a natural consequence of the restricted supply route of sediment, such as mudflows and streams, in many terrestrial environments. Streams constantly change course or meander laterally, but any given portion of their floodplains is infrequently reworked by the stream channel or covered by thick near-channel

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flood deposits. With slow steady subsidence large numbers of fossil soils may become superimposed. The buried Eocene forests of the northeastern part of Yellowstone National Park, Wyoming, consist of about 27 successive buried surfaces containing petrified tree trunks in about 400 m of stream gravel, mudflows, and tuffs (Fig. 3.1). Although some of the petrified trunks may have been transported (Fritz 1979), at least two of the surfaces (which I have personally examined on Specimen Ridge) have a leached horizon overlying a clay-rich horizon and large petrified tree trunks in growth position. Dorf (1964) has counted as many as 500 growth rings in some of these buried stumps, indicating at least that many years of soil formation on some of the surfaces.

Recognizing and interpreting paleosols in such sedimentary settings is not without problems. Some features of a fossil soil may be relicts from the formation of another fossil soil (or pedorelicts in the terminology of Brewer 1964). For example, nodules or clods of an older soil may have been eroded and incorporated in the parent material of a younger soil. If the sediment overlying a buried soil was only thin, the upper portion of the buried soil may also have been altered by soil formation at the higher land surface. This may result in a B horizon of one soil containing structures of the relict A horizon of an older buried soil, as documented in the type Long Reef clay paleosol in the Triassic rocks near Sydney, Australia by Retallack (1977b). In immature paleosols many sedimentary features of the parent material may not have been obliterated. These sedimentary relicts (as they are termed by Brewer 1964) are most commonly bedding and ripple-drift cross-lamination. The upper layers of fossil soils may also have been removed by erosion. This soil material may have been transported and deposited to form a pedolith (in the sense of Gerasimov 1971), a rock unit with sedimentary organization but soil mineralogy and clast microstructure. In some sedimentary successions pedoliths are difficult to distinguish from ordinary sediments, in others they are distinct. Fortunately, subsidence rates in many sedimentary basins are such that these various problems are not insuperable. In such basins, not only fossil soils, but also fossil plants and animals may be well preserved. These can then be integrated into surprisingly detailed reconstructions of past environments (Fig. 3.2). Fossil soils in thick sedimentary successions have been the least studied in the past, but show most promise for the future.

Roeschmann (1971) has indicated that later alteration by diagenesis, metamorphism, or intrastratal solution can also be an obstacle to the interpretation of older paleosols in sedimentary sequences. Each feature of the fossil soil must be assessed separately to determine whether it is due to original soil formation or later alteration. Physical field relations and petrographical textures are usually more convincing evidence in such deliberations than mineralogy or geochemistry. For example, in the Avalon

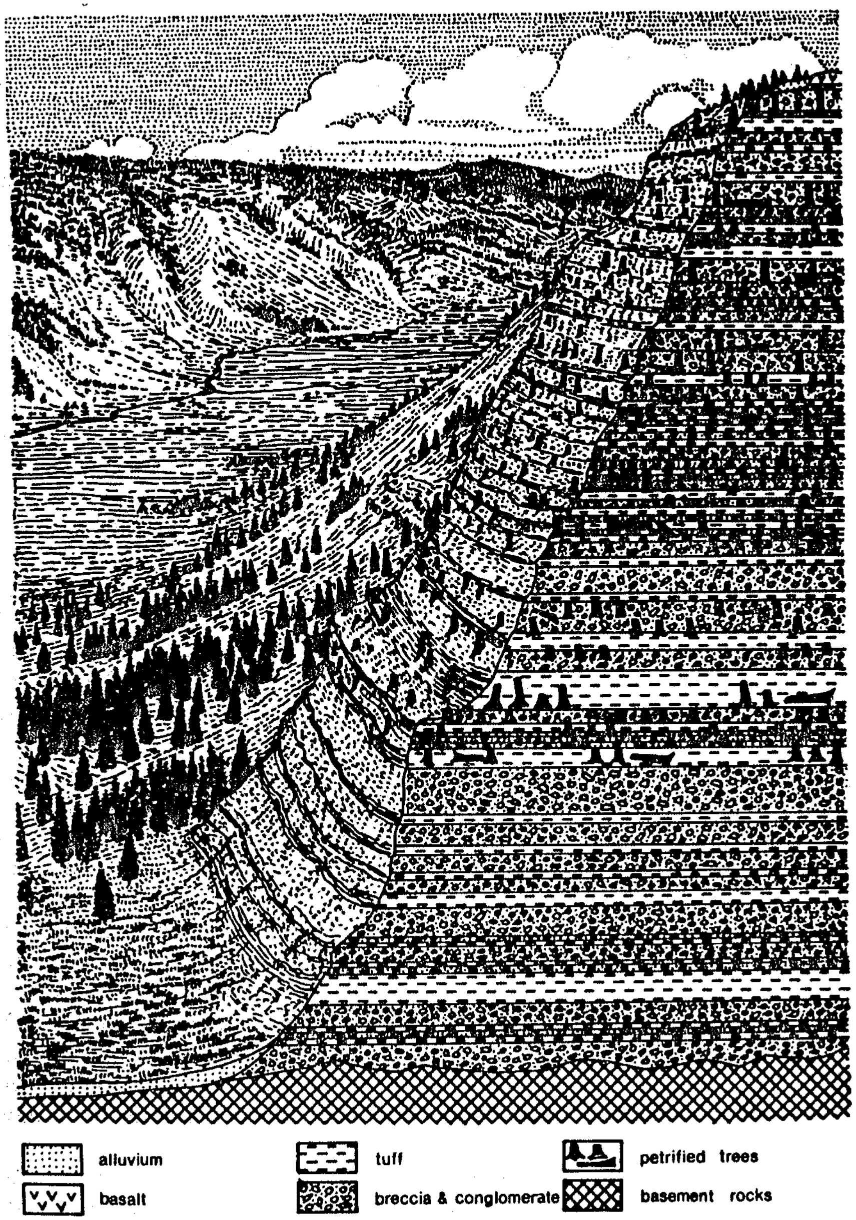

FIGURE 3.1. The Eocene buried forests of Yellowstone National Park, Wyoming. About 27 horizons bearing petrified stumps occur in 400 m of fossiliferous rock. (Modified from Dorf. E. April 1964. The petrified forests of Yellowstone Park. *Scientific American* 210. Copyright © 1964 by Scientific American, Inc. All rights reserved.)

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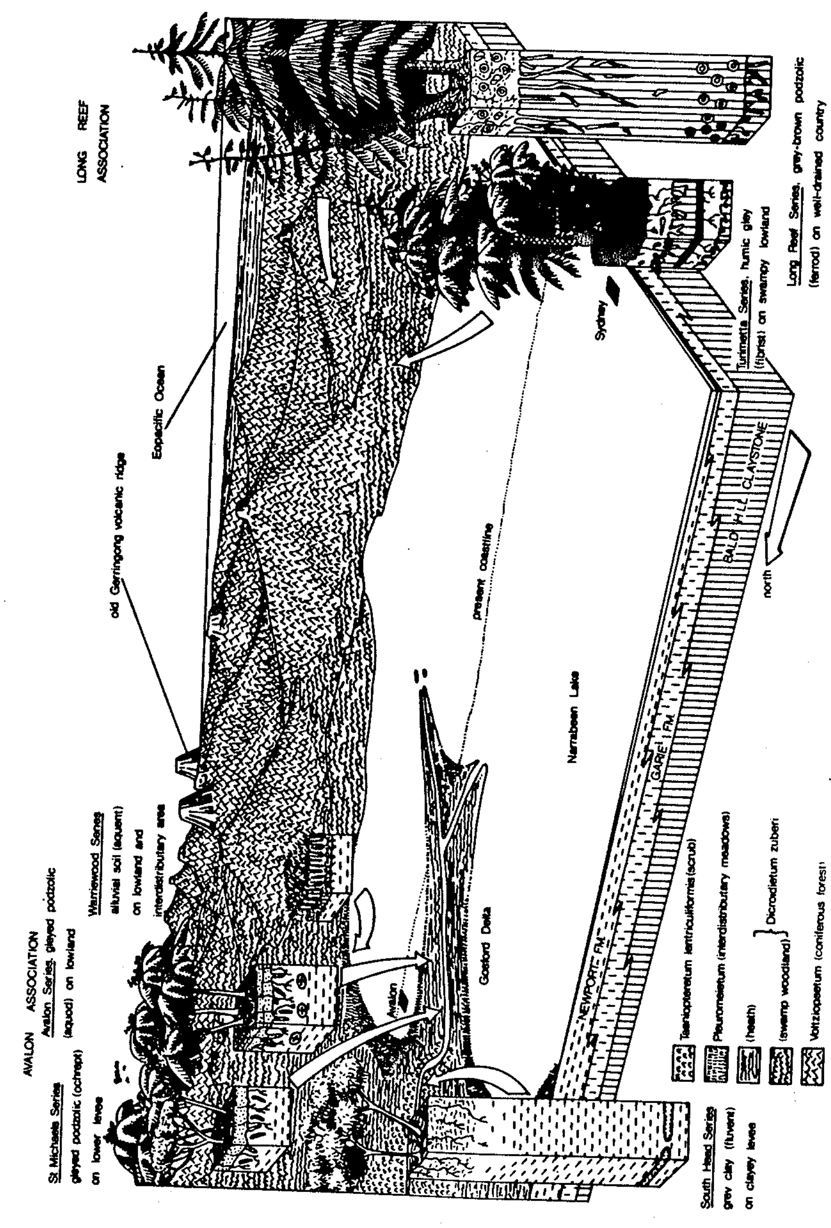

FIGURE 3.2. A reconstruction of the landscape, soils, and vegetation of the Sydney area, Australia during the later Early to earlier Middle Triassic. Lithological symbols as for Figure 3.11. (Redrawn from Retallack [1977b] with permission from the *Journal of the Geological Society of Australia.*)

Series paleosols of the Triassic near Sydney, Australia, the best evidence that siderite nodules were formed in place during soil formation was the rare occurrence of sand-filled insect burrows approaching the nodules from above, but sidling right around them, rather than passing through them (McDonnell 1974; Retallack 1977a). After detailed consideration of all their features, these Triassic paleosols proved impressively well preserved for their age and the few diagenetic alterations found were not critical to interpretation of the paleosols.

