3 The Environmental Factor Approach to the Interpretation of Paleosols

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ABSTRACT

Paleosols ranging in geological age to 3 billion years are widely used as evidence for ancient surface environments—an enterprise dependent on the enormous literature on factors in soil formation, popularized by Hans Jenny. This kind of inference inverts the logic of Jenny in a manner common to geological sciences—deducing paleoenvironments from observed paleosol features, rather than deducing variation in soil features with observed environmental differences. A paleosol is a single product of many past influences, including alteration after burial, and because of this, some environmental relationships with soil color, clayeyness and organic matter are not useful for interpreting paleosols. One relationship that has proven useful for paleosols is that between depth to calcic horizon and mean annual rainfall. A new compilation of data presented here demonstrates that this relationship holds for aridland soils worldwide. The use of this relationship for interpreting paleoclimate from paleosols is illustrated with an example of the Eocene and Oligocene paleosols of Badlands National Park, South Dakota. Other approaches for the study of paleosols include identifying paleosols within a soil taxonomy, and simulating ancient soil development with mathematical process models. Identification of paleosols leads to broad areas on soil maps unless done with several paleosols. Process models often founder on assumptions, and those of the form $\delta x/\delta t$ (where x is a measured soil property) are difficult to apply because time of formation (t) of a paleosol is estimable only to an order of magnitude. Thus the environmental factor approach to the interpretation of paleosols is likely to remain popular for some time to come.

Hans Jenny's book *Factors in soil formation* (1941) was seminal in systematizing and quantifying empirical studies of soil formation. One can argue that his approach was not fundamentally distinct from that implied by Dokuchaev (1883) and other pioneers of soil science, or that the full promise of Jenny's universal equation has yet to be realized (Yaalon, 1975). Nevertheless, Jenny's vision of quantitative study of soil formation by means of soils chosen

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according to an explicit experimental design continues to have an undeniable influence in soil science, as well as in geological studies of soils and paleosols (Birkeland, 1984). Geological sciences include theoretical and taxonomic studies like those of soil science, but Jenny's approach is most similar to other kinds of study, such as the functional morphology of mammals. In such studies are documented the influence of specific factors such as diet on tooth and limb structure of mammals (Dodd & Stanton, 1990). Data to document such relationships needs to be chosen with care to mitigate other possible causes of variation. In Jenny's terms such a carefully chosen data set is analogous to a biosequence, and its study would aim at establishing biofunctions.

This however is only part of the task for geological sciences, which address not only how the world works at present, but what such understanding can reveal about events in the past. For example, with the knowledge that modern grassland mammals have high crowned teeth and elongate slender limbs with few digits, grasslands of the past can be inferred from finding such teeth and limb bones. Uniformitarianism is the term applied in geological sciences for such inferences about the past based on observations of the present. There is nothing special to geology about such logic. Indeed, it is little more than a kind of natural controlled experiment (Shea, 1982), or making inferences about the unknown (in this case events of the past) based on the known (observations of the present). It is this kind of logic that makes the environmental factor approach to the study of soils so useful for the interpretation of paleosols buried within sedimentary rock sequences ranging back in geological age up to 3 billion years (Retallack, 1990). Ongoing research on the relationships between soil features and the natural environment provides a large fund of potential inferences about the past from features observed in paleosols (Birkeland, 1984). This uniformitarian use of these data does not in itself attempt to extend or improve factor functions, although their limitations must be considered, but rather to use them. Such uniformitarian interpretation of paleosols also has limitations. Many details of soil formation cannot be observed in paleosols altered by burial over the millions of years since they formed (Retallack, 1991a).

The establishment of relationships between environment and features of surface soils and the use of such relationships to interpret paleosols are two distinctly different kinds of study in aims, materials, methods, scale and resolution. In this essay they will be illustrated by a specific example of the relationship between mean annual rainfall and depth to calcareous nodules in soils and paleosols. This is just one of numerous soil–environment relationships useful for interpreting paleosols, and only a tabular summary of other relationships can be offered here. None of these are without problems for application to paleosols, but then neither are taxonomic and modeling approaches, which also will be discussed as alternatives to the environmental factor approach.

