Phil. Trans. R. Soc. Lond. B **309**, 105–142 (1985) [105]
Printed in Great Britain

Fossil soils as grounds for interpreting the advent of large plants and animals on land

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[Plates 1–4]

A little-explored line of evidence for the antiquity and nature of early vegetation on land is the soils in which they grew. Vegetation is one of a number of factors known to play an important role in forming modern soils and soil features. As in studies of the role of organisms in modern soil formation, what are needed are fossil soils that supported different ancient ecosystems, but for which other soil-forming factors, such as palaeoclimate, palaeogeographical setting, parent materials and duration of formation, were closely comparable. This preliminary petrographic and chemical study compares four palaeosols; three are from the states of Pennsylvania and New York, U.S.A., and the fourth from the Potwar Plateau region of northern Pakistan. All appear to have formed in a subhumid, seasonally dry, subtropical climate, on the alluvial outwash of major mountain ranges, largely formed of sedimentary and metamorphic rocks, and over periods of only a few thousand years. These palaeosols are here named the Potters Mills clay (late Ordovician), Lehigh Gap clay (late Silurian), Peas Eddy clay (late Devonian) and Lal clay (late Miocene). Successively younger palaeosols show increasing degree of weathering, more clayey texture and better soil structure. Deep burrows are abundant in late Ordovician palaeosols, and are evidence of sizeable (3–16 mm diameter) soil animals. Bioturbation in the surface of the late Silurian palaeosol may have been produced by animals or vascular land plants. Large root traces and remains of leaf litter are indications that the late Devonian palaeosol supported a low diversity, streamside gallery forest. Weak redistribution of iron in this palaeosol may have been produced by phenolic and other herbivore suppressant toxins from these early trees. The late Miocene palaeosol is extensively bioturbated, presumably by termites and other creatures. Judging from its root traces and associated sediments and fossils, it supported gallery forest in a region of grassy savanna groveland. These early results encourage the belief that fossil soils may provide useful evidence for the nature of early ecosystems on land, not only complementary to that of early terrestrial fossils, but also in sedimentary sequences too oxidized and acidic to allow preservation of fossil plants and animals.

Introduction

The earth was fundamentally changed following the colonization of land by multicellular plants and animals. A green living mantle mitigated soil erosion, promoted clay production, changed patterns of alluvial deposition and contributed to the oxygenation of the atmosphere. Much has been written about these general consequences of the advent of land vegetation (Weaver 1969; Schumm 1968, 1977; Ronov *et al.* 1980; Moore 1983). By contrast, the immediate consequences of early land plants on the soils in which they grew have been strangely neglected. A few perceptive individuals, such as Joseph Barrell (1913) long ago recognized the existence of Palaeozoic palaeosols. It is only in the last decade that the sheer abundance of early

Palaeozoic palaeosols has become apparent (Allen 1973, 1974*a*, *b*, *c*; Boucot *et al.* 1974, 1982; McPherson 1979; Retallack 1981; Parnell 1983). It is now possible to choose from a surfeit of palaeosols, those relevant to understanding particular aspects of early terrestrial ecosystems.

In his still influential book, Hans Jenny (1941) argued that the formation of modern soils could be related to a limited number of factors: climate, organisms, geomorphological setting, parent material and time available for formation. To study the role of any one of these factors quantitatively, Jenny recommended seeking study sites in which the other factors were more or less constant. In a demonstration of the role of vegetation in soil formation, he plotted soil reaction (pH), and amounts of organic carbon, clay and carbonate in soils formed under forest and grassland in the prairie–timber transition zone of Illinois. From such relationships he hoped to derive mathematical equations describing soil formation, which he termed biofunctions. His dream of a multivariate mathematical description of soil formation has not yet been realized, but numerous other biofunctions, as well as climofunctions, topofunctions, lithofunctions and chronofunctions have been documented since (Birkeland 1984; Buol *et al.* 1980). In this paper I attempt to examine selected Palaeozoic palaeosols in a similar way. I have searched widely for palaeosols for which palaeoclimate, palaeogeography, parent material and duration of their formation were similar, so that changes in their vegetation over geological time would be especially evident.

Materials and methods

Here I discuss only four palaeosols, and briefly characterize associated profiles. I make no claim that these are representative. Indeed, they were chosen as the best-developed and easiest to understand of the palaeosols examined. They are each presumed to have supported some of the most complex ecosystems of their times. My fieldwork so far has concentrated principally on Cambrian to Triassic red beds in the states of Ohio, Pennsylvania, Connecticut, New Jersey and New York (figure 1). I considered also discussing an analogous modern soil, but thought it better to add a description of a late Miocene palaeosol from the Potwar Plateau of northern Pakistan (figure 1). This comparison is not confused by features of modern soils which are not preserved in buried soils. The palaeosol from Pakistan is part of a larger project currently

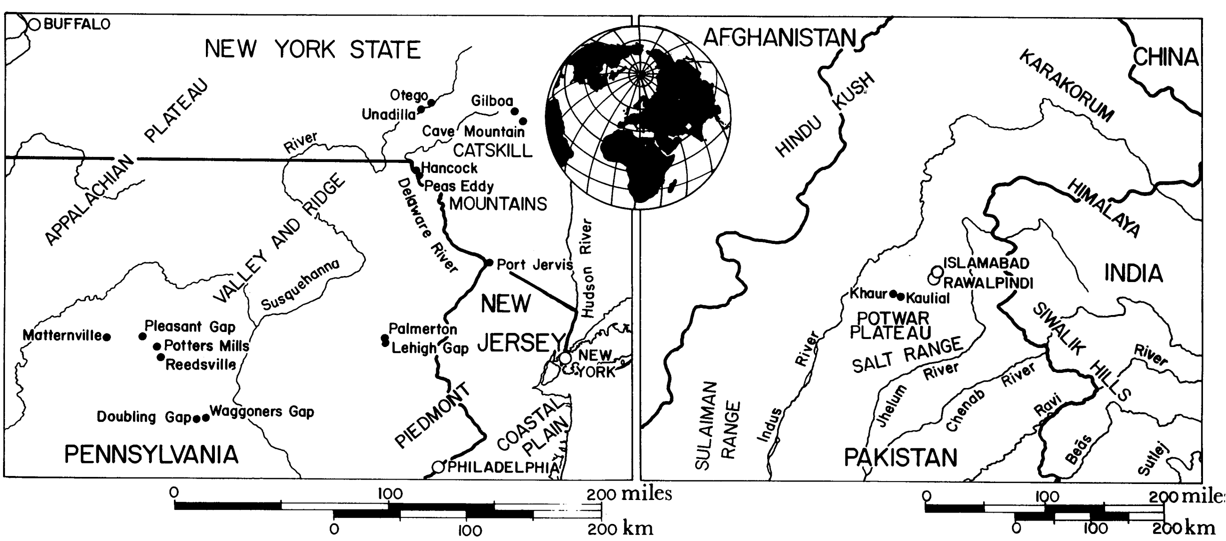


Figure 1. Mentioned localities in eastern North America (left) and northern Pakistan (right).

under-way, which is designed to establish the palaeoenvironments of ramapithecine primates from studies of the fossil soils in which they are found.

Alteration of fossil soils upon burial and destruction of their biota is such that they cannot be studied in exactly the same way as modern soils (Retallack 1981, 1983*a*). Direct measurement of pH and redox potential does not give meaningful results. Routine chemical analyses for available elements used for modern soils are also inappropriate. Nor is it possible to disaggregate effectively lithified rocks for grain size analysis. I have chosen to study palaeosols largely from field observations, point-counting of petrographic thin sections and whole-rock chemical analyses. Because of problems with swelling clay, the Miocene and some of the Palaeozoic rocks were cut and ground under kerosene, rather than water. Chemical analyses were done with a Baird P.S.1 I.C.P. Spectrophotometer in the Department of Geological Sciences, University of Washington, Seattle. Specimens are housed in the Department of Geology, University of Oregon, Eugene (those whose numbers are prefixed by UO), and in the U.S. National Museum, Washington D.C. (prefixed USNM).

This work follows as closely as possible the recommendations and terminology of Brewer (1976) and Birkeland (1984). None of the existing soil classifications can be considered comprehensive and definitive, so three of the leading ones were used here. The U.S. Department of Agriculture classification (Soil Survey Staff 1975) is best for high latitude soils. The Food and Agriculture Organization of the United Nations has produced a classification (F.A.O.–U.N.E.S.C.O. 1977; Fitzpatrick 1980) based on long experience in tropical regions. Mid-latitude soils of stable cratonic areas form the main subject of the classification of the Australian C.S.I.R.O. (Stace *et al.* 1968).

Detailed descriptions of the palaeosols and tabulations of petrographic and chemical data could not be included in this account for lack of space, but are available as an open file report free of charge from the author.

Description and interpretation of the palaeosols

Potters Mills clay palaeosol (late Ordovician)

Diagnosis

Thin (40 cm in compacted palaeosol), dark reddish brown (2.5YR3/4) to weak red (2.5YR4/2) palaeosol, with clayey surface horizon, abundant large (3–16 mm diameter) burrows and with subsurface horizons (beyond 15 cm) riddled with calcareous pseudomycelium and nodules sheathing burrows (figure 2, plate 1).

Location

This is the second palaeosol in the red claystone sequence below the thick sandstone with basal breccia (figure 3), 200 m east of the western end of the southern cutting on eastbound U.S. highway 322, 4 km east of Potters Mills, Centre County, Pennsylvania (figure 1). On U.S. Geological Survey 7.5 minute Spring Mills Quadrangle, this at latitude 40° 45′ 32″ north and longitude 77° 37′ 04″ west.

Geological setting

This palaeosol is near the top of the Juniata Formation, as mapped by Gray & Shepps (1960) and Hoskins (in Berg & Dodge 1981). The palaeosol is gently deformed, dipping to the east

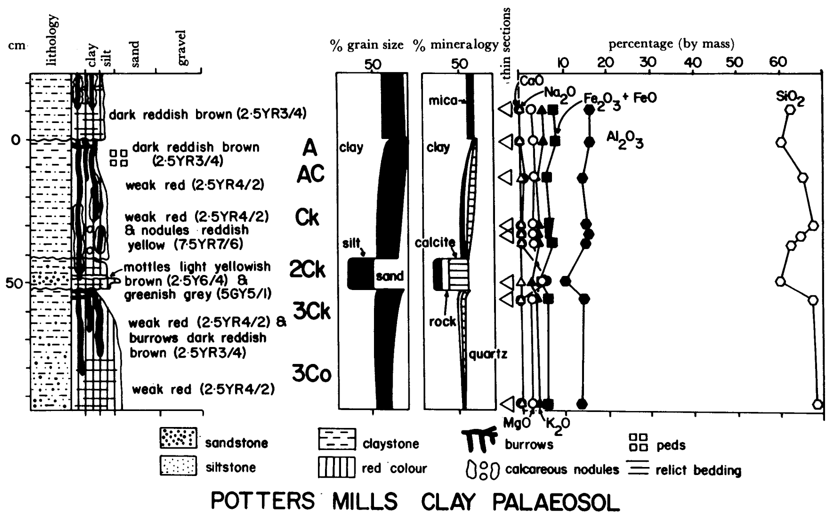


FIGURE 2. Field appearance, petrographic and chemical composition of the Potters Mills clay palaeosol, late Ordovician, central Pennsylvania. Specimen numbers from top down are UOR169 to177. A, AC, Ck, etc. refer to soil horizons (see text). Horizontal axis is percentage (by mass).

(at 21° on magnetic dip azimuth 119°) near the crest of a large anticline. This fold is near the northwestern edge of the Valley and Ridge Province of the Appalachian Orogen (of Rodgers 1970).

Age

The Juniata Formation is late Ordovician (Ashgillian) in age (Ross 1982), or approximately 438–448 Ma in the radiometric time scale proposed by Palmer (1983).

Alteration after burial

Compared with other exposures of the Juniata Formation examined on steeply dipping limbs of regional folds, this palaeosol is little deformed by shearing. Both cleavage and schistosity are barely perceptible and are at a high angle to the bedding. From the degree of alteration of conodonts (about CAI 4) in Ordovician limestones of this area of Pennsylvania, Epstein *et al.* (1977) propose likely burial depths of 4.6–6.6 km and temperatures of about 160–210 °C. This is compatible with the illitic and chloritic composition of clay in the Juniata Formation (Thompson 1970*a*). Thus metamorphic alteration is very low grade (in the sense of Winkler 1979).

Some compaction of the upper clayey parts of the profile probably accompanied burial. Subhorizontal burrows (figure 6) in this part of the profile are only about one third as high as wide, perhaps because of flattening of an originally tubular burrow. Vertical burrows do not show concertina outlines (figure 5), which would be produced by substantial compaction after burial.

Drab mottles within this palaeosol have the same petrographic texture as the red matrix. Like comparable mottles in other palaeosols (Retallack 1983*b*), these may have formed by bacterial reduction of organic matter within the soil shortly after its burial. More extensive drab beds lower in the Juniata Formation may have a similar origin (Thompson 1970*a*).

Although the palaeosol is now red with haematite, this may have formed by dehydration of ferric oxyhydrate minerals during early burial, as described in Pleistocene palaeosols by Walker (1967). The original palaeosol may have been orange or brown with minerals such as ferrihydrite or goethite.

Reconstructed soil

The original soil was probably a brown or red silt, with a clayey surface and abundant deep, vertical burrows. Below the surface were a calcareous pseudomycelium and carbonate sheathing large burrows. Considering its carbonate and reddish colour, the soil was probably mildly alkaline and highly oxidized. Burrowing invertebrates appear to have played an important role in soil formation, promoting clay production and soil structure at the surface, and aeration of the whole profile. The soil-forming process of calcification operated just below the zone of most intense biological activity, and the whole profile was well ferruginized. Both processes and the weak reorientation of clay (skelinsepic to skelmosepic agglomeroplasmic fabric) imply that this soil was well drained. The water table was probably consistently below 50 cm from the surface. It is possible that the calcareous matrix of horizon 2Ck is in part due to a water table at this level, because it has some similarities to groundwater calcrete recognized elsewhere (Netterberg 1969; Mann & Horwitz 1979). There is no mottling or gley mineralization of the palaeosol at any level that would be evidence of permanent waterlogging.

Palaeoclimate

Similar modern soils with which this palaeosol is here tentatively identified, formed in seasonally dry, subhumid, subtropical climates. Under such conditions, ferruginization proceeds during the rainy season and calcification during the dry season. The depth to the calcareous (Ck) horizon in modern soils shows a crude relationship to mean annual precipitation. From the data given by both Jenny (1941) and Arkley (1963), the Potters Mills clay would have formed in a climate of less than 40 cm mean annual rainfall, corresponding to the perarid thorn woodland life zone of Holdridge (1967). Putting aside difficulties in assessing surface erosion and compaction in palaeosols (which were probably slight in this case), comparison with modern soils may not be appropriate for other reasons. Leaching of carbonate is dependent not only on the depth of wetting, but on production of carbonic and other acids by soil biota. This is likely to have been much less for this Ordovician palaeosol than for modern soils. A wetter climate can be inferred from the abundant burrows and thoroughly weathered iron-bearing minerals, well in advance of that found in modern desert soils. Considering these features, and the likely palaeogeography and palaeolatitude of the Juniata Formation (Ziegler *et al.* 1979), a seasonally dry, subhumid, subtropical climate is most likely.

Fossil flora

No megafossil plants were found in or associated with this palaeosol. Processing of samples of associated green and grey shales by J. Gray proved them to be barren of palynomorphs. The only possible traces of plants were calcareous crystal tubes, here referred to as pseudomycelium. This non-committal name is commonly used in descriptions of weakly developed calcareous horizons (Gile *et al.* 1966) in deference to the uncertain origin of these features: perhaps formed by microbial aggregations, as fungal mycelium, as root traces or as small burrows. The origin of carbonate, which sheathes burrows in this palaeosol (figures 4, 7 and 8) is also uncertain. It may be related to soil biota, either microbial or fungal aggregations associated with the burrows, or respiration of burrow occupants.

The degree of development of this palaeosol constitutes additional evidence for plants or plant-like microbes in the original soil. Although development is modest, it is much greater than that envisaged for sterile, prebiotic landscapes (Schumm 1968, 1977), and comparable to aridland soils with sparse and intermittently active, plants and microbes (Buol *et al.* 1980; Birkeland 1984).