METHODS OF STUDY

Although theoretically recognized by all the features of modern soils, different methods of study are needed for older fossil soils than for Quaternary and modern soils. As a general rule, mineralogy and geochemistry, although necessary to consider, are the features of paleosols most susceptible to later alteration and should play a subordinate role to detailed field observations, the nature of horizons and soil peds, micromorphology, and associated fossil plants and animals in the study of older fossil soils.

Fieldwork

There is no substitute for detailed fieldwork in the study of fossil soils. The most striking feature of fossil soils from a distance is often their color, especially bright red and brown horizons. The most diagnostic feature is evidence of fossil roots in place. Other features include clayey, leached, massive prismatic, blocky-jointed or slickensided layers, and a variety of trace fossils, mottles, nodules, and concretions. Some concepts of paleosol horizons' parent material, and of special features, such as later fills of surficial holes left by felled trees (Fig. 3.3), need to be assessed in the field, before they can be confirmed by laboratory studies. Color should also be taken in the field, using a Munsell or other soil color chart, rather than any of the existing rock color charts.

For interpretative, as opposed to stratigraphical, studies of fossil soils, I have found that the soil mapping units of the United States Department of Agriculture (Soil Survey Staff 1951, 1962) are best for several reasons. The names do not imply anything of the nature or origin of the fossil soil and are not dependent on modern soil classification, whose criteria cannot always be applied to paleosols or unequivocally distinguished from diagenetic modifications. A separate name can be given to each particular paleosol. Part of the name relates it to other paleosols of a similar kind in the same area. The paleosols can be interpreted at several conceptual levels within a hierarchy of classification. There is no confusion between paleosols from different areas

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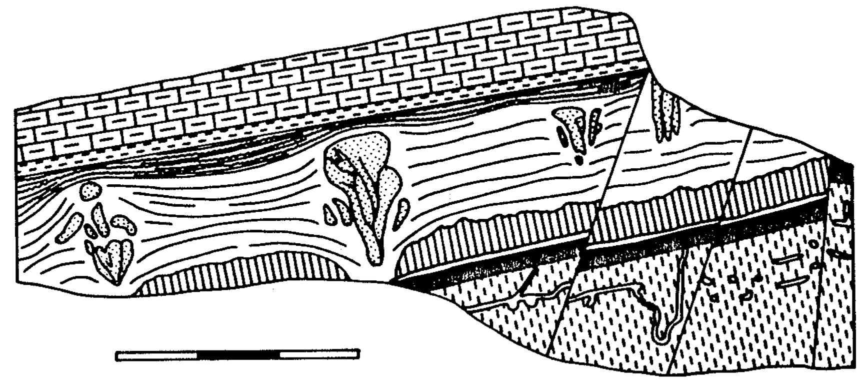

FIGURE 3.3. Scale drawing of a well-differentiated Early Permian paleosol at the contact between the Minnekahta Limestone (brick pattern) and the underlying Opeche Formation, near Boulder Park, Black Hills, South Dakota. The fossil soil is overlain by siltstone (horizontal dashes) and contains conspicuous cradle knolls (stipple). The profile includes an upper leached zone with thin ferruginized surfaces (wavy lines), yellow shale with relict bedding (long lines), red shale (vertical lines), and dark purple shale (black). Brick-red siltstone (vertical dashes) with numerous white mottles forms the parent material of the paleosol. Scale in meters, with no vertical exaggeration.

or rocks of different nature or age, as might arise if paleosols were named by letters or numbers, as is often a way of naming sedimentary facies (for example by Turner 1970; Miall 1977; Gorter 1978). Finally, the units of classification are already defined and widely accepted by soil scientists.

By this system, paleosols of a similar kind are grouped into a series named after a locality, such as the Avalon Series paleosols (of Retallack 1977b). The series is a most convenient unit for classification of paleosols in the field. Individual paleosols may be named by combining the series name with a term for the grainsize distribution in its A horizon, such as the Avalon silt loam paleosol. If individual paleosols need to be distinguished on bases other than the grainsize distribution in the A horizon, then variants and stages can be named from a variety of soil features, for example the Avalon silt loam egg-cup variant paleosol. On the other hand, series of paleosols can be grouped into associations on the basis of unifying characters, such as the Avalon Association paleosols.

Petrographical Thin Sections

Perhaps the most useful method of studying lithified older fossil soils is by petrographical thin sections. Microscopical features are often extremely helpful in deciding which features of the paleosol are original and which are due to later alteration. In addition, particular micromorphological features, such as sepic plasmic fabrics (Figs. 3.4, 3.5), may be diagnostic of soils and soil-forming processes. The terminology developed by Brewer (1964) for micromorphological features of soils has been widely accepted in studies of older fossil soils (Terrugi and Andreis 1971; Allen 1974b; Retallack 1977a; McPherson 1979).

Well-consolidated and cemented fossil soils may be difficult to disaggregate accurately for analysis of the proportions of sand, silt, and clay at different levels of the profile (Spalletti and Mazzoni 1978). Such quantitative information is of great value in assessing the nature of soil horizons and degree of illuviation, and also in naming paleosols. Grainsize distribution is best determined by counting measured grains under a microscope using a point counter. Friedman (1958, 1962) has shown that counts of the long axes of about 500 grains gives results very close to that of sieving fractions of unconsolidated sediments, and that even more accurate statistical parameters of the distribution can be obtained by converting the data with regression equations. Be aware, though, that the widely used Wentworth grainsize scale of geologists is not the same as that usually used by soil scientists. The grainsize scale and textural classes used by the United States Department of Agriculture (Soil Survey Staff 1975, p. 470) are better suited to textural studies of fossil soils.

FIGURE 3.4. Clino-trimasepic plasmic fabric from the A horizon of a well-differentiated paleosol in the latest Eocene to early Oligocene lowermost Chadron Formation (11.6 m in Figure 3.13); Pinnacles area, Badlands National Monument, South Dakota. × 50.

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FIGURE 3.5. Omnisepic plasmic fabric with numerous channel organo-sesquans (after roots) in the B horizon of a well-differentiated paleosol in the mid-Cretaceous upper Dakota Formation, in bluffs 10 m above the river bank north of the Saline River, near U.S. highway 281, north of Russell, Kansas. × 50.

Geochemistry and Mineralogy

As with modern soils, an understanding of the mineralogy and geochemistry of fossil soils is important to their interpretation, although not always diagnostic (Power 1969). Mineralogical determinations are usually based on elemental chemical analysis and on optical properties in petrographical thin sections. The ratios of heavy minerals separated from the fossil soil and its parent material may also be useful in indicating the degree of weathering. For example, zircon and tourmaline are more rapidly weathered than amphibole and pyroxene (Ruhe 1965). X-ray diffraction analysis of clay minerals is also very useful. Clay in the upper horizons of fossil soils commonly gives broader, more diffuse peaks on an x-ray diffractometer trace than clays lower in the same profile or in adjacent sediments. This is probably due to small crystallite size, poor crystallization and mixed-layering common in the upper horizons of soils (Retallack 1977a) and can be another useful indicator of weathering. Such weathered minerals may be more difficult to identify than fresh ones. The geochemical conditions for the stability and origin of each mineral phase must be carefully cross-checked for consistency and to determine which are due to original soil formation and which due to later alteration.

Fossils

Trace fossils, bones, shells, coprolites, plant fossils, or any other vestige of former life associated with fossil soils should be collected and prepared by appropriate techniques. Fortunately, a great deal of basic paleontological work has already been done on terrestrial and near marine organisms, and this may only need to be integrated with study of the fossil soil, in order to reconstruct the soil, its setting, and the ecosystem it supported.

COMMON DIAGENETIC MODIFICATIONS

Unraveling the effects of diagenesis and original weathering can be a major stumbling block in the interpretation of older fossil soils. It is helpful to be aware of diagenetic modifications common in older paleosols. The following is my own list of troublesome diagenetic modifications. Undoubtedly others will be added as research continues.

Reddening of Ferric Oxide Minerals

The diagenetic inversion of yellow and brown ferric gel and goethite to brown or red limonite and hematite causes an appreciable reddening of fossil soils. The reason why paleosols of the last interglacial are commonly redder than those presently forming in the same areas may be partly because of higher temperature and humidity when they formed (Ruhe 1965). However, it is probably also partly due to long-term diagenetic inversion to redder ferric oxide minerals (Walker 1974). In most pre-Tertiary paleosols with horizons stained with ferric oxide, this is mostly hematite. Even the B horizons of humic gley (fibrist) paleosols from the Triassic near Sydney, Australia (Retallack 1977b), are brightly colored. These fossil soils may not have originally been as red as they appear today. Such paleosols may have originally been a variety of pale yellowish, brownish or pinkish colors, and were not necessarily oxisols or lateritic podzolic soils.

Siderite Pseudogley

Siderite nodules may form in waterlogged portions of modern soils (Kanno 1962; Degens 1965) and were evidently an original feature of the B horizons of gleyed podzolic (aquod) paleosols from the Triassic near Sydney, Australia (Retallack 1977a, 1977b). However, even previously

1967; Hardan 1971; Conry and Mitchell 1971). More could be done, particularly with older fossil soils associated with Miocene, Pliocene, and Pleistocene hominoid fossils. Spectacular new finds of such fossils are changing concepts of our own evolution as a species (Johanson and Taieb 1976; Leakey et al. 1976; Bishop 1978), but many questions remain. How and when was the evolution of *Homo sapiens* related to forest, savanna, and prairie environments? To what extent can the spread of grasslands be attributed to the use of fire by hominoids? Did the giant beasts of the Pleistocene become extinct because of overkill by hominoids, because of their effect on the ancient environments or because of other environmental changes? Detailed studies of fossil soils addressing these questions have not yet been forthcoming, but enough is now known of the geological occurrence of early hominoid fossils to indicate their potential. In the Middle Silts and Gravels Member of the Kapthurin Formation of Kenya, between 700,000 and 230,000 years old, Tallon (1978) reported hominoid remains (probably *Homo erectus*) and stone artifacts scattered over the surface of a fossil soil, which had a calcrete 2 m below the surface. This occupation site was evidently on lakeside flats just south of the nose of a trachyte flow, which was used to quarry the artifacts. Further studies are needed to establish the nature of the calcrete and the different kinds of lakeside vegetation and soils. A variety of fossil soils have also been found in association with hominoid remains, perhaps up to 2 million years old, in the Chesowanga area of the northern Rift Valley of Kenya (Bishop, Hill, and Pickford 1978), and also in association with hominoid remains 9 to 12 million years old (mid-Miocene) in the central Kenyan Rift Valley (Pickford 1978). Paleosols may also be useful in evaluating the habitats of Miocene hominoids from the Siwalik deposits of Pakistan (Pilbeam et al. 1977a, 1977b; Behrensmeyer, personal communication 1979).