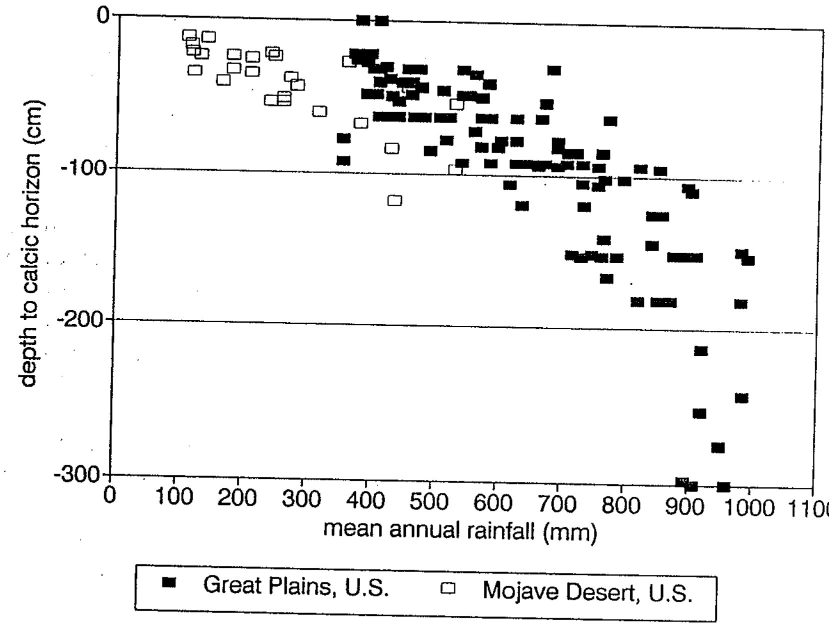

Fig. 3-1. The relationship between depth to calcic horizon in soils and their mean annual precipitation for the North American Great Plains (Jenny & Leoanrd, 1935; Jenny, 1941) and Mojave Desert (Arkley, 1963).

DEPTH TO CALCIC HORIZON AND MEAN ANNUAL RAINFALL

Previous Work

One of the most impressive quantifications of factors in soil formation in Jenny's (1941) book and an earlier paper (Jenny & Leonard, 1935) was a scatter plot (Fig. 3-1) of the depth to the zone of calcareous nodules in soils vs. mean annual rainfall, with the linear relationship $D = 86.74 - 0.2835P$, where D is the depth (cm) down from the surface to the top of the calcic horizon of a soil, and P is the mean annual precipitation (mm) at that site. I have replotted these data in metric units from Jenny's plot of 106 soils, giving a correlation coefficient (r) of 0.81 and a standard error (s) about the regression of ± 22 cm, using computer routines of Davis (1973). This relationship is an excellent example of a climofunction in the sense of Jenny (1941) because not only other aspects of climate, but also vegetation, topographic relief, parent material and time were more or less constrained for the set of soils considered. The relevant soils were on an east–west transect across postglacial (less than 15 000 yr old) loess of rolling grasslands of the central North American Great Plains.

Several subsequent studies have refined this work. Ruhe (1984) examined another transect of soils in this same area, and faulted Jenny's relationship because the time for formation and vegetation history at either ends of the transect were not exactly comparable. The eastern Udolls in the transect formed on Peoria Loess some 14 000 yr old, whereas the western Ustolls formed on Bignell Loess no more than 9 000 yr old. The eastern soils also were forested for a part of their postglacial history, unlike the western soils. Arkley (1963) also has been critical of this relationship because in the Mojave Desert he found a different relationship of $D = -2.734 - 0.1508P$. I also have replotted this data set of 26 soils, and calculated a correlation coefficient (r) of 0.75 and standard error (s) of ± 17 cm. The difference between the relationship of Jenny and Arkley is not so profound as apparent from casual inspection of the equations. Mojave and Great Plains data overlap substantially (Fig. 3–1), and may be better fit by a curve as shown schematically for Israeli soils by Dan and Yaalon (1982) and Yaalon (1983). The difference may be due to winter rainfall in the southwestern USA rather than summer rain of the Midwest, and to differences in water movement through the soil related to differences in porosity and structure of the soils in the two areas (Birkeland, 1984). Such differences also explain the differing values of precipitation across the udic–ustic soil moisture boundary in different countries (Yaalon, 1983).