Better evidence for the existence of plants or plant-like microbes in this palaeosol is provided by the abundant burrows (para-aggrotubules: figures 5 and 6). It is difficult to envisage what else would support such populations of invertebrates in well-drained soils. Their pellet-like fill (figure 8) and sheathing carbonate (figure 7) are evidence against their being temporary burrows, used only for resting or aestivation. In several other palaeosols near Potters Mills, these burrows are also quite fresh, as if occupied up until the moment the soils were buried. I suspect that most of these animals were eating some kind of vegetable or plant-derived fodder. Considering the low potential for preservation of organic fossils in such oxidized soils (Retallack 1984*a*), this could mean any kind of plant or plant-like microbe other than those producing large rooting structures which would leave definite traces.

Well-preserved spore tetrads of presumed land plants have been extracted from the late Ordovician (Ashgillian) Elkhorn Formation in Ohio (Gray *et al.* 1982). Numerous and widespread spores, tubular organic structures and cuticle sheets of presumed non-vascular land plants have been extracted from sediments of the overlying Tuscarora Formation of early Silurian (Llandoverian) age in Pennsylvania (Strother & Traverse 1979) and rocks of possibly equivalent age in Virginia (Pratt *et al.* 1978). Comparable palynomorphs have been reported from other regions in rocks of similar (Llandoverian) and slightly older (Ashgillian) age (Boucot & Gray 1982). Megafossils of enigmatic erect plants have been reported from Llandoverian rocks of Maine (Schopf *et al.* 1966; Niklas 1982). Thus it would not be surprising if the Potters Mills clay palaeosol supported large non-vascular plants. Judging from its lack of discernable root or rhizome traces, any plants present were unlikely to have been vascular or to have stood very far off the ground.

Fossil fauna

As already discussed, the burrows (para-aggrotubules) of this palaeosol (figures 4–8) are interpreted as permanent dwelling burrows of invertebrate animals. None of these burrows were observed to branch and all were blunt-ended, with no evidence of living chambers or turnarounds. They are best assigned to the trace fossil genera *Skolithos* (vertical ones) and *Planolites* (subhorizontal ones). They lack funnel-like structures associated with trace fossils variously assigned to *Tigillites* and *Monocraterion*, and are more elongate than *Cylindricum* and *Neoskolithos* (Alpert 1974; Häntzschel 1975; Bradshaw 1981). Presumably they were excavated by elongate invertebrates with a body width ranging from 3–16 mm.

No body fossils have been found in the non-marine part of the Juniata Formation, nor are terrestrial organisms known with certainty from other late Ordovician rocks. One possibility is that the burrows of the Potters Mills clay were excavated by annelid worms. Earthworm-like fossils have been found in Middle Ordovician marine rocks in Quebec (Morris *et al.* 1982). The Potters Mills clay would have been rather dry for soft-skinned annelids to maintain a permanently open burrow. Relict bedding around the burrows deep within the profile provides evidence that this part of the palaeosol was not exploited in the same way that earthworms burrow in modern soils. For these reasons the burrows are more likely to have been excavated and occupied by some kind of arthropod. Both scorpions and millipedes have been regarded as early colonizers of the land, but neither have been found in rocks older than Silurian (Rolfe 1980; Shear *et al.* 1984). Furthermore, the earliest scorpions appear to have been aquatic (Kjellesvig-Waering 1966). Eurypterids had evolved by Ordovician time (Størmer 1955), and there are indications that some Silurian forms could have breathed air and moved out of water (Størmer 1977; Rolfe 1980). Even so, it would be surprising if they could dig and live in dry soil.

Parent material

Some claystone clasts in horizon 2Ck were possibly derived from other palaeosols higher within the drainage basin. Generally, however, the degree of weathering of mineral grains this low in the palaeosol is slight. Much of the source area was probably rocky and barren.

Not all of the clay in the surface of this palaeosol is likely to have been pedogenic. Stepwise increases in clay content low within the profile (figure 2) are relicts of original bedding. Some of the observed gradational fining upwards of grain size also could have been a feature of the original sedimentary unit on which the palaeosol developed.

Palaeogeography

The Juniata Formation represents alluvial outwash of the high Taconic Mountains, which formerly existed to the east in the present area of the piedmont and coastal plain of eastern North America (Rodgers 1970). This range consisted mainly of sedimentary, low-grade metamorphic and plutonic rocks, judging from the composition of Juniata sandstones (Yeakel 1962). Palaeocurrents of Ordovician streams flow northwest (Yeakel 1962), where the Juniata Formation becomes thinner and interbedded with shallow marine rocks (Meckel 1970). Ziegler *et al.* (1979) have reconstructed this part of the world at low palaeolatitudes in the southern hemisphere during Ordovician time.

The degree of oxidation and development of the calcareous horizon of the Potters Mills clay are indications that it formed on moderately well drained parts of the landscape. Presumably this would have been an alluvial terrace, perhaps no more than a metre or so above the water table. Associated sandstones appear to have formed in the beds of sandy braided streams (Cotter 1978). Judging from reconstructed palaeogeography (Meckel 1970; Dennison 1976), the Potters Mills clay was some 120 km from the mountain front to the east and at least 200 km inland from marine rocks of similar age to the west.

Time for formation

In the qualitative scheme of Retallack (1984*a*), this palaeosol is weakly developed. Its calcic horizon is differentiated only to a modest extent (Stage I, of Gile *et al.* (1966) or stage 1 of Wieder & Yaalon (1982). Comparable modern calcic horizons form over periods of less than 7000 years in the desert of New Mexico (Birkeland 1984). About 3000–4000 years can be considered a reasonable rough estimate, if the probably less acidifying effect of late Ordovician soil biota is balanced against greater rainfall on the Potters Mills clay. Degree of development of this palaeosol is slight by modern standards, but is nevertheless significant.

Identification

I hesitate to identify this palaeosol within modern classifications of soils because so many features of Ordovician landscapes were different from those of today. Some Precambrian palaeosols formed under less oxygenated atmospheres than at present may represent extinct kinds of soils (Retallack 1981). As more is learned about Palaeozoic palaeosols, some of them also may be better placed in new categories, rather than identified with modern soils.

Within the U.S.D.A. classification, this palaeosol has a weakly developed calcareous profile, like Aridisols, and a red and uniform profile, like Oxisols. The abundance of burrows is evidence of fauna and probably also rainfall higher than usual for Aridisols. The red colour may be diagenetic. High values of MgO, CaO and K_2O are evidence of base saturation greater than usual for modern Oxisols or Ultisols, such as the soil described by Ahmad & Jones (1969). There is, however, some controversy concerning the high amounts of K_2O found in Precambrian and early Palaeozoic palaeosols (Blatt *et al.* 1980). Some feel that it remained in the soil in the absence of higher land plants, whereas others feel that it was introduced during metamorphism. The dolomitic composition of carbonate nodules in the Juniata Formation (Horowitz 1965), can be taken as evidence of high base saturation, because magnesian calcite and dolomite have been found only in such base-rich modern soils as Solonetz (F.A.O. classification) and Mollisol (U.S.D.A.: Doner & Lynn 1977). Additional features unlikely for Oxisols are the shallowness of the most strongly altered horizon and the persistence of about 1% feldspar in the AC and C horizons. Among Inceptisols, it is difficult to exclude the possibility that this palaeosol was an Ochrept. This is considered unlikely because these form in cooler climates than envisaged for the palaeosol and they also tend to be lighter in colour. The Potters Mills clay palaeosol shows greatest similarity to Oxic Ustropepts.

Within the F.A.O. classification, this palaeosol can be distinguished from Ferralsols, Nitosols, Xerosols and Yermisols for the same reasons already outlined for equivalent U.S.D.A. categories. The Potters Mills clay is most like Calcic Cambisols in this system.

The palaeosol is similar to a number of categories in the C.S.I.R.O. classification. It is better developed than Red Calcareous Soils and not as clayey as Red Clays. Horizon 2Ck may have been hard setting, but was not siliceous or clayey as in Red Hardpan Soils. The Potters Mills clay is most like Calcareous Red Earths, although it is rather shallow and poorly developed compared with most of these extant soils.

Modern soils comparable to the Potters Mills clay develop on alluvial and undulating plains and gentle hillslopes, on a variety of sedimentary and igneous parent materials of felsic composition. At present, they support mostly savanna, open grassy woodland and shrub steppe in warm, subhumid to arid climates.

Other Ordovician palaeosols

A similar palaeosol is formed in sediments immediately overlying the palaeosol described here (figure 3) and another was seen in road cuttings into the Juniata Formation on old U.S. highway 322, 1 km east of Reedsville, Pennsylvania. Other palaeosols in the road cutting near Potters Mills, such as those underlying the described profile (figures 2 and 3), do not appear as well developed. They have abundant burrows, but little clay formation or calcic horizon development. Similar very weakly developed palaeosols were also seen in road cuttings into the Juniata Formation 1 km west of Matternville, Pennsylvania. In the humid climate of modern Pennsylvania, claystones and shales are weathered much more rapidly than associated sandstones. In old road cuttings into the Juniata Formation examined at Waggoners Gap, Doubling Gap and Pleasant Gap, Pennsylvania, there are also palaeosols, but all are poorly exposed. From these limited observations only two kinds of late Ordovician palaeosols are currently recognized in Pennsylvania: weakly developed calcareous red clays like the Potters Mills clay, and very weakly developed red silty soils, a palaeosol type as yet unnamed. These probably formed on moderately elevated terraces of interfluves and on near-stream floodbasins, respectively.

Thick, red palaeosols of late Ordovician age have also been reported from Nova Scotia (Boucot *et al.* 1974; Retallack 1981). Small reduction spots in surface mounds which separate swales filled with redeposited soil, could be traces of organic matter in areas of this palaeosol stabilized by clumps of non-vascular land plants (Dewey, in Boucot *et al.* 1974).

Lehigh Gap clay palaeosol (late Silurian)

Diagnosis

Thick (160 cm in compacted palaeosol) dark red (10R3/6) to dusky red (10R3/3) profile with a substantial (now 25 cm thick) clayey surface horizon and subsurface horizon (80 cm down) studded with small carbonate nodules (figure 15, plate 2).

Location

This palaeosol crops out 250 m southeast of the northern end of the large cliff exposures beside Red Hill Crescent, Palmerton, Carbon County, Pennsylvania. These exposures are at the northern end of the succession exposed through Lehigh Gap, which is 1 km south of here. On U.S. Geological Survey 7.5 minute Palmerton Quadrangle, the palaeosol is at latitude 40° 47′ 43″ north and longitude 75° 37′ 05″ west.

Geological setting

The palaeosol is in the lower part of the Bloomsburg Formation as mapped by Epstein *et al.* (1974). The top of the palaeosol near road level dips steeply (65° north on a magnetic dip azimuth of 355°), and is buckled into a medium scale (2 m amplitude) kink fold. This deformation is associated with the Stony Ridge–Godfrey Ridge Decollement Zone, which runs through Palmerton, nearby to the north (Epstein *et al.* 1974). This is one of the thrust faults near the southeastern margin of the Valley and Ridge Province of the Appalachian Orogen (Rodgers 1970).

Age

The Bloomsburg Formation in eastern Pennsylvania may range in age from latest Wenlockian to earliest Pridolian parts of the late Silurian (Hoskins 1961; Berry & Boucot 1970). The intervening Ludlovian has been radiometrically calibrated at 414–421 Ma before present (Palmer 1983).

Alteration after burial

The prominent fracture cleavage and quartz veins in this exposure (figure 9) are evidence of substantial metamorphism and deformation. According to Epstein *et al.* (1974), the mineralogy of the Bloomsburg Formation in this area tallies well with the quartz–muscovite–albite–chlorite subfacies of the greenschist facies of regional metamorphism. Additional evidence for a higher degree of metamorphic alteration of the Lehigh Gap clay compared with the Potters Mills clay, comes from alteration of conodonts (CAI 4.5–5) and coalified debris (vitrinite reflectance of +3.5) in Silurian to mid-Devonian marine rocks of this area. From these data, Epstein *et al.* (1977) propose burial temperatures of 220–260 °C and depths of 6.7–7.9 km. These are the kinds of depths in which contacts between sand grains are largely concavo-convex and in places sutured (Taylor 1950), and both kinds of grain contacts can be seen in the sandy part (horizon 5Co) of the Lehigh Gap clay palaeosol.

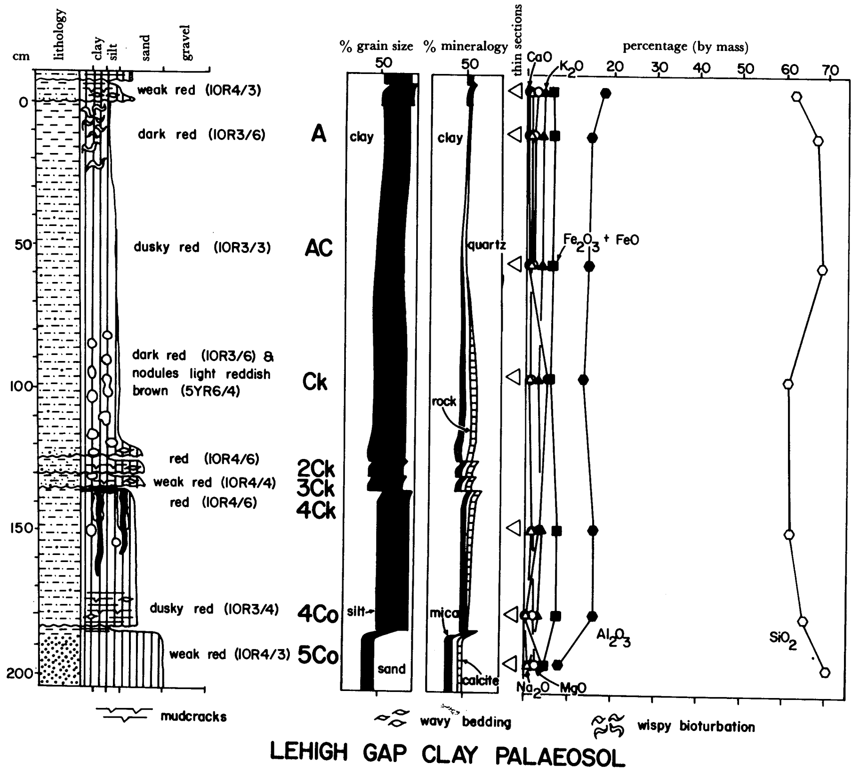


FIGURE 15. Field appearance, petrographic and chemical composition of the Lehigh Gap clay palaeosol, late Silurian, eastern Pennsylvania. Specimen numbers from top down are UOR154 to 160. Unspecified symbols are as for figure 2. Horizontal axis is percentage (by mass).

The thickness of this palaeosol may have been substantially altered during metamorphism. Compressed grain contacts are evidence of compaction. Quartz veins, on the other hand, have the appearance of tension gashes found in areas of local tectonic thickening. There is some microfaulting of carbonate nodules within the profile, but these are mildly flattened and close to their presumed originally equant shape, like those of the Lal clay palaeosol. Vertical burrows deep within the profile do not appear to be folded or deformed appreciably. Although some compaction of the profile is likely, it was probably not extreme (more than half) because of the silty and sandy texture of the original soil.

Early diagenetic dehydration of orange and brown ferric oxyhydrate minerals to the present red haematite, discussed for the Potters Mills clay palaeosol, is also likely.

Reconstructed soil

The original soil was probably a thick brown or red silty clay with an horizon of abundant, small calcareous nodules (figures 13 and 14) at depth (80 cm in the compacted palaeosol). It was well bioturbated (figures 10–12) near the surface (25 cm in the compacted palaeosol). Most of the mica and feldspar and some rock fragments were weathered to clay for a considerable distance below the surface (now 70 cm: figure 15). Soil pH in the upper part of the profile may have been mildly alkaline to neutral, but it was probably alkaline below that. From its colour, carbonate and weakly reoriented clay (isotic to skelinsepic microfabric), the palaeosol appears to have been highly oxidized and well drained, with permanent water table well below 2 m. The principal soil-forming processes appear to have been clay production at the surface, calcification at depth and ferruginization throughout the profile. These are all similar in kind to processes envisaged for the Potters Mills clay, but their scope and the time over which they were active appear to have been much greater in this palaeosol.

Palaeoclimate

Like the Potters Mills clay palaeosol, this one is also typical for subhumid, seasonally dry, subtropical climates. The depth to the top of the calcic horizon, as it is preserved at present, is similar to that found in modern soils receiving 40 to 75 cm mean annual rainfall, corresponding to Holdridge's (1967), arid, very dry forest life zone. It may have been wetter than this if allowance is made for compaction and surface erosion. It is difficult to be certain whether this difference from the rainfall estimate gained from the Potters Mills clay is due to a wetter climate or to the acidifying effect of a more substantial soil biota in the Lehigh Gap palaeosol.