CONCLUSIONS

The study of fossil soils is just beginning. Compared to other branches of earth sciences there are still few researchers, although their ranks are growing. There are innumerable projects unattempted, many involving major aspects of the evolution of terrestrial ecosystems. Undoubtedly, more will be revealed as work progresses.

Although useful in stratigraphical mapping, fossil soils also provide evidence for interpretation of ancient terrestrial environments. These interpretations are particularly effective when integrated with existing paleontological and geological studies. Such an approach promises to become an important additional way of understanding the past.

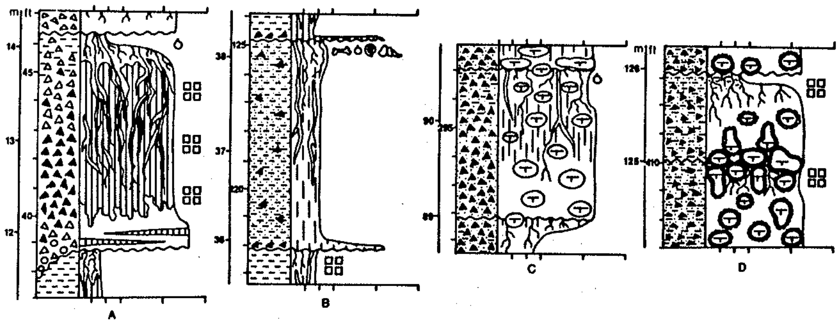

FIGURE 3.13. Oligocene paleosols of the Pinnacles area of the Badlands National Monument, South Dakota: **A,** well-differentiated paleosol of the Chadron Formation. **B,** weakly differentiated paleosol, with sparse gray mottles, in the Scenic Member of the Brule Formation. **C,** weakly differentiated caliche-bearing paleosol, with sparse gray mottles, in the Poleslide Member of the Brule Formation. **D,** weakly differentiated caliche-bearing paleosol, without larger gray mottles, in the Sharps Formation. Scale measurements refer to a measured section in the Pinnacles area in which the uppermost non-redeposited Cretaceous marine rocks are at about 8.3 m. Lithological key as for Figure 3.11.

The Impact of Humans on the World's Land Surfaces

Even if the species *Homo sapiens* soon becomes a diagnostic fossil of one of the briefest biostratigraphical zones in geological history, its effect on soils of the world is already conspicuous and irreversible. Modern cities, dams, parking lots, and highways are reshaping the landscape. Plowing, chemical fertilizers, forestry, irrigation, flood mitigation works, waste disposal, acid rain, and introduced animals and plants are altering the nature of soils, usually to our detriment. These problems are not new. More than 2,300 years ago, Plato lamented the spread of rocky, barren wastes in the Greece of his day, compared to the fertile soils that had been cleared of forest and cultivated by his ancestors (Glacken 1956). About 4,000 years ago much of the mighty cedar forests of Lebanon had been cut for Phoenician ships, towns, and export, and farmers terraced the steep cultivated slopes in an unsuccessful attempt to prevent more serious soil erosion (Lowdermilk 1943). Until about 5,000 years ago parts of the Sahara Desert were far less forbidding wastes than they have been since. Neolithic nomadic tribesmen have left numerous rock drawings which indicate that elephant, rhinoceros, lion, panther, giraffe, antelope, cattle, ram, and ostrich once lived there (Huzayyin 1956).

Much has been learned about the effects of ancient clearing and civilizations through the study of Quarternary fossil soils (see, e.g., Ložek

nonwaterlogged soils become waterlogged when they subside below the water table. Such early diagenetic gleization is the best explanation for siderite nodules, spherulites, crystal tubes, and replacement of the walls of earthworm burrows in the A and organic horizons of others of the Triassic paleosols mentioned. Diagenetic gleization (or pseudogley of Roeschmann 1971) may also be responsible for the greenish and bluish hue of many older pedoliths and unoxidized soil horizons interbedded with redbeds. There is no evidence from the Triassic study cited that diagenetic gleization proceeded to the extent of reducing original ferric oxide minerals. This is in agreement with other studies indicating that ferric oxide minerals, once formed, are highly insoluble and unlikely to be reduced by the action of normal groundwater alone (Eaton 1942; Millot 1970).

Clay Diagenesis

Clay minerals are notoriously susceptible to diagenetic alterations of such a scope and complexity (Millot 1970) that it is very difficult to be certain of the original clay mineralogy of older paleosols. In Carboniferous clayey paleosols of the Ruhr district of West Germany, chlorite has been regarded as a diagenetic alteration product of montmorillonite, vermiculite and biotite, and kaolinite has been regarded as an alteration product of illite and muscovite (Roeschmann 1971).

Silicification

A variety of cherty rocks, cemented by silica, are found in or were formed by ancient soils. The silcretes of Australia, South Africa, the United States, and Europe are the most problematic of these. The cementing silica as well as the cemented orthoquartzites of silcretes may be derived from the weathering of stable land surfaces over long periods of geological time. In some cases it seems that the silica was transported great distances from where it was leached out of a variety of soils, so it is not necessarily genetically related to the material which it has transformed into a massive flinty rock. This process is probably at least partly diagenetic, although not yet completely understood. Different views on the nature and origin of silcretes have been conveniently collected in a volume edited by Langford-Smith (1978).

Ganister is a coal miner's term for indurated silicified sandstone, no thicker than 1.5 m, with over 90 percent angular grains of quartz in the grainsize range 0.5 to 0.15 mm. They commonly contain fossil roots and underlie coal seams in the Carboniferous coal measures of England (Williamson 1967). These are evidently the upper horizons of fossil soils and could not have supported the vegetation, indicated by the fossil roots, in their present indurated state. From my study of Triassic ganisters from near Sydney, Australia (Retallack 1977a), I concluded that the silica cement was derived largely from the early diagenetic mobilization and reprecipitation of opal phytoliths from plants, fecal pellets, and airborne dust in the original soil.

Similarly, the induration and complete silicification of petrified peats is probably also partly diagenetic. Kidston and Lang (1921) regarded the dead areas, wound reactions, and unequal enlargement of cells in otherwise well-preserved stems of *Rhynia major* from the early Devonian Rhynie chert of Scotland, as responses of living plants to the infiltration of silica-rich groundwaters from nearby fumaroles. Substantial influx of silica into this petrified peat was probably coincident with more aquatic conditions which destroyed the *Rhynia* marshes. The peaty substrate with well-preserved remains of these plants in growth position is overlain by layers with abundant crustaceans, algae, and the more aquatic vascular plant *Horneophyton* (emended from the original name of Kidston and Lang, by Barghoorn and Darrah [1938]). The living plants probably did not live in extremely high silica concentrations, nor in the indurated chert of today. The chert must have been silicified and indurated very early in diagenesis. This is apparent from the exceptional preservation of uncrushed herbaceous remains, and anatomical detail, in this and other petrified peats of various geological ages (Schopf 1970; Ting 1972; Basinger and Rothwell 1977; Runnegar 1977).

Sharpened Boundaries

The delineation of nodules, concretions, and horizons is often much sharper in older fossil soils than in modern soils. This may be in part a diagenetic segregation of chemically incompatible parts of the paleosol. Particularly noticeable is the often sharp delineation of gray reduced areas around fossil roots within horizons stained red with ferric oxide minerals, in older paleosols (Fig. 3.6).

The superficial appearance of sharp contacts can also be due to differential weathering of more indurated portions of a paleosol. In the Avalon and Warriewood Series paleosols of the Triassic of the Sydney Basin (Retallack 1977a), the lower boundaries of the silicified A horizons (ganisters), never prove to be as distinct in polished slabs as they appear in the field.

Physical Compaction

Depending on the depth and other conditions of burial, fossil soils may be flattened or develop jointing or other structures. According to Roesch-

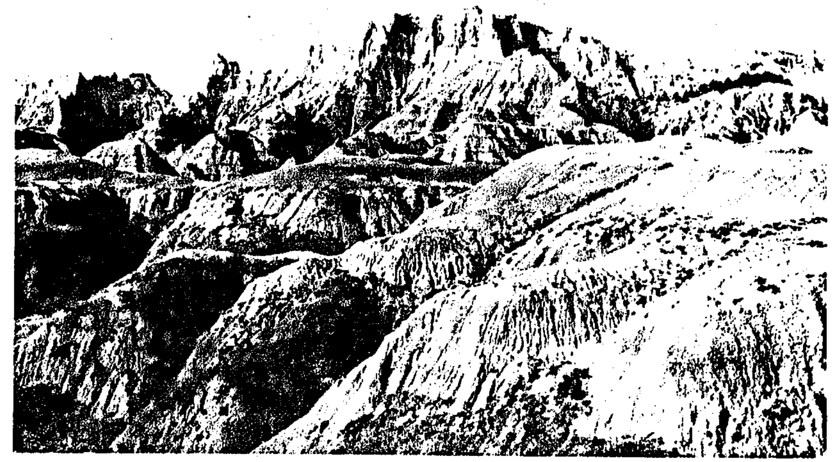

FIGURE 3.12. The Badlands National Monument of South Dakota, near the Pinnacles, an area with about 100 m of topographical relief. The late Cretaceous marine rocks (runneled lower slope), are capped by thick well-differentiated paleosols (light and dark bands), overlain by the latest Eocene to early Oligocene Chadron Formation (smooth gentle slopes), mid-Oligocene Scenic Member of the Brule Formation (light and dark banded unit in lower cliffs), mid to late Oligocene Poleslide Member of the Brule Formation (less strongly banded unit), and late Oligocene Sharps Formation (light colored rocks of far mountaintop).

resorted ash, has incised 10 m into floodplain claystones. In this area of the Badlands, it is apparent that small tracts of prairie emerged during the later Oligocene as the grassy portions of savanna expanded at the expense of trees. Increasing aridity, lowered water tables, and coarser ashy substrate appear to have accompanied these changes. Fire was probably also important, although not in evidence.