Another group of soils from the southwestern USA was used by Marion et al. (1985) to calibrate a computer model of the depth of carbonate in aridland soils. They concluded that carbonate depths reflected a slightly more humid climate of the past rather than present mean annual rainfall. Subsequent computer modeling studies have shown that the depth of carbonate in soils is controlled by a complex mix of soil porosity, rainfall infiltration rates, soil respiration, available Ca, dust influx, temperature and other factors (McFadden & Tinsley, 1985; Mayer et al., 1988; McFadden et al., 1991). Because biological productivity also is related to mean annual rainfall and to soil respiration rates (Leith, 1975), the depth to the calcic horizon also can be used as a proxy indicator of rangeland productivity (Munn et al., 1978).

A New Compilation

From the perspective of high-resolution Quaternary studies these climofunctions need to be used cautiously. From the perspective of low-resolution geological interpretation of ancient paleosols; however, it is impressive how such a relationship could emerge at all from such a complex natural phenomenon. Indeed there have been indications in the literature that this relationship between depth to carbonate nodules and rainfall holds in such different soil-forming environments as the Serengeti Plains of Tanzania (de Wit, 1978), the Indo-Gangetic Plains of India (Sehgal et al., 1968) and the deserts of Sinai and Negev (Dan & Yaalon, 1982). These hopeful signs stimulated me to compile as many data as could be found in the literature on depth to the calcic horizon of soils whose mean annual precipitation also is known. The data set now stands at 317 soils, and includes soils from the Great Plains of North America (Jenny & Leonard, 1935; Jenny, 1941; Ruhe, 1984); the desert Southwest of the USA (Arkley, 1963; Marion et al., 1985); the pampas of Argentina (Fadda, 1968; U.N.FAO, 1971; Soil Correlation Committee for South America, 1967; Plaza & Moscatelli, 1989); the central Spanish meseta (del Villar, 1957); the Serengeti Plains of Tanzania (de Wit, 1978; Jager, 1982); the Kalahari Desert of Botswana (Siderius, 1973); the Sinai–Negev deserts of Israel (Dan et al., 1981; Dan & Yaalon, 1982); the Mesopotamian Plains of Syria, Iraq, and Iran (Mulders, 1969; Al Taie et al., 1969; U.N.FAO, 1977; Hussain et al., 1984); the Indo-Gangetic Plains of India (Sehgal et al., 1968; U.N.FAO, 1977; Sidhu et al., 1977; Ahmad et al., 1977; Bhargava et al., 1981, Vinayak et al., 1981, Murthy et al., 1982; Courty & Féderoff, 1985); the south Russian Plain (Dokuchaev, 1883; Glinka, 1931; U.N.FAO, 1981); the southwestern Siberian Plain (Fedorin, 1960; Bal & Buursink, 1976); the western Mongolian Plain (Nogina, 1976); the north China Plain (Thorp, 1936; Bronger & Heinkele, 1989); the Riverina district of central southeastern Australia (Stace et al., 1968); intermontane basins of the South Island of New Zealand (McCraw, 1964; Raeside & Cutler, 1966; Leamy & Sanders, 1967; Soil Bureau Staff, 1968; Orbell, 1974); and the polar deserts of Antarctica (Campbell & Claridge, 1987) and Greenland (Tedrow, 1970, 1977). Where climatic data were not provided within these soil studies they were obtained from published tabulations and isohyet maps (Alt, 1932; Watts, 1969; Hoffman, 1975; Lydolph, 1977; Taha et al., 1981; Ruffner, 1985).