Fossil flora

No megafossil plants were found in association with this palaeosol, although some degree of vegetative cover can be inferred from its surface bioturbation, degree of development and depth of the calcic horizon. The abundant tubular features (ortho-isotubules and metagranotubules) are so indifferently preserved (figures 10–12) that they could be interpreted as traces of animals or plants. Their irregular shape and branching is suggestive of rhizomes and other root-like structures. On the other hand if they are animal burrows, then the existence of some kind of plant-derived fodder can be inferred, even if it was not substantial enough to leave a trace.

At one locality within the Bloomsburg Formation, only 22 km west along strike from Palmerton and still well east of marine-influenced parts of the formation, coaly layers yielded a compressed plant axis 1–2 cm in width (Willard 1938). This was identified by W. C. Darrah as a primitive land plant, although its exact nature is still in doubt. Remains of nematophytes form coaly layers in other Silurian rocks of Pennsylvania (Strother & Traverse 1979; P. K. Strother, personal communication). These are thought to have been non-vascular land or semi-aquatic plants (Lang 1937; Niklas 1982). The oldest megafossil vascular plants known in North America are latest Silurian (Pridolian) rhyniophytes, *Cooksonia*, from New York (Banks 1969). Megafossil vascular plants have also been reported from rocks of possible mid-Silurian (Wenlockian) age in Libya, and from Irish mid-Silurian (Wenlockian) and Welsh late Silurian

(Ludlovian) rocks (Boucot & Gray 1982). A vegetative cover of xeromorphic rhyniophytes is more likely than nematophytes, for the well-drained Lehigh Gap clay.

Fossil fauna

There are abundant, large (5–12 mm diameter) subhorizontal and vertical burrows (para-isotubules) in a separate palaeosol low within the Lehigh Gap clay. This was probably formed before the Lehigh Gap clay, as a soil similar in many ways to those found associated with the Potters Mills clay. It attests to the persistence of similar communities during late Silurian time. Scorpions and eurypterids are known from late Silurian (Ludlovian) rocks in western New York (Kjellesvig-Waering 1966), but only eurypterids have been found in the Bloomsburg Formation (Hoskins 1961). Similar fossils have also been found elsewhere, as well as millipedes, known from late Silurian (Ludlovian and Pridolian) rocks, and very doubtfully from older Silurian rocks (Rolfe 1980; Shear *et al.* 1984). Silurian scorpions appear to have been aquatic (Kjellesvig-Waering 1966). So were most Silurian eurypterids, although some may have been amphibious (Størmer 1977; Rolfe 1980). Eurypterids or millipedes appear to be the most likely burrowers in the palaeosol below the Lehigh Gap clay.

Although rather more like traces of plants than animals, wispy bioturbation (some irregular ortho-isotubules and most of the metagranotubules) of the uppermost Lehigh Gap clay, could be traces of soil fauna. If so, then they represent a different assemblage of smaller animals than those in the lower palaeosol. This difference could be related to the likely better drainage of the Lehigh Gap clay, compared to palaeosols with large vertical burrows.

Parent material

The Lehigh Gap clay palaeosol formed on silty alluvium. Some traces of original bedding are preserved throughout the profile. Since some of these form varve-like, fining-upwards, laminae, it is possible that this material was deposited in a depression that was especially deeply inundated during floods. This different depositional setting from the Potters Mills clay may explain why the Lehigh Gap clay shows little sign of having developed on an upward-fining sedimentary unit. Few clay clasts likely to have been redeposited from erosion of other soils were seen. As indicated by Epstein *et al.* (1974), the source terrain of these sediments was probably rocky and barren.

Palaeogeography

Red beds of the Bloomsburg Formation represent alluvial outwash of the Taconic Mountains to the east, deposited in response to a mild resurgence (Salinic disturbance, see Rodgers 1970) of earlier earth movements. Compared with their late Ordovician extent, these mountains were much reduced in height and width (Berry & Boucot 1970). Judging from the composition of Bloomsburg sandstones (Epstein *et al.* 1974), these mountains mainly consisted of sedimentary, metamorphic and plutonic rocks, but were now eroded more deeply into their metamorphic core than during Ordovician time. In Silurian reconstructions, this area of Pennsylvania remains at low southerly palaeolatitudes (Ziegler *et al.* 1979).

Marine and brackish-adapted marine fossils have been found in the Bloomsburg Formation (Hoskins 1961; Giffen 1979), but these are no closer than 24 km from the Lehigh Gap palaeosol. The mountainous source terrane (as reconstructed by Berry & Boucot 1970), was about 80 km to the southeast of the Lehigh Gap clay. This alluvial plain was dissected by loosely sinuous, bedload streams which may have been more sinuous and carried a greater suspended load in bottomlands to the west (Epstein *et al.* 1974). Within this landscape the Lehigh Gap clay formed on high alluvial terraces and interfluves, at least two metres above permanent water table. Less developed soils, with abundant large burrows may have formed in lower lying parts of the landscape, closer to streams.

Time for formation

The Lehigh Gap clay palaeosol is surprisingly well developed for its antiquity. In the scheme of Retallack (1984*a*), the profile as a whole is weakly developed. The calcic horizon is nodular (early stage II of Gile *et al.* (1966), and stage 2 of Wieder & Yaalon (1982)), a degree of development reached after 8000 to 15000 years in the present desert of New Mexico (Birkeland 1984). Considering qualifications also pertinent to interpreting time of formation of the Potters Mills clay, a reasonable time for formation of the Lehigh Gap clay would be of the order of a few thousand years.

Identification

As for the Potters Mills clay, there is petrographic and chemical evidence that the base status of this palaeosol was much higher than in Oxisols of the U.S.D.A. classification. High base status can also be inferred from ferroan dolomite nodules reported from the Bloomsburg Formation in this area by Epstein *et al.* (1974). Although these authors use this composition as evidence against an origin in soils, dolomite has been recorded in modern soils of high base status (Doner & Lynn 1977). Bioturbation of the surface and depth of carbonate are both indications of climate more humid than usual for Aridisols. The Lehigh Gap clay palaeosol appears closest to the same kinds of modern soil as the Potters Mills clay: Oxic Ustropepts (U.S.D.A.), Calcic Cambisols (F.A.O.) and Calcareous Red Earths (C.S.I.R.O.). Because of its better development and more substantial thickness, the Lehigh Gap clay fits more comfortably within each of these categories than the Potters Mills clay.

Other Silurian palaeosols

The palaeosol profile immediately below and partly overprinted by the Lehigh Gap palaeosol (135–185 cm in figure 15) is similar to several others exposed in the cliff near Palmerton. These very weakly developed palaeosols are also similar to un-named palaeosols associated with the late Ordovician Potters Mills clay. Presumably, they formed in lower parts of the landscape, closer to streams. A third kind of late Silurian palaeosol was seen in the dip slopes of Bloomsburg Formation exposed west of the Scenic Overlook off westbound Interstate highway 84, 4 km east of Port Jervis, New York. These also are riddled with subhorizontal and vertical burrows of comparable diameter, but the burrow fill and surface of the palaeosols is greenish grey (5GY6/1 to 5G6/1) and their subsurface horizons are dusky red (10R3/2). These differences in colour may be due to more frequent waterlogging, and perhaps also, marine influence.

Red palaeosols with calcareous nodules of late Silurian age have been reported also from Nova Scotia (Boucot *et al.* 1974).

Peas Eddy clay palaeosol (late Devonian)

Diagnosis

Thick (1 m after compaction), weakly developed clayey palaeosol, with a greenish grey (5BG5/1), clayey A horizon and weak red (10R5/2) to dark reddish grey (10R4/1) B horizon, penetrated by large root traces (figure 16, plate 3).

Location

This palaeosol is below a shaly parting with root traces and impressions of *Archaeopteris* leaves, in the low wall of rock on the outside (southern) of the bend in old New York highway 17 (figure 17), 3.4 km east of Hancock, and 1 km directly across the river from Peas Eddy, Delaware County, New York. On U.S. Geological Survey 7.5 minute, Fishs Eddy Quadrangle, this is at latitude 41° 57′ 23″ north and longitude 75° 14′ 32″ west.

Geological setting

By the mapping of Fisher *et al.* (1961) and Rickard (1975), this palaeosol is in the upper Walton Formation of the West Falls Group. Rodgers (1970) regards this hilly region of central New York state, the Catskill Mountains, as a distinct physiographical and structural province of the Appalachian Orogen.

Description of plate 1

Potters Mills clay palaeosol, late Ordovician, central Pennsylvania.

Figure 3. Road cutting on U.S. highway 322 4 km east of Potters Mills. Arrow indicates top of described palaeosol. Scale is marked in feet (1 foot ≈ 30 cm).

Figure 4. Partly overlapping vertical burrows (*Skolithos*) with sheathing carbonate (light coloured quasicalcitan) and clay fill (at arrow), from horizon Ck. Scale bar is 1 cm, and specimen number UOR182C.

Figure 5. *Skolithos*, vertical burrows (para-aggrotubules) from horizon 2Ck (= A horizon of underlying palaeosol). Scale bar is 1 cm, and specimen number UOR179B.

Figure 6. *Planolites*, subhorizontal burrows (para-aggrotubules) from A horizon. Scale bar is 1 cm, and specimen number UOR182B.

Figure 7. Petrographic thin section under crossed nicols of burrow-sheathing carbonate (quasicalcitan), from horizon Ck. Scale bar is 1 mm.

Figure 8. Petrographic thin section under crossed nicols, showing burrow-sheathing carbonate (quasicalcitan, lower left) and indistinct granular fabric of para-aggrotubule (at small arrow), from horizon Ck. Scale bar is 1 mm.

Description of plate 2

Lehigh Gap clay palaeosol, late Silurian, eastern Pennsylvania.

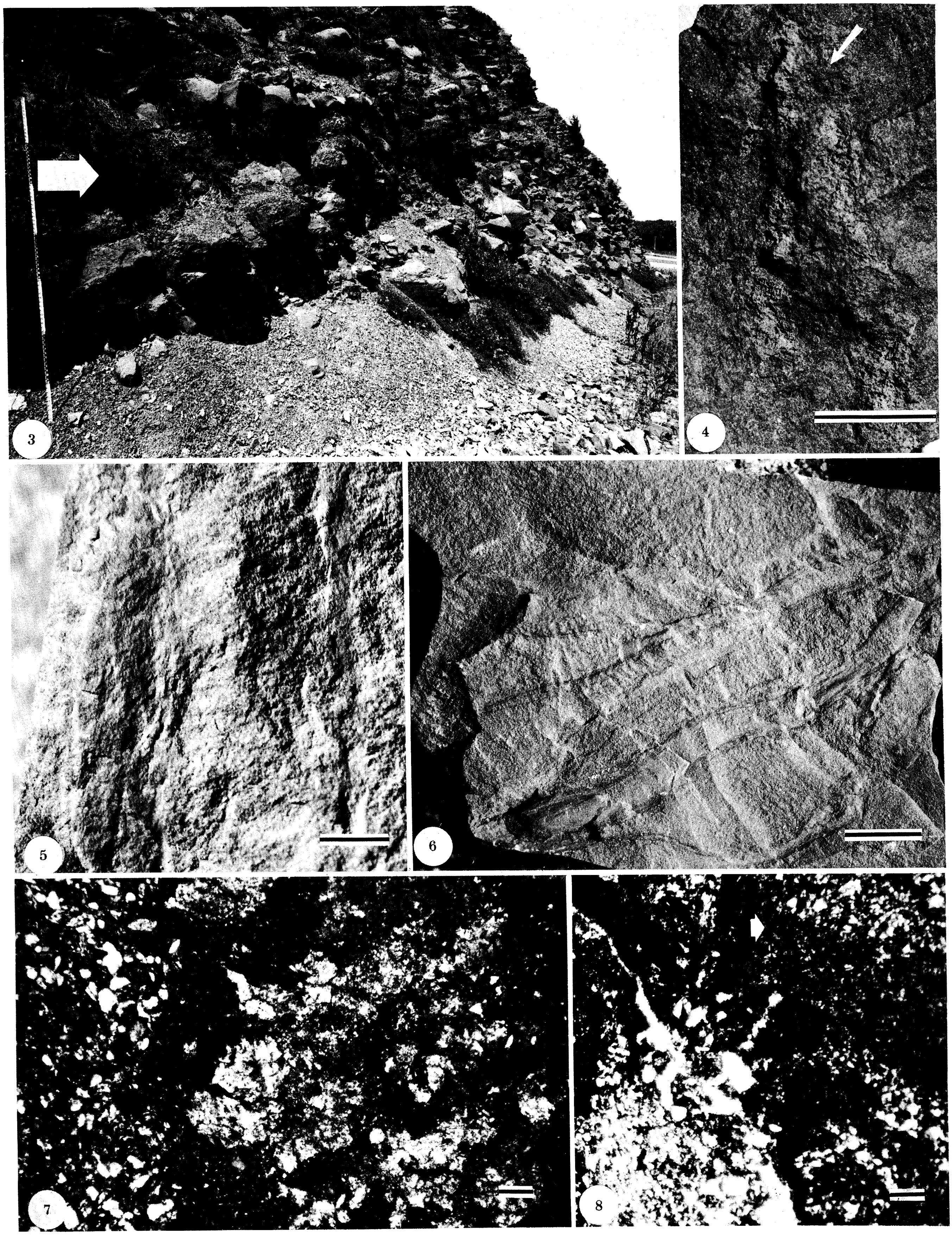
Figure 9. Steeply dipping palaeosol in road cutting near Palmerton. Arrows indicate tops of horizons A (top left) and 4Ck (lower right), both obscured by prominent fracture cleavage. Hammer indicates scale.

Figures 10 and 12. Large (at large arrow) and small (at small arrow) burrow (?) or rhizome (?) traces (ortho-isotubules and metagranotubules), on plane parallel to bedding (figures 10 and 12) and on fracture cleavage plane (figure 12), from A horizon. Scale bars are 1 cm, and specimen number UOR155C.

Figure 11. Petrographic thin section under plain light of burrow (?) or rhizome (?) trace (ortho-isotubule), here irregular and filled with material more clayey (darker) than matrix, from A horizon. Scale bar is 1 mm.

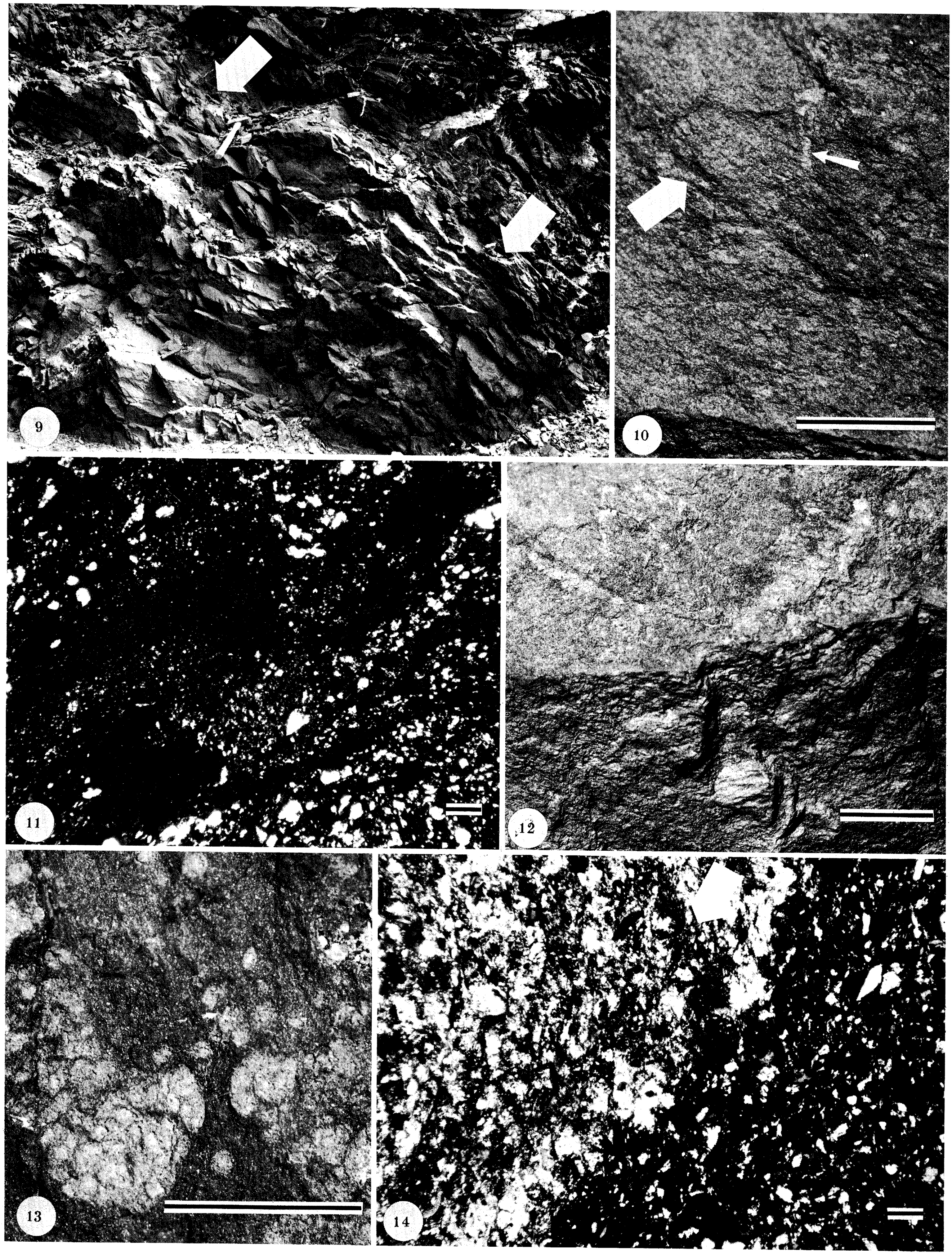
Figure 13. Caliche nodules occur both isolated and forming tubular aggregates, within horizon Ck. Scale bar is 1 cm, and specimen number UOR157A.