Probable savanna or pampas paleosols have also been reported by Andreis (1972) and Spalletti and Mazzoni (1978) from the Eocene to Oligocene Sarmiento Group in Chubut Province, Argentina. These paleosols have many characteristic soil microstructures and are associated with numerous fossil boli of dung beetles, diverse fossil mammal remains, and phytoliths (cited as evidence of grasses).

Continuing studies of fossil soils such as these may provide evidence for the origin of a number of different kinds of plant communities which are poorly represented in the fossil record, such as savanna, prairie, desert, and chaparral vegetation.

and nature of the early savanna and prairie communities and also of the coevolution of mammals and plants in these environments. David L. Dilcher and I have recently initiated a project with this exact aim, based on the Oligocene succession exposed in the Badlands National Monument of South Dakota. This is one of the richest fossiliferous areas in the world for the remains of extinct mammals, whose teeth and limbs appear adapted initially to forest and woodland and later to more open savanna or prairie conditions (Webb 1977). Apart from endocarps of hackberry (*Celtis*), possibly cached by rodents, almost all the leaves, trunks, and pollen of this vegetation appear to have decayed away long ago (Clark, Beerbower, and Kietkze 1967; Bjork and Leopold, personal communication). Our preliminary mapping in the area indicates that there are at least 87 successive fossil soils in the succession exposed in the Pinnacles area of the Badlands National Monument (Fig. 3.12). Laboratory examination of these is not complete, but some conclusions can be made from field observations. The thick fossil soil developed on the erosional unconformity between the latest Eocene to Oligocene Chadron Formation and latest Cretaceous marine rocks (Wanless 1923; Pettyjohn 1966) and most of the paleosols excavated in the Chadron Formation, have leached A horizons and reddish B horizons and numerous thick and deep root mottles (Fig. 3.13A). These fossil soils probably developed under forests and woodlands. In the overlying Scenic Member of the Brule Formation, such soils are only found in close association with sandy levee and channel deposits, indicating that gallery woodlands lined watercourses. Fossil soils of the floodplain at this level have much sparser large root mottles, abundant fine rootlets, and less pronounced B horizons (Fig. 3.13B), and probably were formed under savanna vegetation. Similar paleosols have also been recognized in the Orella Member of the Brule Formation in nearby Nebraska (Schultz, Tanner, and Harvey 1955; Schultz and Stout 1955; Harvey 1960). The whitewashed appearance of outcrops of the Poleslide Member of the Brule Formation, higher in the Badlands succession, is largely due to abundant carbonate nodules. Although there is some overlap of these from higher into lower paleosols, these nodules are most common in the lower portions of the paleosols (Fig. 3.13C). Some of the rare reduced root mottles of these paleosols have also been found in the nodules, and are evidence that they are caliche nodules in place. Caliche is also common in many younger Tertiary formations of the Great Plains of North America (Swineford, Leonard, and Frye 1958; Reeves 1970). It probably indicates increasingly warm and semiarid climate. Higher in the Poleslide Member there are some fossil soils of this kind in which I have not seen any of the large reduced root mottles over considerable lengths of these extensive outcrops (as in Fig. 3.13D). At the level of the Rockyford Ash Member of the Sharps Formation, a thick stream channel deposit, largely of

mann (1971), the gleyed paleosols underlying Carboniferous coals of the Ruhr district of West Germany were compacted from 25 percent in sandy to 70 percent in clayey sediments. Such values need to be derived independently for each particular case, using deformation of structures and comparing the bulk density of the fossil soil with that of analogous modern soils.

Slickensides and prismatic jointing can form both at the surface in a soil, and during deep burial. The nature of the bounding surfaces (cutans) in petrographical thin sections is the best way of discriminating whether they were formed in the original soil or during later diagenesis. On this basis, the tessellated pavements and prismatic jointing in massive B horizons of the Long Reef Series paleosols from the Triassic near Sydney, Australia appear to be diagenetic in origin (Retallack 1977a).

Copper, Uranium, and Vanadium Mineralization

Certain metals are commonly mobilized and reprecipitated in fossil soils during diagenesis. Within the Long Reef Series paleosols mentioned previously, paratacamite, atacamite, and rare native copper may fill the cleat of coalified stick debris and the central portion of gray root mottles.

The uranium and vanadium ores of the Chinle and Morrison Formations of the western United States may be similar. Most of the mottled beds and redbeds of these units which I have seen are fossil soils, although not generally recognized as such. Many of the fossil logs in these formations, sometimes heavily mineralized, are scattered about on the surface of these fossil soils. Some of the impervious layers which confined ore-bearing

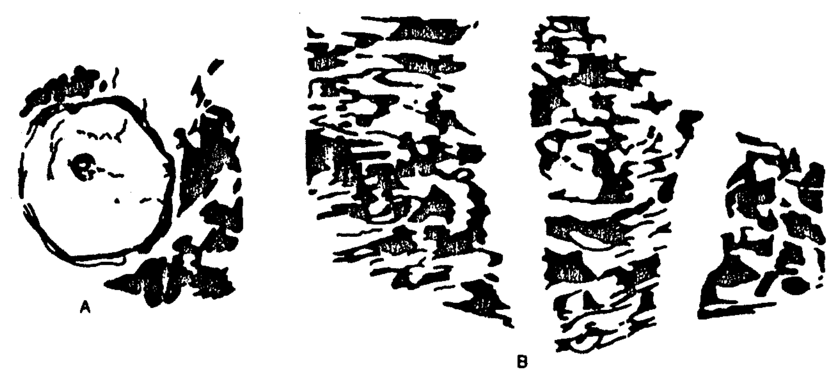

FIGURE 3.6. Ferric mottles (black) outlining gray mottles after roots, from the later Early to earlier Middle Triassic, Turimetta clay slightly eroded phase paleosol, north of Sydney, Australia: **A**, horizontal section including a root with clay-filled cortex, separating carbonaceous remnants of woody stele and periderm. **B**, vertical section. Natural size. (Redrawn from Retallack [1977a] with permission from the *Journal of the Geological Society of Australia*.)

solutions and also some of the organic layers which have been mineralized, are also probably fossil soils. The minerals are thought to have been dissolved in groundwater draining from uranium-bearing crystalline or volcanic rocks in uplands, and precipitated at a "redox front" (a transitional zone from positive to negative Eh or oxidizing to reducing conditions) in the lowlands (Stanton 1972; Granger and Warren 1978). This may have occurred during the formation of the fossil soils in these deposits, as well as during their later diagenesis. Basin-wide studies of these fossil soils, and particularly interpretations of their ancient water tables, may prove to be a powerful exploration tool. To my knowledge, this has not yet been suggested or attempted.

INTERPRETATION

Some aspects of ancient terrestrial environments can only be interpreted from the study of fossil soils. For other aspects, the study of fossil soils may be a valuable independent check on conclusions reached by other kinds of research. The study of fossil soils has lagged far behind other geological and paleontological studies, such as analysis of sedimentary basins, sedimentary petrography, heavy mineralogy, sedimentology, paleocurrents, palynology, paleobotany, paleontology, and paleoecology. In areas where these other studies have already been completed, a study of fossil soils may serve as a critical focus to integrate other information into a more detailed concept of a particular ancient terrestrial ecosystem than was hitherto possible.

Water Table

The nature and depth to the B horizon of some kinds of paleosols is a reasonable guide to the depth of the water table, or at least the zone about which it most commonly fluctuated. Frequent wet periods are indicated by siderite nodules or spherulites in the B horizon (as with all these features they must be demonstrated to be original), by asepic or undulic plasmic fabrics, by an apedal or massive structure, by shallow root systems with thicker roots spreading laterally rather than downward, and also by more humified organic matter at the surface. More frequent dry periods may be indicated by reddish mottled or nodular B horizons stained with ferric oxide minerals, by well-differentiated peds, by animal burrows, by deeper root systems, and by less humified organic matter at the surface. Even better drained and more arid conditions are indicated by soils without a clear B horizon or with other indicators, such as caliche nodules. It is also possible that this last kind of paleosol was so immature before burial that there was

on a pre-Oligocene surface in a deep drill hole in the Ross Sea, Antarctica (Ford and Barrett 1975).

Forests would have had a greater stabilizing effect on the landscape than preexisting kinds of vegetation. Even today, forests have not achieved total cover of the landscape, but they were probably much less effective in controlling upland erosion during the Devonian and Carboniferous. Schumm (1968) has pointed out that the Euramerican Carboniferous coal measures have more and thicker clastic partings than the spectacularly thick early Tertiary coals of North America, Germany, and southeastern Australia. He suggested that this is due to greater erosion of less vegetated uplands during the Paleozoic. Further study of woodland and other fossil soils in sedimentary basins should refine this hypothesis considerably.

Tertiary Emergence of Grassland Soils

The development of savanna, steppe, prairie, and pampas vegetation on the plains, and of grassy vegetation above the snowline in alpine regions, also had a major effect on landscapes of the world. On the plains, new kinds of soils formed under the grassy swards, mollisol (of the U.S. Dept. of Agriculture), or chernozem, prairie soil, black earth, rendzina, and wiesenboden of older nomenclature. Evidence from fossil plants indicates that monocotyledonous angiosperms probably evolved during the early Cretaceous (Doyle 1973), but true grasses do not appear until the early Tertiary (Litke 1968). The best known megafossil record of grasses is in the Miocene to present sediments of the Great Plains of North America (Elias 1942; Thomassen 1979). Other than this the fossil record of grasses is generally poor. The emergence of grasslands within the interior of all the major continents at various times during the Tertiary is better indicated by other fossil plant remains and fossil mammals. In particular, the high crowned (hypsodont) teeth of grazing mammals indicate that coarse grassy fodder was widely available. On such grounds, savanna or pampas may have appeared in Argentina as early as the Eocene (Patterson and Pascual 1972). In Africa, there were probably considerable areas under savanna vegetation during the Oligocene, and it was probably more extensive during the Miocene and Pliocene (Tanner 1978; Axelrod and Raven 1978). In North America, savanna began to emerge as a vegetation type during the Oligocene, with steadily decreasing numbers of trees culminating in prairies as extensive as those of today by Pliocene times (Webb 1977). Savanna and steppe of central Asia appears to have become widespread during the Miocene, expanding into western Europe and China by the Late Miocene (Osborn 1910). In Australia, savanna may have been present during the Pliocene, but there is little evidence of savanna and grassland until the Pleistocene (Kemp 1978; Martin 1978).