A few ground rules were used for selection of soils for this database, which is presented here in full (Appendix 3–1) in the hope that others will expand upon it. All are late Pleistocene and Holocene soils on unconsolidated sediments other than clay or limestone in low lying or rolling terrain of free drainage, more or less as in Jenny and Leonard's (1935) original data. The calcic horizon was taken as the horizon of most abundant micritic carbonate nodules that appear to be in place. This was not always the highest nodule or carbonate in a profile, especially within monsoonal soils which may have at least a few calcareous nodules throughout the profile (Retallack, 1991b). Although soil horizons above the calcic horizon may be weakly calcareous because leached of carbonate, this measure of the depth to the horizon of carbonate acccumulation is not the same as the depth of leaching of a calcareous parent material, which apparently reflects time for formation rather than climatic conditions (Birkeland, 1984). Carbonate "veins" or "cement" were not accepted as a calcic horizon, because these could be of groundwater origin, rather than pedogenic origin (Mann & Horwitz, 1979; Carlisle, 1983; Lander, 1990; Kaemmerer & Revel 1991; Wright & Tucker, 1991). Solid carbonate layers ("K horizons" of the USA or "tosca" of Argentina) also were not included because they indicate soils of great antiquity and more complex history (Gile et al., 1966; Pazos, 190), nor were soils with stone lines or other indications of redeposition in alluvial parent materials (Courty & Féderoff, 1985). Also not included were soils of hills and steep slopes, nor soils on bedrock, limestone, beach rock, obvious local sand dunes or clay. In most cases, these various complications were apparent from micromorphological, geochemical or other studies of the soils, but there are bound to be suspect soils that evaded detection. A literature compilation such as this is no substitute for a careful selection and study of all the soils by a single investigator, as can be seen from the more highly correlated results of Jenny and Leonard (1935) and Arkley (1963).

Despite these strictures, there remains considerable variation in environmental factors within the data of my compilation. Climate, for example, ranges from frigid to tropical, and from mildly seasonal to monsoonal. Most of the soils support grassland, but many are under dry woodland and desert shrubland. Their topographic setting ranges from extensive alluvial plains to loess-mantled rolling terrain. Parent materials include volcanic ash, quartzofeldspathic silty alluvium and boulder till. Time for formation varies from late Pleistocene (perhaps 50 000 yr old) to late Holocene (perhaps as young as 500 yr).

Even with those unwieldy data the points define a clear relationship between depth to calcic horizon and mean annual rainfall (Fig. 3–2). The relationship is not linear, but best-fit by the curve $D = -40.49 - 0.0852P - 0.0002455P^2$, which has a correlation coefficient (r) of 0.78 and a standard error (s) about the curve of ± 33 cm. Application of an F test to an analysis of variance (following methods of Davis, 1973) shows that this fit is highly significant at the 1% level, and is at least an equally significant advance over the fit provided by linear regression. The fitting of higher-order polynomial curves however, gave insignificantly improved fit. This is a surprisingly clear relationship for a data set of such diverse origins and it encourages faith that something fundamental about the way soils form has emerged above the noise of local variations in soil history and other factors. Better results have been obtained locally by Jenny and Arkley with more carefully and deliberately constrained field studies. Such studies remain a promising direction for further research, and can be constructed from the raw data of Appendix 3–1.

Problems in Application to Paleosols

From the perspective of interpreting paleosols, not the relationship of depth of carbonate to rainfall, but rather the relationship of rainfall to depth of carbonate is needed. Regression of this relationship for the new compliation yields the equation $P = 139.6 - 6.388D - 0.01303D^2$, with a correlation coefficient (r) of 0.79 and a standard error (s) of ± 141 mm. The broad error envelope is in part a consequence of compression of the data toward the soil surface (Fig. 3–2). This level of resolution may not be helpful for studies of late Quaternary paleosols, but can be valuable for paleoclimatic interpretation of paleosols in ancient sedimentary rocks. Furthermore, there is the prospect of interpreting paleoclimate from subsets of these data—climosequences constrained more narrowly to approximate other paleoenvironmental conditions of particular paleosols. For example, in the Great Plains data of Jenny and Leonard (1935) and Jenny (1941) the relationship is $P = 418.5 - 2.335D$, with a correlation coefficient of 0.81 and a standard error of ± 108 mm.