Figure 14. Petrographic thin section under crossed nicols of recrystallized caliche nodule (upper left), with zone of red stain within margin (quasisesquan), from horizon Ck. Scale bar is 1 mm.

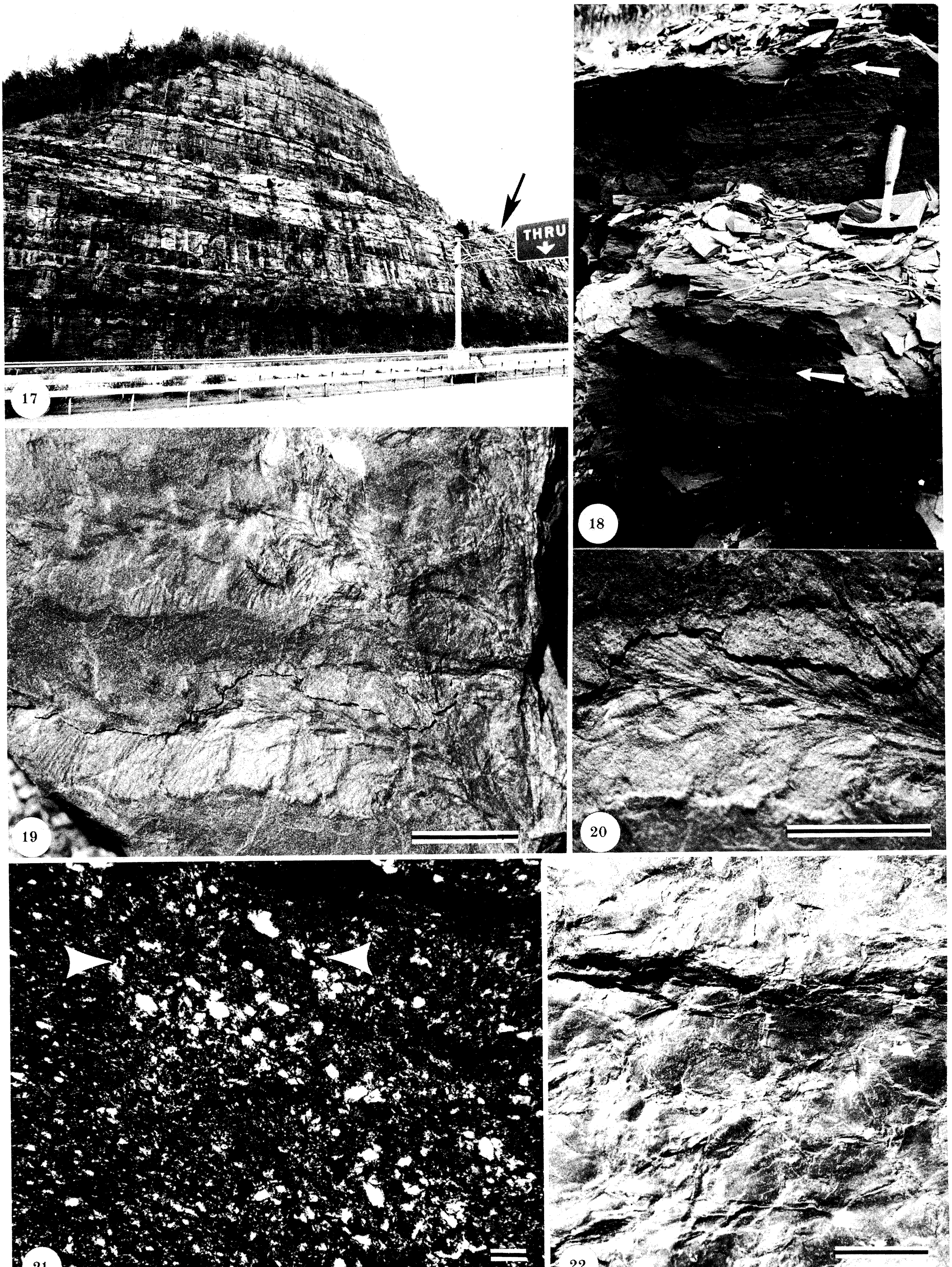


FIGURES 3–8. For description see opposite.

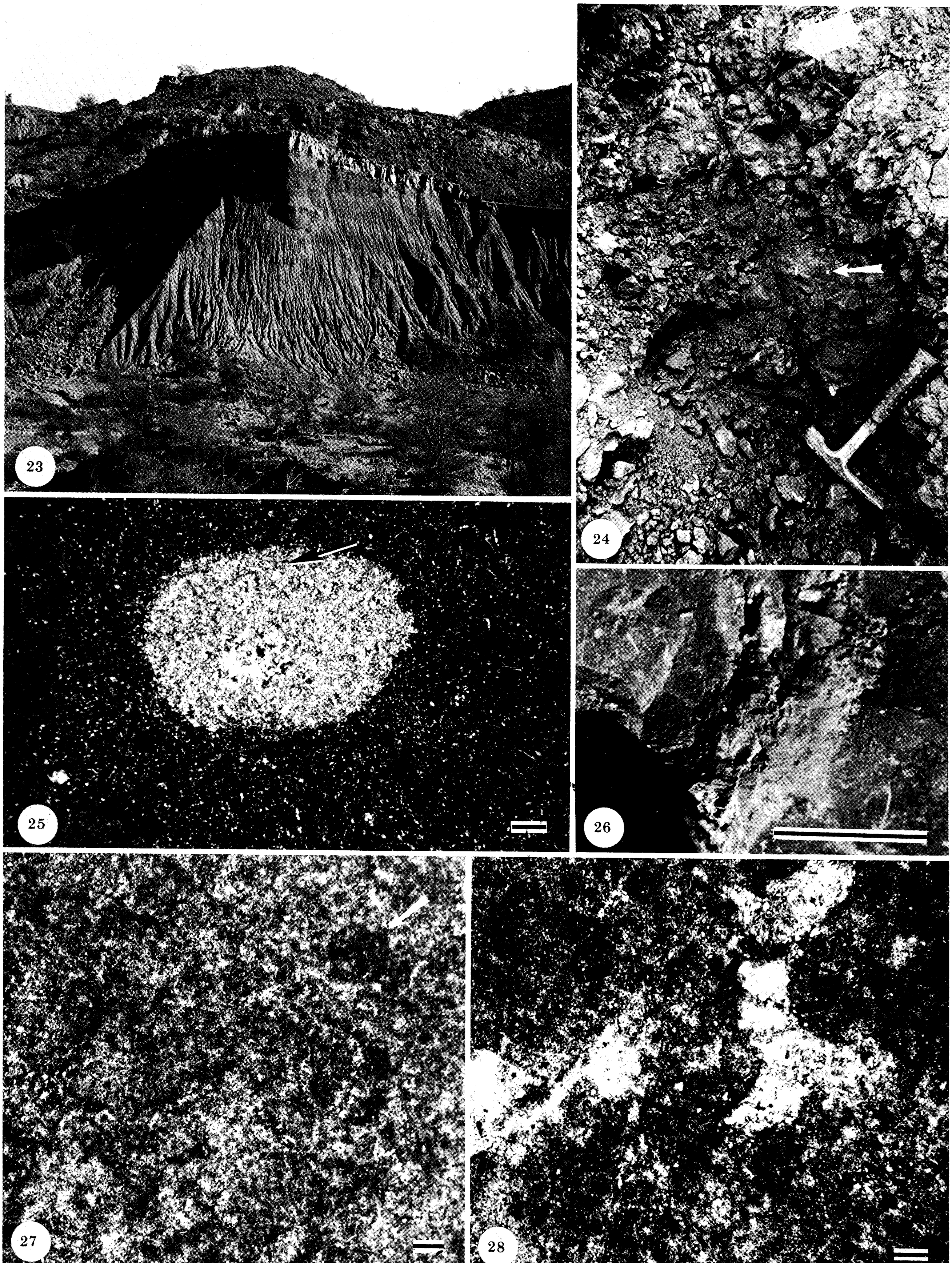
(*Facing p.* 118)



FIGURES 9–14. For description see p. 118.



FIGURES 17–22. For description see p. 119.



Figures 23–28. For description see opposite.

Age

The Walton Formation is early late Devonian (Frasnian) in age (Rickard 1975), or some 367–374 Ma old in Palmer's (1983) radiometric time scale.

Alteration after burial

Studies of the metamorphic mineralogy and coalification of fossil wood and kerogen in the stratigraphically lower, middle Devonian, Gilboa Formation in this region (Friedman & Sanders 1982) and of alteration of conodonts in correlative marine limestones (Epstein *et al.* 1977), all indicate burial temperatures in the range 160–210 °C and burial depths of 4.6–6.6 km. Analysis of the degree of annealing of fission tracks in apatite and zircon in earliest late Devonian sandstones of this same area (Lakatos & Miller 1983) gave comparable results: burial temperatures before unroofing some 125 Ma ago of 120–200 °C and burial depths of 4–7 km. This is the realm of very low grade metamorphism, comparable to that postulated for the Potters Mills clay palaeosol. This degree of metamorphism is surprising considering the mild deformation of these rocks (Rodgers 1970), and the likelihood that much of the chlorite in these rocks is detrital in origin (Liebling & Sherp 1976, 1980).

The abundant concavo-convex and sutured contacts of quartz grains in the lower part (horizon 5C) of this palaeosol are compatible with compaction of sandstones noted elsewhere

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Description of plate 3

Peas Eddy clay palaeosol, late Devonian, south-central New York.

Figure 17. Roadcuttings on new (foreground) and old (at arrow) New York highways 17, viewed from south. Arrow also indicates position of described palaeosol.

Figure 18. Peas Eddy clay palaeosol, sparse root traces (at arrow) and its conspicuous relict bedding. Horizon A is the shale on top and the hammer rests on horizon Bt.

Figures 19 and 20. Fossil leaf of *Archaeopteris halliana* (Arnold 1939), poorly preserved and with marking attributed to partial decomposition in leaf litter, from A horizon. Scale bars are 1 cm, and specimen number UOP6641.

Figure 21. Petrographic thin section under crossed nicols of root trace with associated organic matter (black) and filled with material more silty than matrix (between arrows), from A horizon. Scale bar is 1 mm.

Figure 22. Poorly preserved subhorizontal root traces from A horizon. Scale bar is 1 cm, and specimen number UOR148.

Description of plate 4

Lal clay palaeosol, late Miocene, northern Pakistan.

Figure 23. Badlands overlying the Lal clay, showing red palaeosols associated with buff-coloured sandstone palaeochannels, and scattered among yellow to brown palaeosols away from palaeochannels. Stratigraphic thickness of section between prominent sandstone capping badlands and sandstone forming foreground is 13.4 m.

Figure 24. Excavated Lal clay paleosol, showing top of A horizon (large arrow) and drab haloed root traces (for example at small arrow). Hammer indicates scale.

Figure 25. Petrographic thin section under crossed nicols of small caliche nodule, with a zone of red stain within margin (quasicalcitan), from horizon Bt. Scale bar is 1 mm.

Figure 26. Freshly exposed, large, drab-haloed root trace. The central dark portion is black (5BG2.5/1) flanked by halo of greenish gray (5GY6/1) within a matrix of dark reddish brown (5YR3/4). The specimen is from horizon 3Co (= A horizon of underlying palaeosol). Scale bar is 1 mm.

Figure 27. Petrographic thin section under plain light of inundulic sesquioxidic claystone, vomasepic in association with root traces and showing spherical micropeds (for example at arrow), from horizon ABs. Scale bar is 1 mm.

Figure 28. Petrographic thin section under plain light of burrows of soil insects (meta-isotubules), from A horizon. Scale bar is 1 mm.

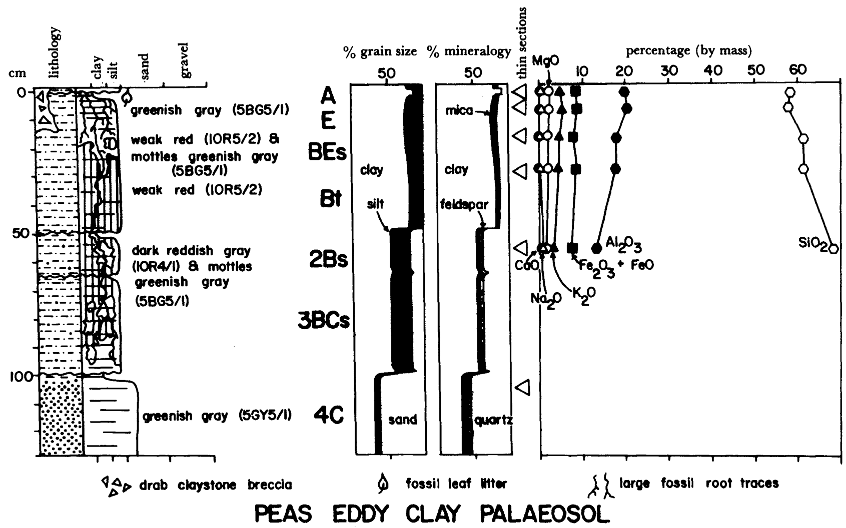

FIGURE 16. Field appearance, petrographic and chemical composition of the Peas Eddy clay palaeosol, late Devonian, south-central New York state. Specimen numbers from top down are UOR148 to 153. Unspecified symbols are as for figure 2. Horizontal axis is percentage (by mass).

(Taylor 1950) at burial depths like those proposed for this region. Root traces in the upper portion of the palaeosol appear flattened and deformed into concertina outlines. Also produced by compaction and metamorphism is the strong reorientation of clay (omnisepic microfabric) in the upper portion (horizons E to Bt). There was probably significant compaction of this clayey part of the palaeosol.

Barrell long ago (1913) advanced an explanation for the red colour of these rocks, similar to the dehydration of less red, iron oxyhydrates, proposed by Walker (1967) for Pleistocene palaeosols. The drab colour of the upper portion of this palaeosol may not have been original either. Light-coloured surface horizons in modern soils may form by leaching of iron and clay to subsurface horizons (Birkeland 1984). There is some textural and chemical evidence (figure 16) that this has occurred to a limited extent in the Peas Eddy clay palaeosol. Drab hue can also be developed in surface horizons shortly after burial because of bacterial reduction of remnant organic matter, as is likely for drab-haloed root traces in this and other palaeosols (Retallack 1983*b*). An additional explanation for drab-topped palaeosols is apparent from their suspiciously frequent occurrence immediately underlying sandstone palaeochannels in late Devonian rocks of New York state (Friend 1966). These palaeochannels commonly contain pyritized logs. Bacterial reduction of this organic matter in either the poorly oxygenated base of the stream or after its burial may have been sufficient to reduce underlying red claystone as well (Friend 1966). Comparable diagenetic pseudogley has been proposed to account for minerals in Carboniferous palaeosols (Roeschmann 1971). Aware of this problem, I chose the Peas Eddy clay as the most likely palaeosol to have had originally a light-coloured surface horizon, because it is separated from a palaeochannel by another drab palaeosol 42 cm thick. Despite this separation from the palaeochannel and textural evidence for slight lessivage, it remains likely that the surface horizon of the Peas Eddy clay was greyish brown, or some similar warm colour, rather than its present greenish grey.