Evidence from fossil soils is likely to give a much clearer idea of the age and Triassic of the U.S.S.R. from the southwestern flanks of the Ural Mountains south to the Donetz Basin (Danilov 1968; Chalyshev 1969); from the early Cretaceous of England and France (Allen 1959; 1976; Batten 1973; Meyer 1976); from the late Cretaceous of Mexico (McBride 1974); and the Eocene and Miocene of South Carolina (Johnson and Heron 1965). Laterites may be formed in several ways, but are apparently initiated as an horizon of deep forested soils (McFarlane 1976). Such soils were evidently widespread during the Cretaceous and early Tertiary (Phillobos and Hassan 1975; Singer 1975; Abbott, Minch, and Peterson 1976; Thompson, Fields, and Alt 1977; Nilsen 1978; Nilsen and Kerr 1978; Blank 1978, McGowran 1979; Cox 1979). The so-called "karst bauxites" may be genetically similar to laterites and have been extensively reviewed by Nicholas and Bildgen (1979).

Organic soils (or histosols of the U.S. Dept. of Agriculture) are evidently more ancient than trees. The early Devonian Rhynie chert of Scotland is a petrified peat with remains of vascular land plants in growth position (Kidston and Lang 1921). However, the appearance of swamp forests and woodlands was probably a great stimulus to the development of thick organic horizons in wetland soils. The paleosols of the Euramerican Carboniferous coal measures have been given a number of nongenetic names, such as underclay, seat earth, fire clay, tonstein, and ganister (each explained by Williamson 1967). These fossil soils are limited in variety. Many of them were forested and most of them more or less gleyed (Huddle and Patterson 1961; Roeschmann 1971; Feofilova 1977).

Thick organic horizons and oxidized forest soils were additions to an expanding array of soils forming on the earth after the Devonian, but more ancient kinds of soils continued to form on other parts of the landscape. Caliche-bearing paleosols have been found in Permian and Triassic rocks of Scotland (Steel 1974; Watts 1976, 1978); in the Late Triassic and Jurassic of Connecticut and Massachusetts (Hubert 1977a, 1977b); in the late Cretaceous and early Tertiary of France (Freytet 1971, 1973); in the early Tertiary of France, England, Belgium, and the Netherlands (Buurman 1975); and the early Tertiary of California (Peterson and Abbott 1979). Fossil soils developed on limestone with karst topography have been found in the Middle Devonian of western Canada (Maiklem 1971; Wardlaw and Reinson 1971); in the Carboniferous of England (Walkden 1974), Kentucky (Walls, Harris, and Nunan 1975), and Missouri (Keller, Wescott, and Bledsoe 1954); in the Permian of New Mexico (Dunham 1969; Estaban and Pray 1977); in the Triassic of Italy (Bosellini and Rossi 1974); in the Jurassic of Italy (Bernoulli and Wagner 1971); in the Cretaceous of Greece (Faugeres and Robert 1969); in the Cretaceous and early Tertiary of France (Freytet 1971); and in the Eocene of Spain (Estaban 1972). Weathered zones at unconformities have also been reported on pre-Pennsylvanian surfaces in Colorado (Hubert 1960; Power 1969); within the Permian of New South Wales (Loughnan 1975); within the Jurassic of Italy (Folk and McBride 1976); and

not enough time to form a B horizon. If this is the case, then sedimentary relicts and other indications of immaturity should also be in evidence.

Soil Chemistry

Assessing the original chemistry of a fossil soil is often a major obstacle to interpretation. Features such as base saturation, cation exchange capacity, and pH are critical to parts of classifications of modern soils. These chemical features are irreparably changed upon burial, and further changed during compaction, diagenesis, and metamorphism. For example, buried Quarternary soils in Iowa are often completely saturated with bases throughout the profile, but analogous modern soils are only 40 percent saturated in the B horizon, but completely saturated at depths of about 1.5 m in the C horizon. The pH is usually over 6.0 throughout the buried soils, but in analogous modern soils it is 5.3 in the A horizon, 4.0 in the B, and 5.0 in the C horizon (Ruhe 1965).

These chemical features are better assessed from the mineralogical phases thought to have been in the original soil. The original pH or Eh can be assessed from the ranges of values in which these minerals form or are stable, as outlined by Krumbein and Garrels (1952) and Baas-Becking, Kaplan, and Moore (1960). Acidic conditions (neutral to low pH) are indicated by kaolinite, by ganister or other siliceous A horizon without carbonate, and by red ferric oxide mottles, concretions, nodules, surface crust, and diffusion ferrans. Alkaline conditions (neutral to high pH) are indicated by carbonate (including siderite nodules) and by other features indicated by Northcote and Skene (1972). Oxidizing conditions (positive Eh) are indicated by red ferric oxide mottles, concretions, nodules, surface crust, and diffusion ferrans. Reducing conditions (negative Eh) are indicated by gley colors, by pyrite framboids, and by siderite nodules, intercalary crystals, spherulites, and crystal tubes. Sodic and saline soils (with Na^+ as the dominant exchangeable cation) may be characterized by surface salt crusts or casts of such crystals in sediment, by prominent domed columnar peds in an argillic B horizon or by other features indicated by Northcote and Skene (1972). Additional investigations into these and other indicators of original soil chemistry would be of great value for identifying fossil soils.

Soil Fauna

The variety of traces of soil fauna discovered in fossil soils to date include Triassic burrows of earthworms, cicada-like insects and large vertebrates (Retallack 1977a), Eocene dung beetle boli (Andreis 1972), and Oligocene land snails and coprolites (Wanless 1923; personal observations). The spectacular vertical corkscrew-like burrows (called *Daemonelix*) often more than 10 m deep in the early Miocene Harrison Formation of northwestern Nebraska (Fig. 3.7) have even been found with entombed skeletons of burrowing rodents. The *Daemonelix* burrows indicate that the water table in these soils was seldom closer than 10 m below the surface, for otherwise the rodents would have drowned. The nature and occurrence of traces of soil fauna can supply useful constraints in the interpretation of fossil soils. Further studies of these traces can be expected to reveal much about the evolution of such animals as earthworms and ground-dwelling social insects and rodents.

Vegetation

The former vegetation of some fossil soils, particularly gleyed and organic soils, may be preserved in or around their upper surface. Plant material is usually not preserved in more oxidized fossil soils. However, poorly preserved pollen in oxidized Triassic fossil soils proved sufficient to gain an idea of its vegetation (Retallack 1977b). As a general rule, fossil plants are much better preserved in flood, pond, or lake deposits than in direct association with fossil soils. More detailed understanding of fossil plants gained from such better preserved material is often useful also in interpreting fossil soils. The fossil plants may indicate whether the vegetation was a forest, savanna or grassland, and also show adaptations to conditions such as aridity, salinity, or waterlogging. This may serve as an independent check on deductions from their likely fossil soil. The study of fossil soils and plants can be combined to gain a better understanding of both.

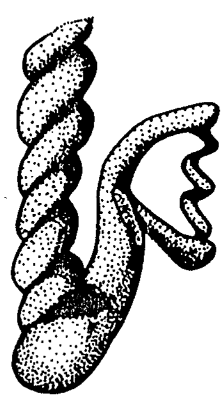

FIGURE 3.7. *Daemonelix*, burrow system of a rodent from the early Miocene Harrison Formation of Nebraska. Reduced one-tenth natural size. (Redrawn from Barbour [1897].)

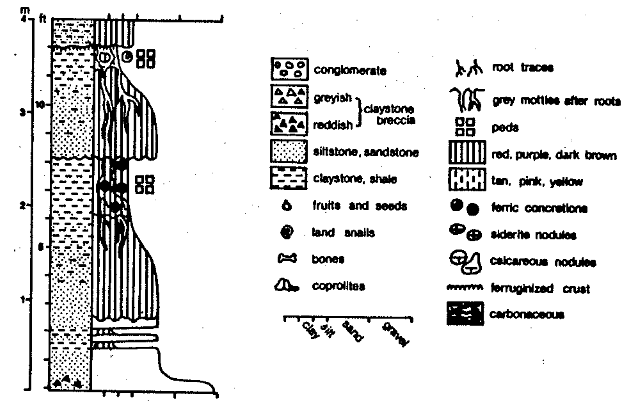

FIGURE 3.11. Later Early or earlier Middle Triassic type Long Reef clay paleosol north of Sydney, Australia. The ferruginized lower B horizon of the type profile includes pedorelicts of the A horizon of an underlying paleosol.

soft volcanic sand grains better preserved lower in the profile. The reddish material with isotic fabric in lower horizons is mainly hematite and goethite, with some siderite. Soil-forming processes have resulted in considerable accumulation of total iron (Fe_2O_3 + FeO) in the B horizon, a moderate accumulation of Al_2O_3 in the A and B horizon, and loss of CaO, MgO, Na_2O, and K_2O throughout the profile. These paleosols were identified as gray brown podzolic (of Stace et al. 1968), ferrods (of Soil Survey Staff 1975), and Uf2 (of Northcote 1974).

I have seen similar fossil soils, not yet studied in detail, in the late Mississippian or early Pennsylvanian Manning Canyon Formation, across Utah Lake, from Provo, Utah; in the early Permian uppermost Opeche Formation, near Boulder Park, in the Black Hills of South Dakota (Fig. 3.3); in the late Triassic Lockatong Formation of the Newark Basin, near Phoenixville, Pennsylvania; in the Late Triassic Chinle Formation of the Petrified Forest National Monument, Arizona; in the early Cretaceous Otway Group of southern Victoria, southeastern Australia; in the mid-Cretaceous upper Dakota Formation in Russell County, Kansas; and in the Eocene Clarno Formation, on Camp Hancock, near Clarno, Oregon. Other aspects of the geology and paleontology of these various places are discussed by Tidwell (1967), Tranter and Petter (1963), Olsen (1978), Wycoff et al. (1972), Douglas (1969), Hattin, Seimers, and Stewart (1978), Baldwin (1976), and references therein. Similar well-differentiated reddish paleosols have also been reported from the Pennsylvanian of Colorado (Hubert 1960); the Permian and Triassic of southern Germany (Ortlam 1971); the Permian accentuated soil-forming processes associated with preexisting vegetation of lesser stature. With woodlands and forests appeared the first well-differentiated soils, such as podzols (of conventional terminology) or spodosols, ultisols, and alfisols (of the U.S. Dept. of Agriculture classification, Soil Survey Staff 1975). These are soils with two clearly recognizable layers, an upper or A horizon leached of materials such as iron, aluminum, and clay, and a lower or B horizon in which these materials accumulate. Further study of such mid-Paleozoic paleosols may clarify how, when, and where the first woodlands and forests arose.