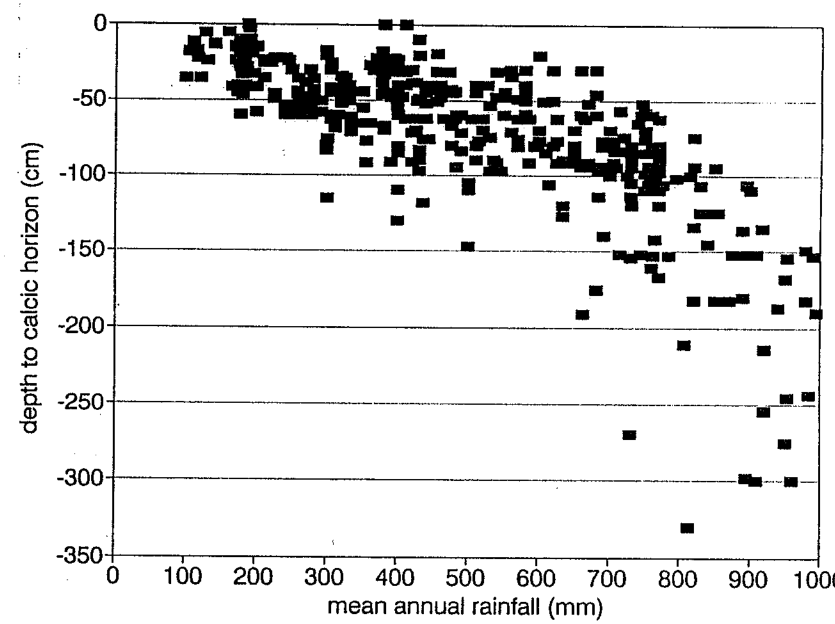

Fig. 3–2. The relationship between depth to calcic horizon in soils and their mean annual precipitation from a new compilation of 317 aridland soils from all over the world (see text for sources).

There are troublesome, but not unsuperable, additional problems in applying these data to the interpretation of paleoclimate from paleosols, because of different atmospheric concentrations of CO_2 during the geological past, the erosion of soils before burial and their compaction after burial. The first of these is difficult to circumvent because there are no soils presently forming under the kind of global greenhouse climate of elevated atmospheric CO_2 envisaged, for example, during mid-Cretaceous geological time. Indeed, Ice Ages like that which have persisted for roughly the past 2 million years are comparatively rare events in earth history. In the distant geological past these Ice Ages with their waxing and waning of large continental ice sheets have persisted for durations only a few tens of millions of years, some 290 million years ago (Carboniferous–Permian boundary), again some 440 million years ago (Ordovician–Silurian boundary), 600 , 1000, and 2300 million years ago (all Precambrian; Hambrey & Harland, 1981). Analysis of gas bubbles in ice cores indicates that CO_2, which is currently at levels of 300 mg L^{-1}, may have been as high as 400 mg L^{-1} during a warm climatic period a few thousands of years ago, and as low as 200 mg L^{-1} during the glacial maximum, 15 000 yr ago (Neftel et al., 1982). Isotopic study of paleosol carbonates combined with a theoretical model for the C isotopic systematics of soils have been used as evidence that atmospheric CO_2 was no more than 700 mg L^{-1} during the Eocene, but perhaps on the order of 1500 to 3000 mg L^{-1} during Early Cretaceous time (Cerling, 1991). This order of magnitude variation in atmospheric CO_2 abundance is less than the two orders of magnitude difference between CO_2 abundance in the atmosphere and in some modern highly productive soils (Brook et al., 1983), but during such greenhouse times in earth history rainwater would have been somewhat more acidic. Thus it is possible that the level within soils at which nodules formed may have been deeper for any given mean annual precipitation. How much deeper for a given level of CO_2 is difficult to calculate, because the calculations are dependent on estimates of soil respiration. Estimates can be made using the numerical models of McFadden and Tinsley (1985) and Mayer et al. (1988).