Reconstructed soil

The original Peas Eddy clay was probably a brownish grey clay, darker and more broken by roots near the surface (to 14 cm in the greatly compacted upper portion of the palaeosol) and with prominent reddish brown mottles and more conspicuous sedimentary relicts at depth (now to 96 cm). Large changes in grain size within the palaeosol are relict from the sediment on which the soil formed. The surface shale is clearly bedded, and probably represents a cumulic horizon, added by flood waters restrained by trees in the soil. Unlike many late Devonian palaeosols in New York, the Peas Eddy clay is non-calcareous. Thus, pH was probably neutral to acidic throughout the profile. This difference from associated palaeosols may have been due to the acidifying effect of more lush vegetation on the Peas Eddy clay and due to neutralization from a shallow water table near streams. Evidence of waterlogging includes the preservation of leaf impressions in the surface, the purple tinge of the ferruginized horizon (perhaps due to organic matter; Bown 1979; Retallack 1983*b*) and the drab colour and little weathering observed in the basal sandstone (horizon 4C). Water table probably fluctuated within the limits of horizon 3BCs (64–96 cm in compacted palaeosol). Despite this evidence of waterlogging, there is little development of gley mineralization, such as nodules or crystals of pyrite or siderite. Perhaps, as can be inferred from relict bedding, the time of formation of this palaeosol was short. As in the older palaeosols already considered, formation of this one included clay production from feldspars and ferruginization. Increase in iron content at the expense of magnesium on ancient floodplains of the Walton Formation, has been demonstrated by geochemical studies of Liebling & Scherp (1976). A distinctive new soil forming process, by comparison with older palaeosols, is the leaching of clay to a subsurface (B) horizon, or lessivage (of Duchafour 1982). This process has not proceeded in this palaeosol to the extent that the B horizon qualifies as argillic. In addition, there appears to have been some eluviation of iron, as also noted by Roberson & Eichenlaub (1971) for other parts of the Walton Formation.

Palaeoclimate

Waterlogged, wooded soils of this kind are not diagnostic for climate. Groundwater forests may form even in regionally arid climates, such as in northern Lake Manyara National Park, Tanzania (Douglas-Hamilton & Douglas-Hamilton 1975). Other red, calcareous palaeosols of late Devonian age in New York and elsewhere on the 'Old Red Continent' are evidence of subtropical, semi-arid to subhumid climates (Allen 1973, 1974*a*, *b*; Woodrow *et al.* 1973). Tepee structures (Allen 1973, 1974*b*) and surface cracking in some of these palaeosols (Wells 1969) are indications of variation in rainfall, which may have been seasonal. Because both soils and plants during late Devonian time appear to have been more like those of today, more confidence can be placed in these palaeoclimatic interpretations than in those based on older palaeosols.

Fossil flora

There are several lines of evidence that the Peas Eddy clay was forested: thickness of the profile, differentiation of a B horizon (figure 16), large root traces (figures 21 and 22), a large (30 cm) stump-like hole (cradle knoll) filled with palaeochannel breccia, near the surface of the palaeosol, and impressions of *Archaeopteris halliana* leaves in the surficial shales (figures 19–20). This palaeosol thus confirms Beck's (1964) conclusion based on the abundance of

Callixylon logs and *Archaeopteris* leaves in deposits associated with palaeochannels, that parts of these late Devonian landscapes were 'heavily forested'.

Evidence for forests away from streams is less secure. Those palaeosols in red beds away from palaeochannels that I have examined in New York, all have root traces of lesser diameter than palaeosols associated with palaeochannels, such as the Peas Eddy clay. These former soils of well-drained floodplains are also thinner and calcareous. These preliminary observations confirm Barrell's (1913) view that late Devonian forests formed streamside galleries whereas dry interfluves were colonized by herbaceous or bushy vegetation.

Other fossil soils with small root traces, drab and partly mineralized with pyrite, have been examined in the road cutting opposite the Peas Eddy clay palaeosol; again, these probably bore different vegetation.

The vegetation of these other kinds of palaeosols remains to be established by further collecting. Late Devonian (Frasnian) fossil plants of New York are diverse and have features indicating that they lived in a variety of habitats (Banks 1967; Beck 1957, 1967, 1981; Hueber 1961; Grierson & Banks 1963; Carluccio *et al.* 1966; Bonamo & Banks 1967; Matten & Banks 1966, 1967; Scheckler & Banks 1971 *a*, *b*; Banks & Grierson 1968; Stubblefield & Banks 1978; Serlin & Banks 1978; Hueber & Banks 1979; Matten 1981). Although the community palaeoecology of this ancient vegetation remains to be worked out in detail, it is likely that partitioning of different plant formations in various parts of the landscape had already begun. This vegetation was probably intermediate between low-growing herbaceous and semi-aquatic vegetation of earlier times and the physiognomic and structural diversity of modern vegetation.

Fossil fauna

Burrow-like features (meta-isotubules and metagranotubules: figure 21) near the surface of this palaeosol could have been excavated by soil animals, but both are so irregular that they could equally be root traces. Some of these tubular structures are mantled by remnants of ferruginized organic matter. This does not prove their identification as only root traces, because soil animals may feed on decaying roots.

The *Archaeopteris halliana* leaves forming a fossil litter are poorly preserved (figures 19 and 20). Local creasing and loss of detail on these impressions is similar to that seen in leaves partly decayed by microbes and enveloped by fungal hyphae (Retallack 1976). Perhaps some of the crease-like markings were created by minute soil animals. No definite traces, such as nibble marks, were seen.

A variety of soil organisms had evolved by late Devonian time. Doug Grierson and Patricia Bonamo have recently discovered various small arthropods, including myriapods and arachnids, in middle Devonian rocks of New York state (Shear *et al.* 1984; Rolfe, this symposium). Amphibians are well known in latest Devonian (Fammenian) rocks in Greenland, and trackways from Australia are evidence of an older (Frasnian) origin (Warren & Wakefield 1972). Supposed trace fossil evidence of them from late Devonian rocks of Pennsylvania proved to be mistaken (Caster 1938). Further afield, arachnids and possible insects have been found in the early Devonian (Siegenian) Rhynie Chert of Scotland (Rolfe 1980), and early Devonian (Emsian) Nellenköpfschichten near Alken, West Germany (Størmer 1977).

The paucity of evidence for soil fauna in the Peas Eddy clay palaeosol contrasts with this fossil record of land animals and with the abundant burrows in Ordovician, Silurian and other

Devonian palaeosols. Could it be, as argued by Swain & Cooper-Driver (1981), that *Archaeopteris* possessed phenolic substances which protected it from ultraviolet radiation and were toxic to herbivores, pathogens and decomposers? Their main reason for proposing such a view is the wide distribution of such substances in fossil and living plants. The weakly developed Bt horizon of the Peas Eddy clay palaeosol could be considered additional evidence for this proposal. Plant phenolic substances leached from leaves by rainfall are known to play an important role in translocation of iron (Duchafour 1982; Fisher & Yam 1984). The evolution of leaves, plagiotropic shoots and greater size of *Archaeopteris* compared with its smaller trimerophyte ancestors, would have increased the flux of leachates into the soil with detrimental effects on soil biota. Translocation of iron in the Peas Eddy clay palaeosol is not nearly as extensive as in modern Spodosols (U.S.D.A. classification) formed under species-poor heaths and conifer forests. This may be because the Peas Eddy palaeosol formed over a short period of time, as inferred from its relict bedding. It is also likely that phenolic substances have increased in both effectiveness and concentration through geological time (Swain & Cooper-Driver 1981).

Parent material

Large changes in grain size within this palaeosol (figure 16) are probably sedimentary rather than pedogenic. Its alluvial parent material consisted of several beds of different grainsize, showing an overall upward fining trend.

Palaeogeography

The Peas Eddy clay formed on alluvial outwash of the Acadian Mountains, formed by continental collision of Laurentia and Baltica (Ziegler *et al.* 1979). From the mineralogical composition of sandstones in this alluvium, Allen & Friend (1968) suggested that this range was of Himalayan proportions, and formed largely of low-grade metamorphic rocks, with minor sedimentary and volcanic strata. This part of North America was situated at low southerly palaeolatitudes at this time (Ziegler *et al.* 1979).

As reconstructed from isopach data and other considerations by Barrell (1914), the mountains lay about 160 km east of the Peas Eddy clay palaeosol. The piedmont and alluvial plain was probably only about this wide, because there is evidence of marine influence in rocks associated with the Peas Eddy clay. Palaeosols with pyritic nodules are a few metres stratigraphically above this palaeosol in road cuttings north across the road. Fragments of articulate brachiopods were found low in cuttings on new New York highway 17 several metres stratigraphically below this palaeosol (100 m east from viewpoint of figure 17). The Peas Eddy clay is associated with a palaeochannel sandstone showing a steep and deep, cut bank. Observations such as this and the common shale in this formation, thought to have been deposited in oxbow lakes (Allen & Friend 1968), are indications of sinuous streams of mixed load. Since the Peas Eddy clay appears to have been waterlogged for a part of the year, it probably formed on low levees and swales flanking a coastal stream or deltaic distributary.

Time for formation

In the scheme of Retallack (1984*a*), this palaeosol is weakly developed. Relict bedding is present throughout the profile (figure 18). Although there is discernable differentiation of A and B horizons, this has not proceeded to the point where they qualify as albic or argillic. By using as a modern analogue the development of clayey B horizons in soils on stream deposits in the dry climate of northern Pakistan (Ahmad *et al.* 1977) and of the San Joaquim Valley, California (Arkley 1964; Harden 1982; Birkeland 1984), the B horizon of the Peas Eddy clay would have formed within 10000 years. This seems excessive in view of the prominent relict bedding and lack of pronounced gley mineralization. On spoil dumps of very permeable loess only 100 years old in subhumid Iowa, Hallberg *et al.* (1978) report clay enrichment of 3.5%, somewhat more than for the Peas Eddy clay (1.4%). A time of a few hundred to a few thousand years is likely.

Identification

The B horizon of the Peas Eddy clay palaeosol has not quite enough clay (figure 16: an 8% increase is needed for a soil as clayey as this) to qualify as an argillic horizon and much too much clay for a spodic horizon as defined by the U.S.D.A. Considering also its relict bedding, this palaeosol was probably an Inceptisol. Because of its low chroma and other indications of waterlogging, the palaeosol is most like Aquepts, but it is difficult to decide between Tropaquepts and Haplaquepts. It is more likely to have been a Tropaquept, because these are darker, more colourfully mottled and found in warmer climates than Haplaquepts. This palaeosol is probably not a Gleyic Cambisol in the F.A.O. classification, because even considering compaction, it was probably periodically waterlogged within 50 cm of the surface. It is more like Eutric Gleysols within this classification. In the C.S.I.R.O. classification, the palaeosol is most like a Gleyed Podzolic soil, but it is a very weakly developed example. Humic Gleys of this system have a comparable degree of development, but the Peas Eddy clay has not nearly enough organic matter in its A horizon to qualify as one of these.

These modern soils are formed in low lying parts of the landscape, sometimes on rocky basement, but more usually on alluvium or colluvium of poorly drained depressions or extensive fluvial or deltaic lowlands. Their parent materials are usually of felsic composition. The strong influence of local waterlogging overrides other soil-forming factors. Although found mostly in humid climates, they can form under almost all climates. Their modern vegetation also varies, from swamp forests to sedge marsh and mossy bogs.

Other Devonian palaeosols

Two other New York palaeosols were seen more strongly developed than the Peas Eddy clay. Both are non-calcareous, have drab surface horizons, immediately underlie palaeochannels and are within the Oneonta Formation, also of late Devonian (Frasnian) age. One of these, high in the southwestern end of the road cutting beside northbound Interstate highway 88 1 km west of the western Unadilla exit, has large root traces penetrating a surficial relict clayey bed, as well as an underlying sandstone bed. Another palaeosol in the quarry on the west slope of Cave Mountain, between Ashland and Jewett (fossil locality number 4 of Stubblefield & Banks 1978), shows little in the way of root traces, but excellent soil structure. There are clear slickensided soil clods (peds) near the top of the profile and strong (dusky red, 10R3/2) sesquioxidic stain and nodules below that. Although no megafossils were found in association with these palaeosols, it is likely that they supported forests like the Peas Eddy clay.

Another common kind of palaeosol in late Devonian rocks of New York state consists of calcareous, nodular, red, silty clay, with smaller (less than 1 cm diameter) root traces. Palaeosols of this kind were seen in the Oneonta Formation in road cuttings along U.S. highway

30 on the descent to Gilboa from Grand Gorge, and along Interstate highway 88 northbound near the Unadilla exit mentioned above, as well as northbound just past the Otego exit. These palaeosols are not associated with palaeochannels, and may have supported shrubby or herbaceous vegetation in interfluves.

Just across the road north of the Peas Eddy clay, also in the late Devonian (Frasnian), Walton Formation, a third kind of palaeosol was found. This is drab in its upper part, but its outcrop is stained red in the lower part of the profile from modern oxidation of original pyritic blebs. Similar gleyed sulphidic soils form today under coastal salt marsh and mangrove vegetation (Blilely & Pettry 1979; Coultas 1980).

The best documented late Devonian (probably Famennian) palaeosol from beyond New York is in Victoria Land, Antarctica (McPherson 1979). This palaeosol shows some differentiation of the A and B horizons like the Peas Eddy clay, but unlike the New York palaeosol, the Antarctic one is calcareous. From what can be gathered from published information on other Devonian palaeosols in Pennsylvania (Walker & Harms 1971), Nova Scotia (Boucot *et al.* 1974), the United Kingdom (Allen 1973, 1974*a*, *b*; Parnell 1983) and China (Boucot *et al.* 1982), they are mostly like the calcareous nodular palaeosols of the Oneonta Formation in New York.

Lal clay palaeosol (*late Miocene*)

Diagnosis

Thick (1 m at present) dark reddish brown (5YR4/4 and 5YR3/4) clayey palaeosols, with few surface, large (1–2 cm), drab-haloed root traces, more clayey and studded with small (3–4 mm) sesquioxidic and calcareous nodules below the surface (figure 29, plate 4).

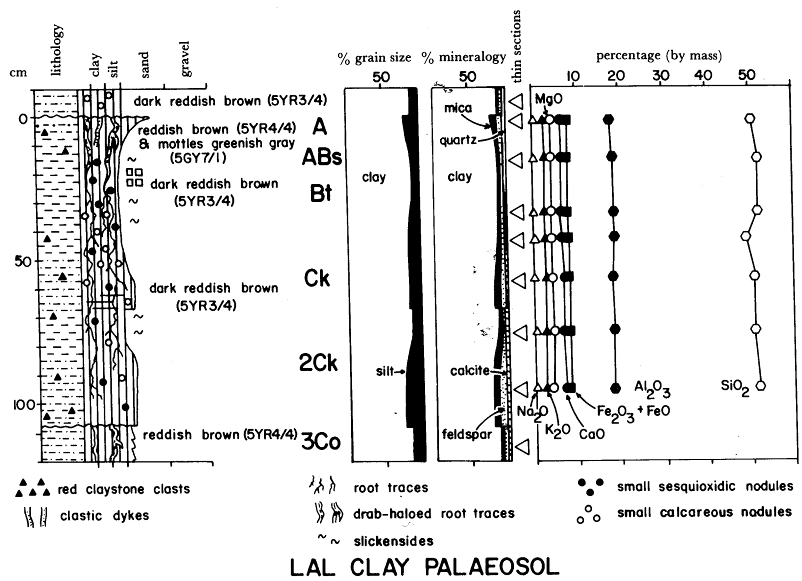

FIGURE 29. Field appearance, petrographic and chemical composition of the Lal clay palaeosol, late Miocene, northern Pakistan. Specimen numbers from top down are USNM353777 to 353785. Unspecified symbols are as for figure 2. Horizontal axis is percentage (by mass).

Location

The Lal clay palaeosol is in badlands (figure 23), at a stratigraphic level 8–9 m above a buff-coloured sandstone exposed in the bed of a tributary of Kaulial Kas, 2 km southwest of Kaulial village, near Khaur, in the Potwar Plateau region of northern Pakistan. On Atock District 1:50000 map number 43C/11 this site is at military grid reference 748144. The name Lal does not refer to a geographical locality. It is Urdu for 'red' and is part of a wider system of unpublished names for other Miocene palaeosols in this area.

Geological setting

This palaeosol is at 2292 m in the stratigraphic section measured by Marc Monaghan in Kaulial Kas (Barry *et al.* 1980), and in the Dhok Pathan Formation of the Siwalik Group. The buff sandstone below the palaeosol is the one marked at the stratigraphic level of 'Middle Gray' in section KUL of the facies map of Behrensmeyer & Tauxe (1982, figure 3). Dips on the buff sandstone here are gentle (16° towards magnetic azimuth 134°). These rocks form the northern limb of the broad Soan Synclinorium.