Theoretically this could have happened during the mid-Devonian. At this time there were many plants with limited amounts of secondary wood (Banks 1968). Large, poorly preserved stumps of *Eospermatopteris* have been found in later Middle Devonian (Givetian) rocks of New York State (Banks 1966). The more massive progymnosperm wood, *Callixylon*, may also be as old as the late Givetian (Banks 1968). During the Late Devonian this wood was widespread, and individual trees grew up to 1.6 m in diameter (Beck 1971). This is not to say that these trees necessarily formed woodland communities, although this is likely. Among outcrops of Late Devonian rocks in the area of New York and Pennsylvania described by Woodrow and Fletcher (1967), I recently discovered a number of intriguing, well-differentiated paleosols during preliminary fieldwork with J. D. Grierson. This would be a particularly suitable area for a study of fossil soils and plant paleoecology, as the fossil plants from here are exceptionally well understood (as reviewed by Banks 1966, with continuing contributions by Skog and Banks 1973; Grierson 1976; Bonamo 1977).

Paleosols with distinct leached A horizons and reddish B horizons are common and widespread in Carboniferous and younger terrestrial sedimentary rocks. As an example, the type Long Reef clay paleosol (Fig. 3.11) evidently formed under coniferous forest on volcanogenic sandstone in moderately well-drained rolling land near Sydney, Australia during the earlier Triassic (Retallack 1977a, 1977b). This has a thin brownish yellow (Munsell color 9YR6/8) surface crust (A_1), overlying 33.5 cm of very coarse subangular blocky and slickensided greenish gray (8G6/1) claystone with vomasepic porphyroskelic fabric. This passes gradationally down into 20 cm of weak red (1YR4/2) fine clayey sandstone (B_{1ir}) with isotic porphyroskelic fabric, penetrated by numerous light greenish gray (8GY7/1) vermicular mottles of sandy material around old root channels. Deeper in the B horizon there is 41 cm of apedal red material with only rare gray mottles (B_{2ir}) and below that, 67 cm of dusky red (3YR3/2) claystone (B_{3ir}). This lowest level contains relict pedotubules and undulic porphyroskelic fabric of the A horizon of an underlying paleosol, the rest of which forms a C horizon to the type Long Reef clay paleosol, more than 181.5 cm below its preserved surface. Clays of the paleosol are mainly kaolinite, probably weathered from

In relating occurrences of well-preserved fossil plants to fossil soils I make one basic assumption. As in the modern world, different kinds of fossil soils probably supported different communities of plants, and the same soils supported similar vegetation. Some quantitative approaches have been devised for reconstructing Carboniferous plant communities from their dispersed compression remains (Scott 1977, 1978, 1979) and from their representation in peats petrified in calcareous coal balls (Phillips et al. 1974). In my own studies of Triassic fossil soils and plants of eastern Australasia (Retallack 1977a, 1977b, 1977c, 1978), I found that interpretation of communities from named recurrent fossil associations of plants gave a more balanced appraisal of different kinds of vegetation. This largely qualitative approach was particularly effective when the named associations were based on specific collections of fossil plants which did not appear to have been transported far, and also when combined with studies of fossil soils and depositional sedimentary environments. The different kinds of fossil plant associations and fossil soils can then be matched according to their nature, or preferably by direct correlation with plant fragments associated with the fossil soil.

Type of Soil

If a fossil soil can be identified with a modern soil, then presumably it formed under similar conditions of topography, drainage, vegetation, and climate. Even if complete identification is not possible because of later alteration or lack of diagnostic features, comparison with modern soils may still provide useful information about the ancient environment. Interpretations of fossil soils based on this principle of uniformitarianism have been presented by many authors (Ložek 1967; Ruhe 1970; Allen 1974b; Retallack 1977b).

There is still some debate concerning the classification of modern soils. Some of these uncertainties should be considered when interpreting fossil soils, but ultimate identification of a fossil soil in a classification of modern soils rests on a literal interpretation of a cited authority. The classification of the United States Department of Agriculture (Soil Survey Staff 1975), the more conventional Australian classification (Stace et al. 1968), and the nongenetic classification of Northcote (1974) represent a range of viewpoints on soil classification. The so-called "ecological method" of Duchafour (1978) also shows much promise, particularly if expanded into a more comprehensive form.

Fossil soils should not be strained to fit into a classification of modern soils, as some kinds of soils once formed on the earth are now extinct. The only extinct kinds of paleosol reported to date are the Precambrian paleosols thought to have formed in an anoxic atmosphere (Roscoe 1968;

Frarey and Roscoe 1970; Rankama 1955). Other more subtle kinds of extinct paleosols may be discovered in the future.

Some kinds of fossil soil horizons such as caliche and laterite have been widely identified in the past on too little evidence, without any indication of the nature of the profiles or any attempt to demonstrate that the features observed were original rather than diagenetic. This is especially apparent from reexamination of so-called laterites of the area around Sydney, Australia by Hunt, Mitchell, and Paton (1977). Close attention to detail is necessary if studies of fossil soils are to realize their evident potential.

Basin Tectonics

The nature of fossil soils and their distribution within sedimentary sequences may give otherwise unattainable information on rates of sedimentation, subsidence, uplift, and basin topography. In my work on Triassic paleosols north of Sydney, Australia (Retallack 1977a, 1977b), the distribution of paleosols indicated all these things. The Bald Hill Claystone, 18 m thick, at the base of the succession contains about eight confusingly superimposed and well-differentiated paleosols. The area probably received little sediment for 16,000 years or more because it was very slowly subsiding, freely drained, rolling land. By contrast, the paleosols are completely separated by sediment in the overlying Garie Formation. These humic gley (fibrist) paleosols indicate increased subsidence culminating in deposition of subaqueous lagoonal shale. In the overlying Newport Formation, paleosols are also well separated by sediment, but show several indications of immaturity, such as widespread sedimentary relicts within the profiles. This would indicate a steady subsidence rate of about a meter every 2,000 years. Higher within the Newport Formation, paleosols are seldom preserved. Paleosols are very rare in the overlying braided stream deposits of the Hawkesbury sandstone. This probably is due to very low rates of subsidence, allowing extensive lateral migration of streams and almost total reworking of floodplain deposits.

Allen (1974b, 1974c) has developed several theoretical models to explain the distribution of paleosols and channel deposits expected under varying conditions of subsidence, stream behavior (laternal migration as opposed to channel avulsion), and climatic fluctuations. Slow subsidence of the order of a meter every 5,000 years and periodic channel avulsion may best explain the distribution of paleosols in the Anglo-Welsh outcrop of the Siluro-Devonian Old Red Sandstone.

Climate

Temperature and rainfall are such important factors in forming modern soils that many modern soils are restricted to particular climatic zones.

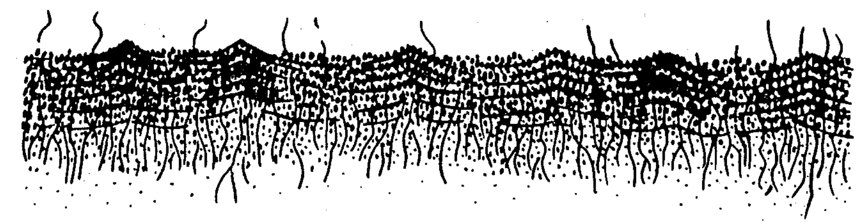

FIGURE 3.10. Schematic drawing of tepee structure in Devonian caliche-bearing paleosols of the Anglo-Welsh outcrop of the Old Red Sandstone. The carbonate nodules (black ellipses) and tubular structures (lines) are largely confined to an horizon about 2 m thick. (Redrawn from Allen [1974b].)

animals. They were evidently vegetated, at least sparsely. Much has been learned of the evolution, anatomy, and morphology of the primitive plants associated with these paleosols (Banks 1968), but it is too early yet to say what kinds of plant communities vegetated these various fossil soils.

McPherson (1979) has made a detailed study of similar fossil soils from the Late Devonian (probably Famennian) Aztec Siltstone of southern Victoria Land, Antarctica. These are riddled with small root casts and burrows, calcareous nodules, and vein networks, and also show tepee structures and color mottling. Sepic plasmic fabrics (as in Fig. 3.4) are common in these paleosols. The down profile decrease in SiO_2 and increase in Al_2O_3, K_2O, and TiO_2 indicates that there was significant illuviation of clay. Increasing total iron (Fe_2O_3 + FeO), CaO, and MnO with depth, is related to a down profile increase in hematite and carbonate. These changes indicate very rudimentary development of A and B horizons, as found, although much better differentiated, in modern forested soils. McPherson compares these Devonian paleosols with red or brown clays, red brown earths, calcareous red earths, and red earth soils of the Australian classification (Stace et al. 1968). Only a few fragmentary lycopod fossil plants have been found in Devonian rocks of Antarctica (Plumstead 1964). Judging from later representatives of this group of plants, these probably colonized wetter habitats than these fossil soils, whose vegetation is still unknown.

As discussed by Schumm (1968), the emergence of a land flora had a great effect on stabilizing stream channels, particularly in promoting meandering rather than braided stream courses. It also served to delay and lessen the devastation of flash flooding after rains in some parts of the landscape. As a result, soil formation on the interfluves was less frequently and less critically interrupted by sedimentation.