Age

Fossil mammals at this stratigraphic level are within the upper part of the '*Hipparion s.l.*' Interval Zone of Barry *et al.* (1982) of late Miocene age. The Lal clay is within the upper portion of palaeomagnetic reversed zone DR3 of Tauxe & Opdyke (1982). From their age calibration and assuming near linear sediment accumulation rates (documented for these rocks by Johnson *et al.* 1982), the Lal clay palaeosol is about 8.3 Ma old.

Alteration after burial

This sequence appears little deformed or metamorphosed. About 1 km of overlying sediments have been measured in Kaulial Kas (Barry *et al.* 1980; Tauxe & Opdyke 1982). Applying the same geothermal gradient used for the Palaeozoic Appalachian palaeosols (Epstein *et al.* 1977), gives a burial temperature for these Miocene palaeosols not significantly above modern ground temperatures in this region.

Root traces and clastic dikes in this palaeosol do not appear to have been greatly deformed. In other parts of this sequence, I have observed undeformed clastic dykes up to 2 m deep. Compaction of these palaeosols was probably slight.

The greenish colour of haloes around root traces in these rocks (figure 26) may be due to bacterial reduction of organic matter following burial of the soil (Retallack 1983 *b*). This is the only gley feature seen in this palaeosol, which in other respects appears to have been well drained.

Dehydration and reddening of ferric oxyhydrate minerals during burial (Walker 1967) is also likely. The finding of Tauxe *et al.* (1980) that red pigment in these red beds is magnetized in a different direction to specular haematite of depositional or early post-depositional origin, could be taken as additional support for this view, if it can be shown that soil structure played no role in orienting red pigment.

Reconstructed soil

The Lal clay was originally a thick, red or brown clay, extensively bioturbated and rather uniform in texture throughout the profile (figure 24). Considering its carbonate and colour, it was probably neutral to mildly alkaline and oxidizing. Water table was probably almost always below a metre from the surface. The principal soil-forming processes appear to have been surface accumulation of organic matter, pervasive bioturbation and destruction of relict bedding, clay formation and calcification below the surface, and ferruginization throughout the profile. Sesquioxidic stain has obscured clay microfabric (inundulic to insepic with spherical micropeds), except in areas of root traces (vomasepic) and burrows (silasepic meta-isotubules). The argillic horizon of this palaeosol probably formed by weathering in place, rather than translocation from higher horizons, for the following reasons: an eluvial horizon is lacking; only a few clay skins were seen; finally, increased clay in the B horizon is more than adequately accounted for by losses of feldspar and mica there.

Palaeoclimate

By comparison with modern soils (Birkeland 1984), several features of this palaeosol are compatible with a warm, subtropical to tropical climate: its clayey texture, strong oxidation, and pervasive bioturbation by roots and soil fauna. Reddish hue could also be cited if there were not some doubt about its time of origin. This palaeoclimatic interpretation is supported by the fossil flora and fauna of northern India and Pakistan, and its palaeogeography and palaeolatitude during late Miocene time (Johnson 1977; Sahni & Mitra 1980; Badgley & Behrensmeyer 1980). The diversity of these fossils, their closest living relatives, and the presence of such organisms as palms and crocodiles, are all compatible with a warm climate.

The various features of the Lal clay already cited are also matched by modern soils of rainy climates (Birkeland 1984). A drier climate of perhaps only 20–50 cm mean annual rainfall (perarid thorn woodland zone of Holdridge 1967) can be inferred by comparison with modern soils (Jenny 1941; Arkley 1963), from the shallow depth (20 cm) to calcareous nodules in the Lal clay. This comparison is weak because of the poor development of the calcic horizon in this palaeosol, because of possible compaction and erosion and because studies of depth to calcic horizon in modern soils are based on soils of climates cooler than envisaged for the Lal clay. Significant compaction is unlikely, judging from little-deformed root traces and clastic dikes. Erosion of the profile is probable, considering that its surface has some unusually large and little-weathered grains. Thus a wetter, perhaps subhumid, rainfall regime is likely and can also be inferred from the fossil record of grasses (Nandi 1975) and grazing mammals (Badgley & Behrensmeyer 1980) and the composition of sandstones (Krynine 1937) and other palaeosols (Johnson *et al.* 1981) in equivalent parts of the Siwalik Group elsewhere in India and Pakistan.

Seasonality of climate can be inferred from the co-occurrence of sesquioxidic and calcareous nodules in this palaeosol. Reddening may have been the main soil-forming process during the wet season and calcification during the dry season. The iron-stained bands (neosesquan and quasisesquan), and the small size and near-spherical shape of the calcareous nodules (figure 25), could be due to cyclical dissolution, ferruginization and calcification. A seasonal climate would also have induced the two to three especially prominent growth bands seen in opercula of land snails collected in this area. Extensive, large, clay-filled burrows in cross-bedded sandstone below the Lal clay palaeosol may have formed when streams were dry.

Fossil flora

Considering the size and depth of penetration of root traces and burrows within this palaeosol, as well as its moderate development, and clayey B horizon, it was probably formed under vegetation of considerable biomass, such as rain forest or semideciduous tropical forest. This is the usual vegetation of modern soils with which it is identified.

Similar vegetation is abundantly represented in the fossil record, especially of wood, found in the Siwalik Group in nearby India (Nandi 1975; Johnson 1977; Sahni & Mitra 1980; Singh & Prakash 1980; Prakash 1981; and references therein). Johnson compares this vegetation to modern gallery forests dominated by *Dipterocarpus* and *Anisoptera* in the lowlands of Assam and Burma.

Distribution of other palaeosols around the Lal clay confirm that these forests formed streamside galleries. Each of the sandstone palaeochannels is overlain and flanked by weakly developed palaeosols, and then either gleyed forested palaeosols or several red forested palaeosols, like the Lal clay. Silty claystones of flood plains away from palaeochannels, include a fourth kind of palaeosol. This is yellow to brown, moderately to weakly developed, with granular cracking (peds and cutans) and fine, copiously branched root traces. In many ways these brown palaeosols are similar to Conata Series palaeosols of Badlands National Park, South Dakota, U.S.A. (Retallack 1983*a*, *b*), which also probably supported savanna vegetation. Since there are also a few isolated, interbedded and laterally impersistent red palaeosols like the Lal clay in these flood plain sequences, it is likely that regional vegetation was savanna groveland, in which extensive savanna was broken here and there by patches of forest.

A similar vision of these ancient landscapes was proposed by Badgley & Behrensmeyer (1980) to explain the mixture of mammals adapted to savanna and forest. The more open parts of the vegetation are less obvious from megafossil plants found in the Siwalik Group in India. Since forested habitats were closer to water, and thus to the kinds of reducing environments in which plants are preserved (Retallack 1984*a*), this discrepancy can be explained as preservational bias. It could also be that late Miocene outwash of the Himalayas was more heavily forested further east in India and Nepal, although this was not indicated by Johnson (1977) in his study of palaeosols in India.

Fossil fauna

The most prominent feature of this palaeosol in petrographic thin sections is the abundance of small (1–2 mm), near-spherical areas of claystone. Some of these have sharp margins and qualify as papules (in the terminology of Brewer 1976). Much of the fabric of this palaeosol is made of similar structures of varying distinctness. Uniform texture and predominance of spherical micropeds (figure 27) are micromorphological characteristics of modern Oxisols (Stoops 1983). Such structures are thought to be produced by the activity of soil fauna, especially termites (Nye 1955). Meta-isotubules throughout the profile (figure 28) also were probably excavated by soil insects. Fossil aardvarks in these sediments (Pickford 1978) are independent evidence for termites.

The fossil vertebrate fauna of this area is diverse, including fish, turtles, crocodiles, birds, antelopes, horses, pigs, giraffes, four-tusker and hoe-tusk elephants, rhinoceroses, tragulids, bear-dogs, hyenas, big cats, rodents, aardvarks, chalicotheres, anthracotheres, and apes (Badgley & Behrensmeyer 1980). Establishing which of these vertebrates lived in the environment in which the Lal clay palaeosol actually formed is not a simple task. All the most productive vertebrate localities are either in stream channel deposits or in associated weakly developed palaeosols of levées (Badgley & Behrensmeyer 1980; Pilbeam *et al.* 1980). Better developed palaeosols such as the Lal clay yield fossils in such low quantities that recognized 'localities' in flood plain claystones are of large extent and so include many different kinds of palaeosols. It is likely that most of these soils were not sufficiently alkaline for the preservation of bone (Retallack 1984*a*). Until better collections are obtained it is best to assign organisms to habitats based on interpreted adaptations and ecology of related living creatures, as has been done by Tatersall (1969*a*, *b*), Badgley & Behrensmeyer (1980), and Gaur & Chopra (1983). Ramapithecine primates may have been restricted to habitats on soils like the Lal clay, considering that these are now thought to be closely allied to orangutans (Pilbeam 1982; Lipson & Pilbeam 1982). Setting aside the aquatic component, most of the other animals probably roamed through gallery forests of the Lal clay. Many of them, especially horses and antelope-like creatures with high crowned teeth, may have ranged farther from streams into grassy interfluves.

Parent material

The Lal clay is part of a sequence of palaeosols all so deeply weathered that little idea of their parent material can be gained from them. Clasts of red clay in some of the palaeosols may be pieces of other soils which contributed parent material to these palaeosols. The least weathered material near the Lal clay is a channel sandstone. Behrensmeyer & Tauxe (1982) have noted conspicuous signs of weathering in similar buff-coloured sandstones. Within the Lal clay there is a notable increase in clay within the B horizon at the expense of feldspar and mica. Although these minerals were probably more abundant in the parent material of these palaeosols, most of them had already been removed in a prior cycle of weathering in soils of the drainage basin upstream.

Palaeogeography

The Siwalik Group is the alluvial outwash of the Himalaya, Karakorum and Hindu Kush mountains, which formed during collision of Indian and Asian continental masses (Parkash *et al.* 1980). These mountains and the Potwar Plateau were probably not so high during late Miocene time as they are at present (Keller *et al.* 1977). The whole Potwar Plateau continues to be elevated and thrust southward over a decollement which comes to the surface in the Salt Range (Opdyke *et al.* 1982). During the Miocene, the Lal clay palaeosol was close to its present distance of about 100 km from the mountain front, and well inland (more than 800 km) from the sea (Sahni & Mitra 1980). Behrensmeyer & Tauxe (1982) have reconstructed local Miocene palaeogeography at the stratigraphic level of the Lal clay. This palaeosol probably formed alongside a system of small streams draining one of the foothill ranges of the high mountains. These streams were moderately sinuous and transported weathered alluvium into a forerunner of the modern Indus River. This major drainage reached well into the high mountains to the north and was choked with little-weathered sands derived from sparsely vegetated metamorphic and crystalline highlands. Drab sediments and soils of these alluvial bottomlands were probably waterlogged and heavily vegetated.

Time for formation

In the scheme of soil development used by Retallack (1984*a*), the Lal clay palaeosol is moderately developed. In soils developed on alluvium in the dry climate of northern Pakistan (Ahmad *et al.* 1977) and the San Joaquim Valley of California (Arkley 1964; Harden 1982; Birkeland 1984) clayey B horizons comparable to that of the Lal clay appear to take more than 10000 years to form. On similar parent materials in the subhumid climate of Iowa (Parsons *et al.* 1962), greater textural differentiation occurred within 14000 years, and in one case within only 2000 years. Other estimates may be gained by considering the weakly developed (lowest stage II of Gile *et al.* (1966) and stage 2 of Wieder & Yaalon (1982)) calcic horizon of the Lal clay. In the intermontane desert of New Mexico this would take from 8000 to 15000 years to form (Birkeland 1984). An average sedimentation accumulation rate of 0.5 m per 1000 years for this sequence has been calculated from palaeomagnetic data (Behrensmeyer & Tauxe 1982; Tauxe & Opdyke 1982). The 2000 years of soil formation implied by this estimate for the Lal clay should be regarded as a minimum, because palaeosols like the Lal clay are better developed than others within the sequence for which this average rate was calculated.

Identification

The Lal clay has several outward features of Oxisols of the U.S.D.A. It is very red, thick and uniform in texture. Boundaries between horizons are gradational and there are few sedimentary relicts. However, its B horizon does not qualify as an oxic horizon because of the persistence of feldspar and mica, and because of high values of MgO, CaO and K_2O (figure 29). These elements, which are major soil cations, are also too high for Ultisols, such as the Sassafras sandy loam soil (of Marbut 1935) and the soil described by Ahmad & Jones (1969). Among Alfisols, the Lal clay has deeper and more abundant root traces than Xeralfs, is too calcareous for Udalfs and not red enough for Rhodustalfs. The Lal clay is most like Oxic Haplustalfs, which form an intergrade between Alfisols and Ustoxes.

Within the F.A.O. classification, the Lal clay also shows affinities with deeply weathered intertropical soils such as Ferralsols and Eutric Nitosols. Considering its presumed high base status, it is best identified as a Calcic Luvisol.

Uniform clayey soils of this kind are classified as Euchrozems, Xanthozems and Krasnozems by the C.S.I.R.O. in Australia. The Lal clay was most like a Euchrozem, considering its calcareous nodules. The other soils are acidic and non-calcareous.

These modern soils are all wooded or forested and form in subhumid, seasonal or monsoonal climates. In warm climates they may form on undulating plains and bottomlands of felsic sediments and rocks. On mafic volcanic rocks, these soils may form in a greater variety of positions on the landscape and within cooler and drier climates.

Other Miocene palaeosols

Although few late Tertiary palaeosols have been described in detail (Johnson 1977; Johnson *et al.* 1981; Retallack 1982), enough is known to infer that Miocene soil profiles were nearly as varied as those of today. Only the other palaeosols at a similar stratigraphic level to the Lal clay are considered here, as an analogue for the geologically older palaeosols already considered.

Within a 60 m stratigraphic interval above the buff sandstone below the Lal clay, five additional kinds of palaeosols were found. Since these share the same fossil fauna and do not appear to be separated by major unconformities, all are considered to have formed in different parts of the late Miocene landscape of this area. Intimately associated with buff palaeochannels of foothills streams are palaeosols with numerous root traces and burrows, and obvious sedimentary relicts, such as bedding. This first kind of palaeosol probably supported well-drained, stream-side, early successional vegetation. A second kind of palaeosol associated with buff palaeochannels is brown to yellow and has a prominent zone of manganese staining below the surface. Sedimentary relicts persist in these palaeosols, which may have supported waterlogged early successional vegetation. As already discussed, the third kind (the Lal clay and similar palaeosols) probably formed under well-drained gallery forest, flanking foothills streams. The fourth kind, farther from buff palaeochannels is yellow to brown, and weakly to moderately developed. These palaeosols have abundant fine root traces, some evidence of clay illuviation, numerous sesquioxidic nodules and some calcareous nodules. They are interbedded with laterally impersistent palaeosols similar to the Lal clay, and probably formed under savanna, interrupted by groves of forest. Associated with blue-grey sandstones thought to have been deposited by a forerunner of the Indus River, is a fifth kind: grey to faintly pink, clayey palaeosols, with large well-preserved root traces. Also associated with blue grey sandstones is a sixth kind: silty, grey palaeosols with abundant calcareous nodules up to 2 cm in diameter. These both appear to have formed under forest, on streamside levees and seasonally dry swamps.

Discussion

Before proceeding with wider implications of the palaeosols for the evolution of land plants, it is well to consider the degree to which soil forming factors other than organisms were comparable. In other words, can these palaeosols provide reasonable data for deducing the biological processes associated with the colonization of land by plants? Time for formation and regional palaeoclimate were roughly similar in each case. Parent materials were also comparable, although there are noticeably more rock fragments in the Ordovician and Silurian palaeosols, and more feldspar in Devonian and Miocene ones. Although all formed on alluvial lowlands adjacent to large fold mountain ranges, there were differences in their geomorphological position within the outwash, and in the size of these mountain ranges. The Himalayan ranges during Miocene time were probably comparable to the late Devonian Acadian Mountains. Both ranges became more impressive in later geological periods. The Ordovician Taconic Mountains were less extensive, and the late Silurian source for the Bloomsburg Formation was only an eroded remnant of the Taconic Mountains. Miocene and Silurian examples appear to have been the best drained of the four palaeosols, and the most closely comparable in their local geomorphological setting. The Ordovician palaeosol may have formed in a low-lying area, adjacent to streams. The Devonian palaeosol was waterlogged for some part of the year. It probably formed in the levée-and-swale topography flanking a near-coastal stream. Although non-biogenic soil-forming factors for the palaeosols selected are fairly similar, they are not identical. This unevenness and the paucity of data preclude a quantitative study of biological activity at the time of formation of these palaeosols. It is to be hoped that more and better chosen palaeosols of future studies will isolate more clearly changes in purely biological soil-forming factors over geological time.

Despite these reservations, the palaeosols discussed allow formulation of a broad timetable for colonization of dry land by large plants. Terrestrial microorganisms were probably present during the Precambrian (Barghoorn 1977; Siegel 1977; Campbell 1979; Gay & Grandstaff 1980; Retallack 1981). From evidence discussed here, well-drained soils were probably vegetated by megascopic non-vascular plants at least by late Ordovician (Ashgillian) time, and by vascular plants during the late Silurian (Ludlovian). Waterlogged and low-lying parts of the landscape were forested by late Devonian (Frasnian) time. Judging from my preliminary observations of palaeosols in the Maunch Chunk Formation (Pennsylvania) and the Calciferous Sandstone near Cockburnspath (Scotland), the afforestation of dry soils was achieved at least by the early Carboniferous. Well-drained forested palaeosols of early Triassic age have been described (Retallack 1976, 1977) and there are also published indications of Permian examples (Retallack 1981).

Judging from the palaeosols described here, animals colonized dry soils by late Ordovician (Ashgillian) times. Some of these animals must have been large for soil-inhabiting invertebrates (3–16 mm diameter), and formed permanent dwelling burrows in low, but still well drained, parts of the landscape. Similar animals persisted into late Silurian (Ludlovian) time, but evidence of animals in better drained soils of high terraces at that time is equivocal. Evidence for soil animals in the late Devonian (Frasnian) palaeosol is unconvincing. Judging from fossils of terrestrial arthropods (Størmer 1977; Rolfe 1980; Shear *et al.* 1984) several kinds of soil animals had evolved by the late Devonian. Their exclusion from early forested soils may have been because of phenolic toxins leached from the canopy, as proposed by Swain & Cooper-Driver (1981) and supported by evidence of mild B horizon development found in the palaeosol described here. Abundant faecal pellets have been described from a fossil Rendzina (an F.A.O. name) of early Carboniferous age in Wales (Wright 1983) and from late Carboniferous Histosols (a U.S.D.A. name) in Kentucky (Scott & Taylor 1983). Faecal pellets filling burrows with the characteristic external morphology of those excavated by earthworms have been described from early Triassic Fibrist palaeosols (U.S.D.A. name: Retallack 1976, 1977). By Miocene time, soil fauna was present in great variety and abundance, almost totally modifying the microfabric of soils, such as the late Miocene palaeosol described here. Termites were probably the main cause of the spherical micropeds observed in this palaeosol. Recognizable nests of termites (Bown 1982), bees and dung beetles (Retallack 1984*b*) have been found in Eocene and Oligocene palaeosols elsewhere.

Numerous causes have been advanced to explain the invasion of land by plants (Boucot & Gray 1982), but the few palaeosols studied here are inadequate to allow choice between competing hypotheses. Studies of long sequences of palaeosols may shed some light on changes in sea level, volcanic activity and climate associated with colonization of land. Studies of palaeosols from higher palaeolatitudes and different climates are needed to assess whether large land plants appeared there earlier than in seasonally dry, intertropical regions, where the palaeosols described here formed.

What little is known of Palaeozoic palaeosols casts some doubt on whether the causes are to be found in extrinsic geological factors alone. There is some support for Gray & Boucot's (1977) concept that the land was first colonized by micro-organisms, then by larger non-vascular plants, and then vascular plants. These were followed by forests and ultimately by more complex vegetation, such as tropical rain forest. The advent of large plants on land is not seen as an especially momentous episode, when they pioneered sterile, bare earth. Large plants and animals appear to have been additions to pre-existing soil biota. Indeed, there are reasons to suspect that large land plants evolved from unicellular soil algae independent of multicellular aquatic plants (Stebbins & Hill 1980). Just as ground disturbed by floods, landslides and human construction is vegetated now by a succession of increasingly complex and bulky communities, so the colonization of land by large plants and animals may have depended as much upon the availability of prepared ground as upon the appearance of terrestrial adaptations. Ecological succession is rapid nowadays because organisms adapted to appropriate conditions are already here. On early Palaeozoic landscapes, where not only the adaptive features but also the environments which would select for them were rare, colonization of the land would have been a very slow process. If the analogy with ecological succession can be extrapolated further, then locations favouring the development of early land vegetation would be those least disturbed by erosion, sedimentation, mountain building, volcanism and changes in sea level or climate. Evidence for such a view should be sought in early Palaeozoic palaeosols formed on bedrock unconformities, as well as within red bed sequences which appear to have accumulated at very low rates.

A surprising discovery of this study is the existence of burrows of sizeable (3–16 mm diameter) animals in late Ordovician (Ashgillian) palaeosols. I am aware that many scientists (for example, Cotter 1982) regard these burrows as convincing evidence for a marine palaeo-environment of rocks of this age, so I here review briefly the evidence for considering these rocks to be ancient soils. The concentration of burrows toward the top of the profile indicates at the very least, an hiatus in sedimentation. This, as well as the gradational alteration and obliteration of sedimentary structure down from that surface, are features of both soils and some marine omission surfaces. The warm hue, now red, of the bed indicates oxidation, which is most common on land. Nevertheless, Palaeozoic red beds containing unquestioned marine fossils are known (Hoskins 1961; Ziegler & McKerrow 1975), and it could be that marine red beds derived from more easily eroded land were more common during Palaeozoic times than they are today. More definitive evidence is the surficial increase in clay at the expense of mica and rock fragments in this bed, difficult to explain other than as soil alteration. Similarly the calcareous nodules of this and the late Silurian palaeosol, although recrystallized, are more like soil caliche of the late Miocene palaeosol, than any marine or lacustrine hardgrounds or nodular beds of my knowledge. Lacustrine and marine carbonates may be nodular and burrowed (Jenkyns 1974; Kennedy & Garrison 1975; Tucker 1978), but tend to form limestone caps rather than small dispersed subsurface carbonate aggregations. Carbonate in marginal marine parts of the Juniata Formation is also more massive (Thompson 1970*b*). Marine and lacustrine carbonates are also associated with shelly fossils, and nothing like this has been found in the Potters Mills clay palaeosol or associated sediments. The sea is likely to have been at least 200 km to the northwest (Meckel 1970; Dennison 1976) at the time this palaeosol formed. Some drab shale beds are interbedded with red beds of the Potters Mills clay and similar palaeosols. These are unvarved, non-calcareous, non-carbonaceous and barren of fossils (including palynomorphs, J. Gray, personal communication). If these drab units were deposited in ponds or lakes, then these were probably short lived and shallow. Most of the sedimentary rocks associated with the Potters Mills clay are sandstone with prominent fluvial features, such as breccias at the erosive bases of channel-like beds, trough cross-bedding and consistent palaeocurrent directions (Yeakel 1962).

What were animals doing out on dry land so early? The burrows are abundant and well preserved. They appear to have been maintained up until the time the palaeosol was covered over. They do not appear to have been later additions to a submerged soil, because the degree of mineral weathering corresponds to their density and because soil carbonate has precipitated preferentially around the burrows (as a quasicalcitan), perhaps in response to pH changes from associated microbial activity or respired carbon dioxide. The burrows are filled with what appear to be faecal pellets, although the preservation of these structures is poor. For these reasons, the idea that the animals were using these burrows as temporary resting places seems unlikely. Other explanations all rely on the organisms having something to eat and do. What kind of animals they were, and whether they were deposit-feeding, detritus-feeding, composting, farming or hunting, it is difficult to say at present. All imply a substantial organic component of the soil. Just as soil microbes or non-vascular plants may have prepared the way for these animals, these burrows were potentially important microenvironments for other colonists. In modern deserts, the burrows of rodents may be small, semi-autonomous communities, including algae, fungi, herbivorous and dung beetles, all protected within the moist, cool burrow from the harsh external environment (Halffter & Matthews 1966; Martin & Bennett 1977). The role of burrows as possible stepping stones for the greening of Palaeozoic landscapes needs to be investigated with additional and more detailed studies of trace fossils in red beds.

Although the causes of the invasion of the land are not yet clear from the evidence of the palaeosols, some consequences are apparent. Schumm (1968, 1977) has speculated that before the advent of land plants, the landscape would have been highly unstable and prone to flash flooding and sheet erosion after rains. In such an environment, mineral grains loosened from bedrock by physical and chemical weathering would be unlikely to be weathered to clay in place, but would be eroded and deposited. Stabilization of the landscape by vegetation would result in greater production of clay and increased weathering of bedrock. This hypothesis is confirmed in general by the palaeosols described here, because the geologically younger palaeosols are increasingly clayey. There is also evidence from the lower horizons of these palaeosols and from associated palaeochannels, that source terrain during Miocene time was much more deeply weathered than during the early Palaeozoic. Nevertheless, the late Ordovician palaeosol described is much more clayey than Schumm's hypothetical initial state. If this state were ever widespread on earth, it was probably well back into Precambrian time, considering the clayey nature of many Precambrian palaeosols (Button 1979; Gay & Grandstaff 1980; Retallack 1981) and the changing amount (Ronov *et al.* 1980) and base status (Weaver 1969) of clays in the sedimentary rock record over geological time. Because there are sandy, texturally immature, soils on the present landscape, a temporal trend in clay production and leaching in soils can only be satisfactorily demonstrated when more palaeosols are known.

Increased soil structure is also apparent in palaeosols of younger age. Despite the burrows, persistent sedimentary relicts are found in the late Ordovician palaeosol. Wispy bioturbation disrupts bedding in the upper 25 cm of the late Silurian palaeosol. Root traces of the late Devonian palaeosol extend through faint relict bedding to depths of almost a metre. The late Miocene palaeosol is so totally bioturbated to a comparable depth, that barely a trace of original bedding remains. Soil microstructure shows parallel increased development (from skelinsepic to vomasepic), thus confirming the importance of bioturbation in promoting soil microstructure (as proposed by Brewer & Sleeman 1969; Brewer 1976). There are parallel changes in soil structure (subangular blocky to spherical micropeds) and horizon differentiation (A–C to A–B–C profiles). Soils with clayey B horizons today (U.S.D.A.: Alfisols, Ultisols, Spodosols and some Oxisols) are almost all naturally wooded or forested. The appearance of a weakly developed, clayey, subsurface horizon formed under late Devonian forests, provides an historical validation of this generalization. The Miocene appearance of spherical micropeds, thought to be produced by termites (Nye 1955; Stoops 1983) is well within the stratigraphic range of termite fossils (Burnham 1978). Thus many of the features characteristic of modern soils appear to have developed in tandem with the ecosystems they supported. Additional studies of palaeosols are needed to determine more precisely when various soil features first appeared and the degree to which they were a cause or a consequence of their associated ecosystems.

Further avenues of investigation have been stressed in this discussion, because I feel acutely the preliminary nature of this study. Early Palaeozoic palaeosols can be considered evidence of ancient landscapes and vegetation, and used to check or enlarge hypotheses based on interpretation of sediments or fossils. Since they record events during periods of non-sedimentation and are found also in oxidizing and acidic environments where fossils are seldom preserved, palaeosols also provide information unobtainable in other ways. Fortunately, early Palaeozoic palaeosols are not rare and there is no shortage of ways to study them.

This study would not have been initiated were it not for the foresight and generous assistance of R. D. Beerbower and J. D. Grierson (State University of New York, Binghampton), who personally showed me almost all the localities in eastern North America considered here. H. P. Banks (Cornell University, Ithaca) and J. Bridges (S.U.N.Y., Binghampton) also gave freely of their knowledge of these rocks. For useful discussion, I thank V. P. Wright (Open University, U.K.), W. D. I. Rolfe (Hunterian Museum, Glasgow), W. G. Chaloner (Bedford College, London) and A. J. Boucot (Oregon State University, Corvallis). Fieldwork in Pakistan was done with A. K. Behrensmeyer (Smithsonian Institution, Washington, D.C.) as part of a joint Yale University and Geological Survey of Pakistan Expedition in 1981, under the general direction of D. R. Pilbeam (Harvard University, Massachusetts). Research costs for the Palaeozoic palaeosols were borne by the Department of Geology, University of Oregon, in preparation of laboratory materials for a new course in Palaeopedology. Laboratory work on the samples from Pakistan was funded by N.S.F. Grant EAR8206183.

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Discussion

J. A. Catt (*Rothamsted Experimental Station, Harpenden, Hertfordshire AL5 AQ, U.K.*). Professor Allen and Professor Retallack have recognized and interpreted buried Palaeozoic soils. Soils should be fairly common in continental successions, but they can easily pass unnoticed because their friable upper horizons were often truncated before burial, and diagenetic changes have usually modified or destroyed many of their characteristic features. However, it is important that they are recognized and that their field relationships and mode of origin are studied in detail, because they have much stratigraphic significance in marking episodes of non-deposition and can provide important palaeontological evidence. Professor Retallack's observation of burrows in an Ordovician soil is especially significant in providing evidence for extremely early land animals, perhaps of a type hitherto unknown, and for a sufficiently dense cover of land plants to feed them. Some palaeoenvironmental evidence can also be obtained from buried soils, but many pedological features likely to be preserved in very old buried soils, such as carbonate concretions, ferruginous segregations, root channels, burrows and illuvial clay coatings, may form in many different climatic zones. Soil features that are more significant climatically, such as humus forms, water soluble salts, iron oxide and clay mineral assemblages, are unlikely to be preserved in unmodified form.

There is also the problem of distinguishing soils *in situ* from sediments containing transported soil materials or with diagenetic features (for example, carbonate concretions, iron segregations, subaqueous burrows) resembling those of pedological origin. The most convincing evidence for a buried soil depends upon compatible combinations of (i) typical vertical changes in composition or frequency of features (depth functions); (ii) characteristic lateral changes in relation to the topography of the associated land surface (catenary changes), and (iii) the co-existence of pedological features within individual profiles that either form together or persist through a sequence of changes. By these criteria, the evidence presented by Professor Retallack for his Ordovician, Silurian and Miocene profiles seems consistent with their interpretation as soils, but the Devonian example is less convincing. It is difficult to imagine clay illuviation occurring in a profile so devoid of soil structure that the original bedding of the fluvial sediment is preserved throughout. Oriented clay bodies may form in coarse fluvial sediments which are periodically flooded by muddy water, and the leaf and root remains might have been transported from vegetation upstream. This suggests that this profile should be interpreted as fluvial sediment unmodified by true soil-forming processes.

G. J. Retallack. There was not room in my paper to detail the various criteria for recognition of palaeosols, a topic I have dealt with elsewhere (Retallack 1983*a*, *b*). Dr Catt's comments on such criteria are thus a welcome addition.

Although the Peas Eddy clay palaeosol shows the most conspicuous relict bedding of those described, I interpret it as a palaeosol because of the presence of large, strata-transgressive root traces, examined both at the outcrop and in petrographic thin sections. Petrographic and chemical data presented are also evidence of limited translocation of iron and clay to a subsurface (B) horizon, confirming colour and textural differences seen in the field. It it had root traces alone it would be classified as a Fluvent (U.S.D.A. classification) palaeosol, but these additional features mark it as an Inceptisol.

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W. G. Chaloner, F.R.S. (*Department of Botany, Royal Holloway and Bedford Colleges, Huntersdale, Virginia Water, Surrey, U.K.*). The report of an Ordovician (Juniata Formation) soil, with its macro-fauna indicated by burrows, raises a question concerning the productivity potential of an early, purely algal, terrestrial flora. It is puzzling to me that terrestrial vegetation, presumably only of algal filaments, could have sufficient productivity to sustain a sizeable metazoan soil fauna. If the interpretation is valid, it appears that this terrestrial fauna may have evolved with little dependence on the evolution of rooted vascular plants.