Mid-Paleozoic Appearance of Woodland Soils

The first woodlands and forests would have had a considerable impact on the world's land surfaces. In some ways their greater biomass would have

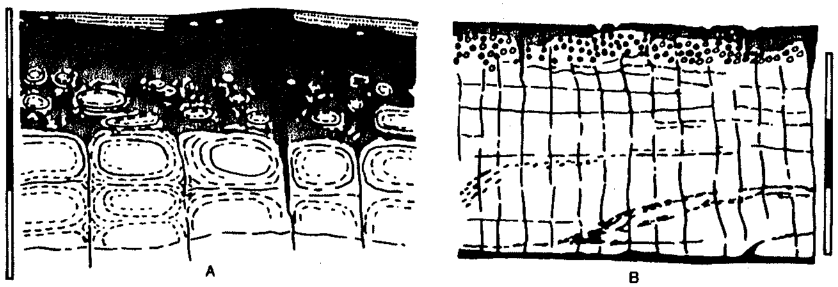

FIGURE 3.9. Late Ordovician paleosols from the Dunn Point Formation, near Arisaig, Nova Scotia; **A**, red soil material (black), with surficial hollows filled with redeposited soil material (stipple) and near surface reduction spots (white ovals), overlying spheroidally weathered andesite. **B**, red soil material swept up into overriding andesite flow. Scales in meters. (Redrawn from Boucot et al. [1974] with permission from the author and the *Geological Society of America.*)

unconsolidated red soil was also prone to slumping, as it forms thick rubbly mudflows filling in ancient channels. The original soils were evidently exposed to the surface with no overlying horizons, and were also unindurated. For these reasons, they should not have been called laterites (by Dewey, in Boucot et al. 1974), as that term is usually understood (McFarlane 1976). Without additional details these fossil soils cannot be accurately identified, but are probably better compared with some of the less differentiated oxisols (Soil Survey Staff 1975) or red earths, calcareous red earths, or krasnozems (Stace et al. 1968). It is quite possible, as suggested by Dewey, that the reduction mottles and the surficial erosion scours indicate clumps of vegetation. This also deserves further attention.

There are also fossil soils in the Red Member of the Moydart Formation of likely latest Silurian age, also in the Arisaig area (Boucot et al. 1974). These consist of about 2 m of red micaceous siltstone with numerous very light gray ellipsoidal calcareous nodules which become more numerous and even fused toward the top of the profile. This kind of fossil soil is also characteristic of Devonian rocks such as the middle Downtonian to middle Dittonian Knoydart Formation, also in the Arisaig area (Boucot et al. 1974) and in many other parts of the late Silurian and Devonian Catskill and Old Red facies of the Northern Hemisphere (Friend, Harland, and Gilbert-Smith 1970; Woodrow, Fletcher, and Ahrnsbrak 1973; Allen 1973, 1974a, 1974b; Leeder 1976). These paleosols resemble modern soils with caliche horizons, forming in warm to hot regions (mean annual temperature 16–20°C) of limited rainfall (mean annual rainfall 100–500 mm). Tepee structures in some of these paleosols indicate periodic wetting, probably during more than one season of the year (Fig. 3.10). These fossil soils also have numerous vertical and branching pedotubules and crystal tubes up to several millimeters in diameter, probably formed by both plant roots and burrowing

However, caution is advisable when interpreting climate from fossil soils for at least two reasons. Firstly, the climatic regime of a soil, particularly of the broader categories at the level of a suborder (of Soil Survey Staff 1975), is usually a generalization which may not apply strictly to each case. Secondly, climate is only one of a number of other factors which must be considered: time of formation, topographical position, soil fauna, vegetation, and parent material.

Caliche-bearing paleosols have been widely considered to indicate warm, semiarid conditions (Allen 1974b; Hubert 1977a, 1977b), and lateritic paleosols to indicate hot, seasonally humid conditions (Abbott, Minch, and Peterson 1976; Peterson and Abbott 1979). While broadly true, there are complications and exceptions to these generalizations which must be considered before applying them to specific paleosols (Reeves 1970; Paton and Williams 1972). As another example, the dominance of podzolic paleosols, even on a variety of parent materials, as well as evidence from associated flora and paleolatitude indicated by paleomagnetic data, indicate that the climate of the area around Sydney, Australia was moderately humid and cool temperate during the early Triassic (Retallack 1977b). Until fossil soils can be identified with greater precision, only such general impressions of paleoclimate will be gained from them.

DIVERSIFICATION OF SOILS THROUGH TIME

Much can be learned from the uniformitarian interpretation of fossil soils. Perhaps the most exciting prospect of continuing paleoenvironmental studies of fossil soils will be gaining a better understanding of the diversification of soils through geological time. As knowledge of fossil soils has lagged so far behind that of other geological and paleontological sciences, such studies can be expected to integrate evidence from a variety of sources into a better understanding of the evolution of terrestrial ecosystems. The critical events for such a scenario were probably the evolution of the atmosphere, the origin of terrestrial organisms, the emergence of rooted plants, the development of woody plants, the expansion of grasslands, and the impact of human beings on the land surfaces of the world. Broadly similar scenarios have been postulated by Yaalon (1971) and Hunt (1972). Existing studies of fossil soils pertinent to these critical events are few, but show much promise for the future.

Abiotic Soils of the Precambrian

Precambrian fossil soils are difficult to recognize with certainty, because one of the most diagnostic features of Devonian and later fossil soils, roots in growth position, are not found in them. Most Precambrian paleosols recognized to date have been below major unconformities. The many problems of this kind of geological setting have already been discussed. There is a need for more earnest theoretical modeling of the likely weathering effects of different hypothetical Precambrian atmospheres and also for more detailed micromorphological and geochemical studies of Precambrian paleosols.

Collins (1925), Roscoe (1968), and Frarey and Roscoe (1970) have described a fossil soil, about 2.45 billion years old, developed on pre-Huronian crystalline rocks underlying the various basal formations of the Elliot Lake Group in the area between Sudbury and Elliot Lake, north of Lake Huron, Canada. This is a zone, up to 16 m thick, of altered biotite granite and greenstone on the pre-Huronian surface. Above unaltered pink biotite granite, there is a thick white rock with granitic texture containing highly altered mafic minerals, plagioclase almost entirely altered to sericite, and scattered inclusions of unaltered granite. Higher in the profile granitic texture is no longer present, even microcline is partly replaced by sericite and only quartz grains persist unaltered. The highest part of the profile is a greenish rock with quartz grains and remnants of microcline floating in a structureless matrix of sericite (mica-illite). Accessory minerals persisting in the altered rock include hematite, magnetite, pyrite, rutile, zircon, monazite, thorogummite (probably thorite originally), garnet, and amphibole. Thinner (about 1 m) profiles, developed on greenstone, consist of a greenish, gray or pale yellowish rock with a high sericite content. Pyrrhotite (or sometimes other sulfide minerals) may form a thin layer, up to 3 cm thick, immediately below the unconformity on these greenstone profiles. The pre-Huronian land surface was evidently leached of most of its CaO, SrO, and MnO, much of its MgO, Na_2O, FeO, and Fe_2O_3, and perhaps a little SiO_2 and Al_2O_3. Strangely, water, Rb_2O_3, and K_2O appear to have accumulated. Other aspects of this alteration are comparable with modern weathering, apart from the general loss of iron, the greater loss of Fe_2O_3 than FeO, and the extreme loss of MnO. These differences are probably due to the absence of oxygen in the atmosphere at that time. An anoxic atmosphere is also indicated by pyritic conglomerates containing placers of detrital uranium minerals, such as uraninite and brannerite, in the overlying Elliot Lake Group. Lack of oxygen is in good accord with modern theory that the Precambrian atmosphere was largely derived from degassing of the earth's interior from volcanic vents and so consisted largely of gases such as CH_4, NH_3, CO_2, H_2, H_2O, H_2S, and N_2 (Berkner and Marshall 1965).

Another paleosol, possibly formed in an oxygen-poor atmosphere about 1.8 billion years ago, has been found in the Tampere area of Finland (Rankama 1955; Eskola 1963). This is a breccia of diorite with a gray schist matrix developed on fresh diorite and overlain unconformably by varved mica schist. Analysis of all these rocks showed a preponderance of FeO over Fe_2O_3, as in the pre-Huronian paleosols north of Lake Huron.

jasper-bearing paleosol (McBride and Folk 1977). Finally, *Microcodium* is a widespread microfossil commonly preserved in caliche-bearing paleosols. Klappa (1978) has interpreted it as a fungus, possibly mycorrhizal, and has discussed numerous occurrences dating back to Jurassic times.

Early Paleozoic Advent of Soils of Vascular Land Plants

The study of fossil soils may make a fundamental contribution to the current debate on when the first vascular land plants appeared. Some (Gray and Boucot 1977) feel that dispersed trilete spores, tracheid-like bodies, and cuticle fragments found in rocks as old as late Ordovician are the first evidence of vascular land plants. Others (Banks 1975a, 1975b; J. M. Schopf 1978; Edwards, Basset, and Rogerson 1979) are only prepared to accept as evidence of the first vascular land plants the complete megafossils of late Silurian age. The debate has opened some very difficult, perhaps insoluble, questions. To what extent do specific structures of fossil plants necessarily indicate their affinities or paleoenvironment? Is the adaptive value of such structures in living plants the same as it was to the first land plants? Which and how many features are important indicators of terrestrial habitat? Were some of the earliest land plants adapted to partial or periodic exposure to the air? Studies of the fossil substrates supporting these early plants should give a more definitive and detailed perspective on the problem. The holdfasts, rhizoids, rhizomes, and roots of primitive plants may be well-preserved in aquatic shales and cherts. Although not preserved as well in more oxidizing environments, such structures may significantly modify soil material. They promote the obliteration of relict structures from the parent material and development of sepic plasmic fabrics, cutans, and glaebules, or leave more obvious pedotubules (this terminology is after Brewer 1964).

Early Paleozoic redbed sequences should be examined in more detail with this aim in mind. First on my list are the Ordovician and Silurian paleosols described by Boucot et al. (1974) from the Arisaig area of Antigonish County, Nova Scotia. Here paleosols cap several columnar-jointed andesite flows in the late Ordovician Dunn Point Formation. The 1.3 m thick profiles consist largely of a homogeneous aggregate of granular hematite and clay, which extends deeper into cracks within the spheroidally-weathered andesite (Fig. 3.9). The lower portion of this altered zone contains upward fining andesite corestones. This is overlain by an horizon with poorly developed spherulitic texture. The upper portion of the red material is blocky, with nodules and irregular patches of chalcedony and carbonate. Near the surface of the profile there are small white reduction spots and irregular pockets, about 1 m wide and 20 cm deep, filled with bedded, redeposited red material. Plastically deformed fragments of the red paleosols are commonly entrained in overriding andesite and ignimbrite flows. The enormous. Erosion resistance of cyanobacterial mats of modern deserts was demonstrated by Booth (1941), who found that runoff from them was clear, while that of adjacent areas was muddy. Today these mats may precipitate carbonate crusts in arid areas (Campbell 1979). As in modern soils (Bloomfield 1964; Zajic 1969), Precambrian microorganisms were probably capable of forming soluble metal chelates (particularly of iron) which could then be transported in solution to other parts of the profile or washed out entirely. Like modern microbes, they were probably also active in decomposing aluminosilicate minerals, releasing different elements at different rates and controlling other chemical processes by regulation of pH. Local oxygen produced by soil microorganisms may have been critical in the formation of insoluble oxides of aluminium and iron, thus fixing these elements within the soil.