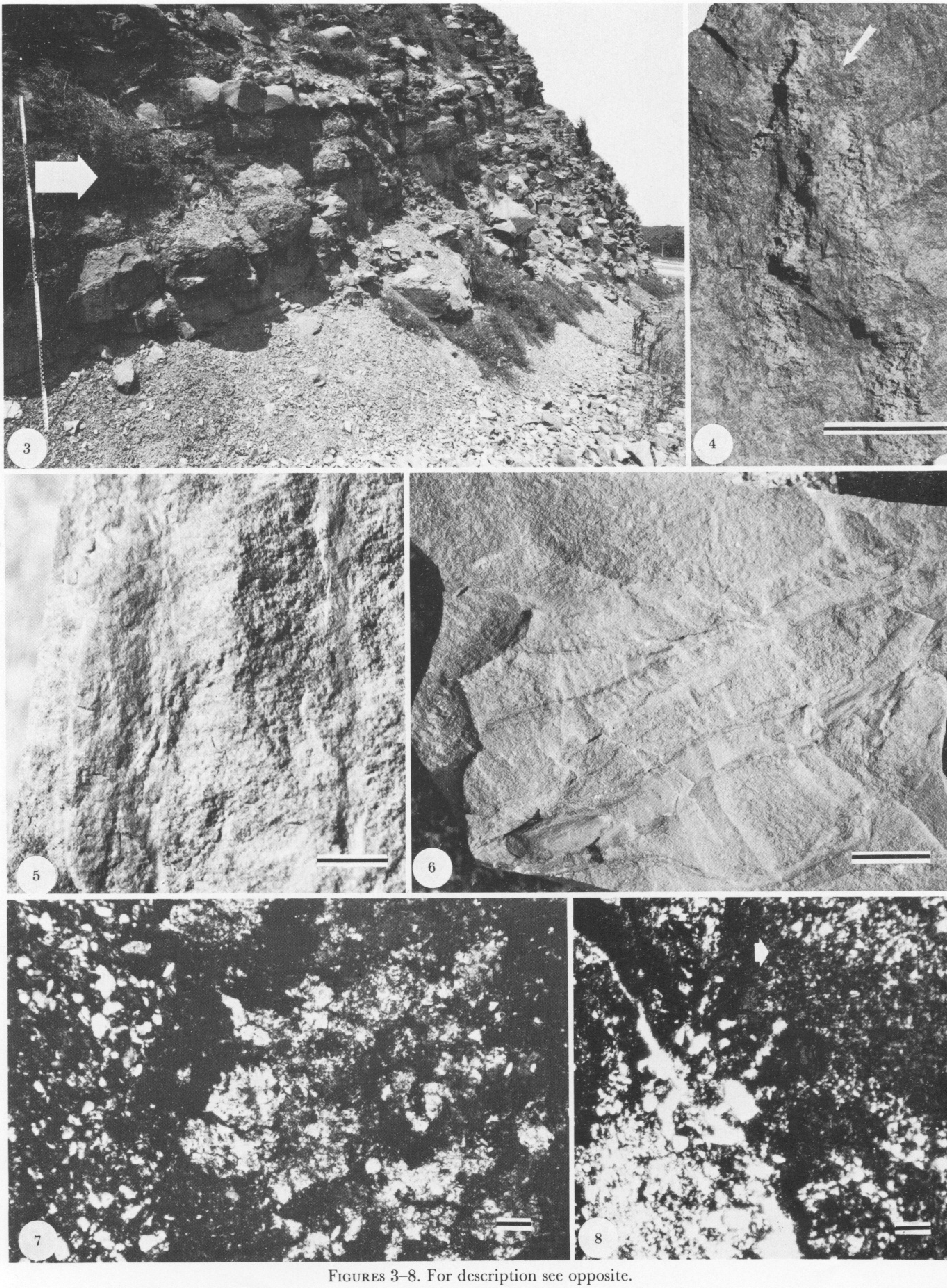
A further point related to the existence of pre-Silurian algal soils is that such a 'terrestrial' algal flora would present a population from which the archegoniates (bryophytes, vascular plants) may have been recruited. This has been suggested on other grounds by, for example, Stebbins & Hill (1980) as a more plausible route of vascular plant origin than the 'land migration' of an already elaborate, aquatic green macrophyte.

Reference

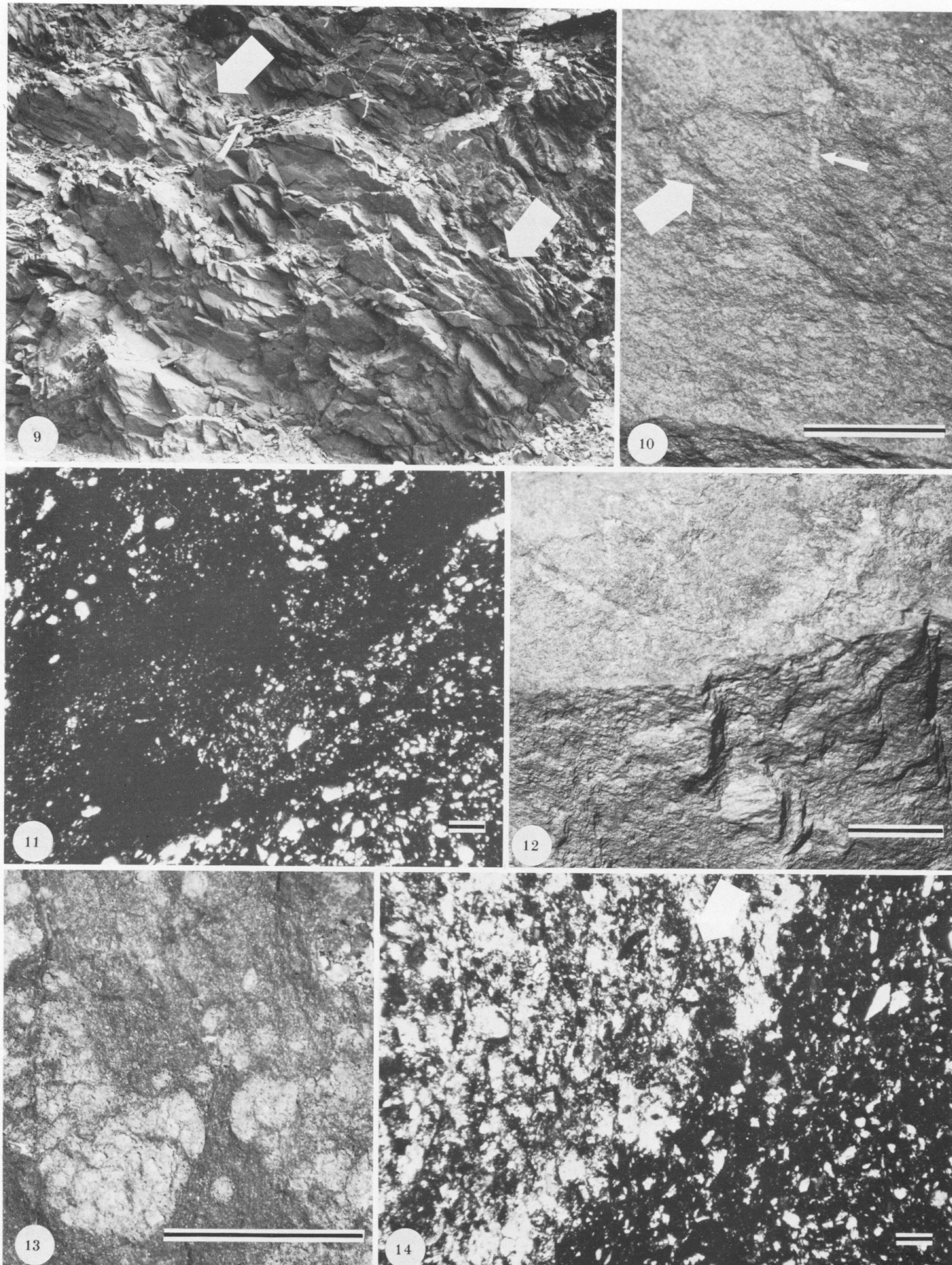
Stebbins, G. L. & Hill, G. J. C. 1980 Did multicellular plants invade the land? *Am. Nat.* **115**, 342–353.

G. J. Retallack. At present it is difficult to assess the nature of burrowing organisms and vegetation of the Late Ordovician palaeosol in the Juniata Formation. A whole range of plants, including lichens, liverworts, algae and totally extinct forms, ranging in habit from thallose to filamentous and unicellular, could have vegetated this palaeosol without leaving obvious physical traces. Although primary productivity of this vegetation may have been high, I have not yet been able to find any evidence that it was. The high density of burrows observed could have been created by a very small population of animals if abandoned burrows were not destroyed by bioturbation. I do agree with Professor Chaloner that these terrestrial metazoans evolved with little dependence on rooted vascular plants, since these appear much later in the geological record. The exact nature of both animals and plants in this palaeosol remain unknown.

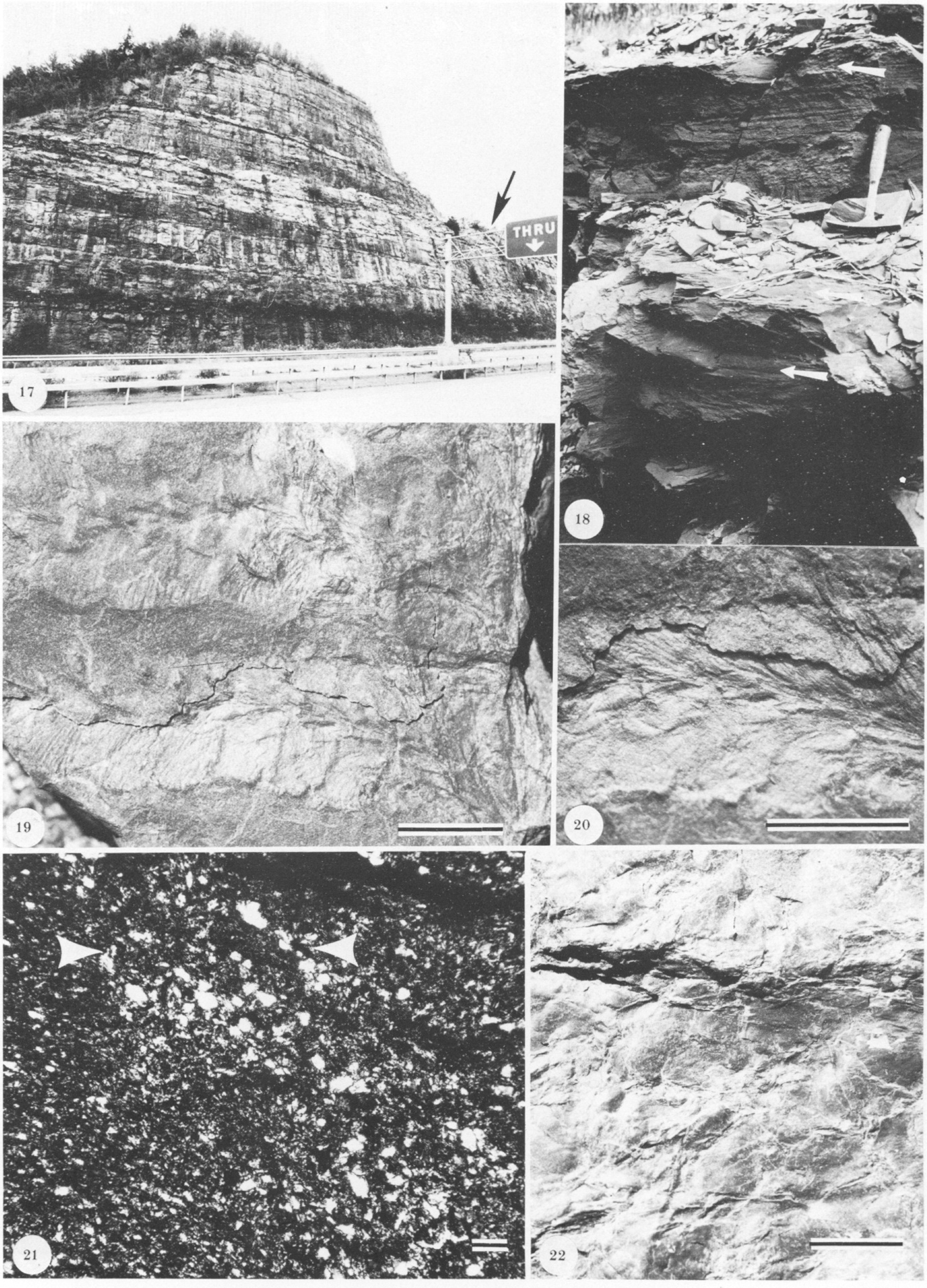
From consideration of this and other palaeosols now known, it does appear that terrestrial ecosystems were built up by degrees. This view is compatible with the idea of Stebbins & Hill (1980), that multicellular land plants evolved directly from unicellular land plants. It is also compatible with the alternative view that the land was prepared for the invasion of multicellular aquatic plants by pre-existing burrows and unicellular land plants. Intermediate scenarios of invasion by extremely small but still multicellular aquatic plants and animals also are possible. Only the idea of large plants pioneering sterile, bare earth now seems unlikely.



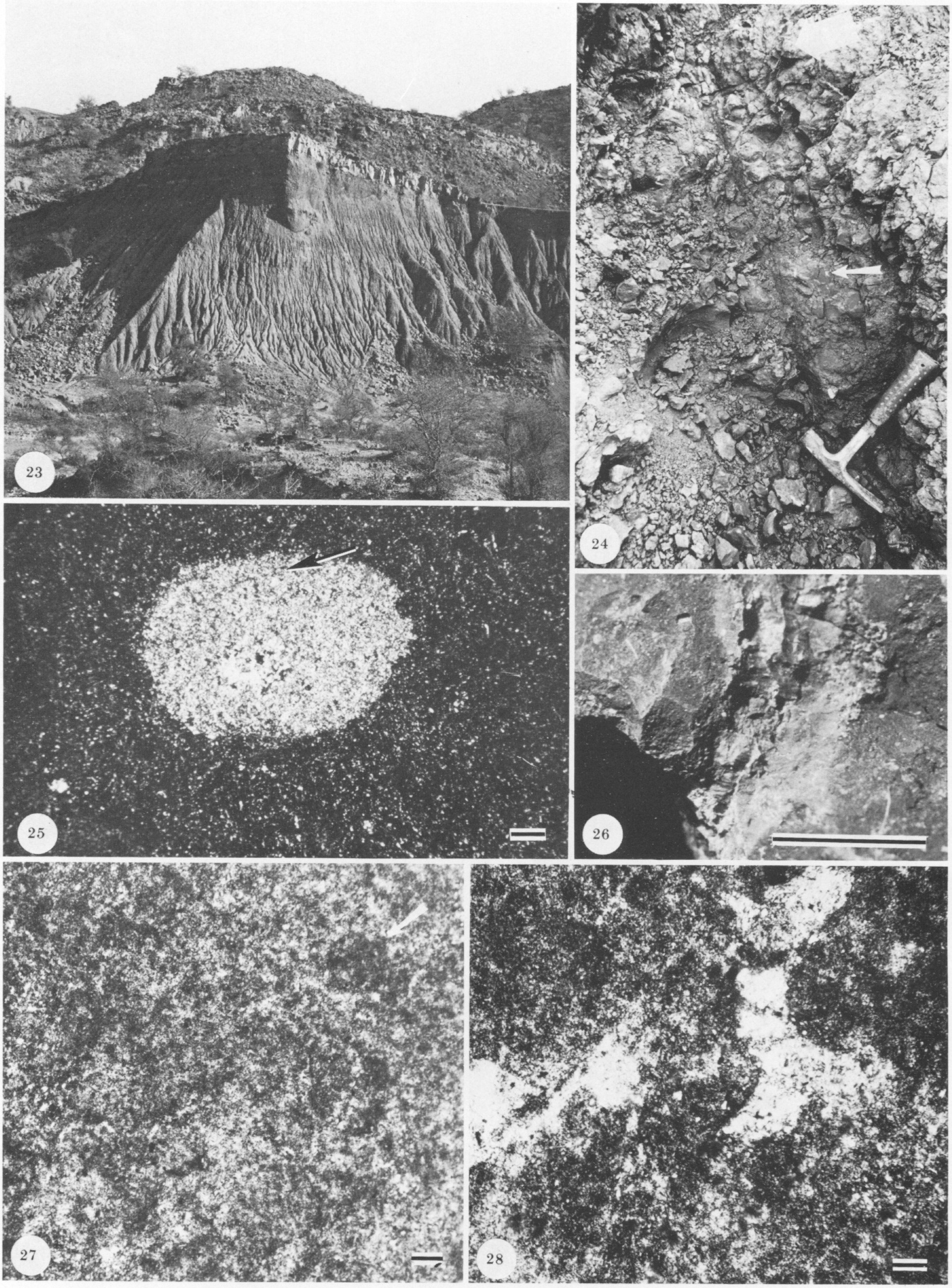
FIGURES 3–8. For description see opposite.



FIGURES 9–14. For description see p. 118.



FIGURES 17–22. For description see p. 119.



FIGURES 23–28. For description see opposite.