Campbell (1979) has suggested that microbial influence of soils may have begun as long as 2.4 billion years ago, on the basis of 0.25 percent reduced organic matter informally reported to her by D. Grandstaff from a paleosol of that age in the Blind River Formation of Ontario, Canada. It is also possible, although presently impossible to prove, that the oxidized surface and likely caliche of younger Precambrian paleosols, and perhaps the surficial sulfides of older ones, owe their clear differentiation to the action of soil organisms. Direct evidence of Precambrian soil microbiota is inconclusive. Cherts filling cracks into basement rocks unconformably underlying the Pokegama Quartzite, a little more than 2 billion years old, contain nostocacean cyanobacteria and also possible budding bacteria (Cloud 1976). However, it is uncertain how often, or if, this material was exposed when the organisms lived. There are similar problems with the supposed terrestrial fungi described by Hallbauer and Van Warmelo (1974) from the base of the 2.3–2.7-billion-year-old Witwatersrand Group of South Africa. Although the carbon is evidently biogenic, the structures are more likely artifacts of their preparation procedures, as are several other such claims (Cloud 1976; Barghoorn, Chap. 2 [this volume]).

Future micromorphological studies of Precambrian terrestrial rocks may be more revealing, judging from discoveries of soil microorganisms in younger rocks. The symbiotic relationship between land plants and endotrophic fungi may be as old as early Devonian, as fungi of that age are preserved within vascular land plants of the Rhynie chert in Scotland (Kidston and Lang 1921; Pirozynski 1976). In the late Devonian Caballos Novaculite, near Marathon, Texas, a variety of filamentous cyanobacteria, green algae, and possible fungi were found in jasper-filled cracks of a paleosol developed on older sabkha deposits (Fairchild, Schopf, and Folk 1973), although these identifications were later doubted by Schopf (1975). *Callixylon* wood found in the Caballos Formation indicates that trees also grew in the area, but it is unclear if, or which, vascular plants grew in the

Although some oxygen in the primitive atmosphere may have come from the photodissociation of water by ultraviolet light, a more productive source of oxygen was probably photosynthesis by marine microorganisms (J. W. Schopf 1975, 1978). There is much that could be learned about this process from fossil soils. Was there more oxygen in and near the sea than on land? Was oxygen dispersed or limited to certain areas? With very low amounts of oxygen, were some elements and minerals of Precambrian soils affected more than others?

The atmosphere evidently contained appreciable amounts of oxygen by about 2 billion years ago, as redbeds began to appear at about that time. In the Huronian sequence north of Lake Huron the advent of a more oxygenated atmosphere is indicated by monazite-rich, hematitic conglomerates of the Lorrain Formation, about 2.1 billion years old (Roscoe 1968). Other Precambrian redbeds of about this age are mentioned by Davidson (1965). Some of these may prove to be paleosols. All deserve more critical attention. Donaldson (1969) and Fraser et al. (1970) have reported a hematitic altered zone on crystalline basement underlying the Athabasca Formation, probably 1.7 billion years old, in northwestern Saskatchewan, Canada. Hematite-stained, altered zones are also widespread just below the base of the Grand Canyon Series in the Grand Canyon, Arizona (Sharp 1940), at a surface older than 1.15 billion years and possibly as old as 1.4 billion years (Livingston and Damon 1968). Other paleosols which may provide evidence on the nature of primitive atmospheres are those mentioned by Hoffman, Fraser, and McGlynn (1970) interbedded with volcanic flows of the Seton Formation, 2.0 to 1.75 billion years old, in the Great Slave fold belt, Northwest Territory, Canada and also by Blades and Bickford (1979) from a 1.5-billion-year-old sequence of volcaniclastic sediments and rhyolitic ash-flows in Missouri, U.S.A.

By about 1 billion years ago, weathering was becoming more like it is today. Williams (1968) has described a fossil soil from the Cape Wrath district of northwest Scotland, probably about a billion years old (Anderson 1965). This was developed on an old topographical surface of Lewisian biotite gneiss, amphibolite, and microcline pegmatite with a relief of at least 600 m, and covered by Torridonian alluvial fan deposits (Fig. 3.8). In the profile developed on biotite gneiss, the upper 30 cm is usually stained very dark red and is massive. Underlying this is a bleached zone of very pale green or grayish green rock with relict gneissic foliation. The bleached zone passes gradationally into light gray, fresh gneiss. The altered zone extends 1 to 3 m below the surface, and up to 6 m locally along joints and around unaltered corestones. Quartz and microcline persist throughout the profile with some corrosion and cracking, but biotite and plagioclase of the original gneiss are extensively altered. The profile developed on greenish black amphibolite is a soft dark greenish gray rock, 1.0 to 1.3 m thick and laced with veins of white carbonate up to 5 mm wide. Quartz and microcline are also little altered in

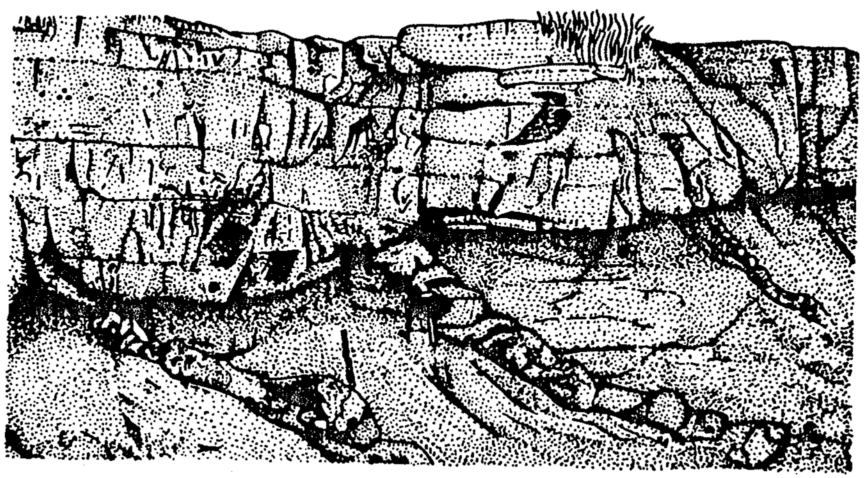

FIGURE 3.8. One-billion-year-old fossil soil on Lewisian gneiss with prominent pegmatite veins, unconformably overlain by Torridonian conglomerate and sandstone, at Sheigra, northwest Scotland. The reddish surface of the fossil soil is shown in heavier stipple. Hammer gives scale. (Redrawn from Williams [1968] with permission from the *Scottish Journal of Geology*.)

this profile, but all the original hornblende and most of the plagioclase have been altered to a pale green micaceous mass with patches of chlorite, carbonate, and iron oxide. Pegmatite veins were little altered. These and the foliation were bent at the unconformity in some outcrops, interpreted by Williams as evidence of Precambrian soil creep. Both the red and bleached portions of the profile on the biotite gneiss have lower SiO_2, FeO, CaO, and Na_2O and higher Fe_2O_3 and K_2O than fresh gneiss. In the red portion of the profile, Al_2O_3, Fe_2O_3, K_2O, TiO_2, and P_2O_5 are at maximum and SiO_2, FeO, and Na_2O at a minimum. Williams believed that the red portion of the profile was a remnant of a thicker, probably podzolic soil. There is no evidence of any different higher horizon, nor is it necessary. The accumulation of aluminium, iron and potassium and comcomitant desilication was more likely produced by lateral flow of ground and surface water, rather than by illuviation more characteristic of modern forested soils. These paleosols were probably developed on freely drained, rolling parts of valleys in a warm, moderately humid climate and oxygenated atmosphere.

Kalliokosi (1975) has described comparable paleosols of similar age from Presque Isle, north of Marquette, Michigan, developed on granodiorite, diabase, and serpentinized periodotite, where these are unconformably overlain by late Keweenawan Jacobsville Sandstone. He interpreted the dolomite-quartz layers in these fossil soils as caliche horizons, indicating semiarid climate. The exact nature of these altered rocks has been disputed

(Lewan 1977; Kalliokosi 1977). Paleosols have also been found in several parts of the Grand Canyon of Arizona at the unconformity immediately below early and middle Cambrian sandstones (Sharp 1940; McKee 1969). Patel (1977) has reported another paleosol at the unconformity covered by early Cambrian rocks in the Saint John District of New Brunswick.

Compared to earlier times, these later Precambrian and Cambrian paleosols were leached of more silica and accumulated more iron, especially Fe_2O_3. This had a considerable effect on shallow marine sedimentation. Banded iron formations became much less common. Silicified shallow marine stromatolites, often including exquisitely preserved microorganisms, became more common (Hargraves 1976; Schopf 1975). More intense terrestrial weathering is probably also a partial explanation, besides long time of formation, for the supermature quartose sandstones deposited in rivers, beaches, and shallow marine continental shelves the world over during the late Precambrian and Cambrian.

Later Precambrian Microbially Influenced Soils

It is likely that the land was partly colonized by algae, bacteria, and viruses, and perhaps even fungi, lichens or liverworts long before the first vascular plants. The green slime of shallow Precambrian seas has been studied in some detail (J. W. Schopf 1978), but did it venture out to become the "scum of the earth"? There is little firm evidence for this idea, and still a need for more perceptive and detailed speculation.

Many of the kinds of microorganisms found in cyanobacterial mats of modern desert crusts were well represented in later Precambrian marine rocks (Campbell 1979). In modern deserts these organisms are capable of withstanding high salinity and long desiccation, with instant reactivation after rain. They are also capable of traveling overland by mechanical expulsion of trichomes during rehydration and by self-propelled gliding wherever there is moisture. Many are of a size easily transported by wind. Fischer (1965) and Sagan (1965) speculated that the high ultraviolet radiation during the earlier Precambrian prevented organisms from colonizing the surface of the ocean, the intertidal zone, and land. They also postulate that radiation was only reduced to tolerable levels by the increasing amounts of oxygen in the atmosphere. Schopf (personal communication 1980) doubts the severity of this radiation on the basis of very old stromatolites and likely planktonic organisms. But even allowing such radiation, microorganisms could have still survived on land below the surface in unconsolidated diaphanous materials and also in "shade oases" of crevices and overhangs. Small outposts of life on land may have been important centers of organic evolution.

The effects of the first soil organisms on soil formation would have been