The problems of erosion of paleosols before burial is less troublesome. Many soils eroded due to misguided agricultural practices of the past are included in the new compilation (Fig. 3–2). In addition, surface horizon structures in many aridland paleosols are distinctive in the field from those in their subsoil horizons. The dark color and organic matter of mollic epipedons are seldom preserved in buried soils that were well drained, but there are commonly granular and platy peds defined by argillans and a variety of root traces and rhizoconcretions (Retallack, 1990, 1991a,b). These can be examined in the field, along with the nature of the contact between the paleosol and its overlying sediments, to determine the likelihood that depth to carbonate nodules has been reduced by erosion of the soil. There is not much that can be done to reconstruct severely eroded paleosols, and they are best ignored for the purpose of paleoclimatic reconstruction. In general, surface horizons of paleosols are widely and well preserved in paleosols within thick sedimentary sequences that accumulated in subsiding sedimentary basins, but compound paleosols with eroded topsoils are widespread in tectonically stable or uplifted continental regions (Retallack, 1986a; Schaetzl & Sorenson, 1987).

A final difficulty is compaction of paleosols due to loading of overburden after burial and to other kinds of tectonic deformation. The degree of compaction of paleosols as geologically young as Miocene can be significant if, for example, they were buried by large volcanic edifices or in such thick sedimentary sequences as Himalayan outwash (Retallack, 1991b). The best direct evidence for degree of compaction is what in geological sciences is called a clastic dike, and in soil sciences is called a silan; that is to say, a former near-vertical crack in the paleosol that has become filled with a contrasting material that buried the paleosol. Such structures become tightly folded with compaction of the surrounding paleosol matrix, and have been useful in reconstructing the original thickness of paleosols as geologically ancient as Precambrian (Retallack, 1986b). Root traces and burrows in paleosols also become folded and flattened with compaction of paleosols, but in these cases original irregularity and flattening diminishes their usefulness as indicators of compaction during burial. Another approach is to determine how much higher is the bulk density of paleosol samples than of samples of comparable surface soils (Retallack, 1991b). Density of paleosols and soils is quite variable and difficult to measure with sufficient accuracy for this to be more than a check on the results of other approaches. The compaction correction method that I prefer is to reconstruct depth of burial from geological data and then estimate likely compaction from standard curves for the compaction of appropriate kinds of rocks with depth (Baldwin & Butler, 1985). Reconstructing the depth of burial may be possible by considering regional stratigraphy and structure, but there also are a variety of other indices of burial including vitrinite reflectance, hydrocarbon maturation, and pollen color alteration (Tissot & Welte, 1984).

Example of Eocene–Oligocene Paleosols and Paleoclimate

A specific case study of the paleoclimatic interpretation of Eocene and Oligocene paleosols in Badlands National Park, South Dakota, may clarify what the factor–function approach can tell us about the past. The Badlands are well known as a long ragged wall of colorful clayey nonmarine rocks between high prairie to the north and the valley of the White River to the south (Fig. 3–3). They have long been famous as the foremost example of badlands weathering (Schumm, 1975) and for a great variety of well-preserved fossil mammals (Emry et al., 1987) now known to range in age from late Eocene to early Oligocene (somewhat older than previously thought; Swisher & Prothero, 1990). My own studies of the numerous paleosols within this sedimentry succession have focused on a single measured section in the Pinnacles area of the National Park (Fig. 3–4), which has been used as evidence of grassland ecosystems (Retallack, 1982, 1984b, 1988a, 1990), the completeness of stratigraphic sections (Retallack, 1984a), factors in sedimentation (Retallack, 1986a), and paleoclimatic change (Retallack, 1992a).

Many of these paleosols are studded with calcareous nodules that have all the earmarks of being original soil nodules. They form definite horizons below the surface horizons of the paleosols and some show mild surface ferruginization. In petrographic thin section, micritic matrix can be seen to locally replace pre-existing grains and local sparry calcite displaces soil peds (soil clods) and fills hollows after root traces (Retallack, 1983a). Similar nodules are commonly found as clasts in channel deposits at the same stratigraphic level (Wanless, 1923). Fossil skulls, tortoises (*Stylemys nebrascensis*), and other hollow groups of bones are noticeably less crushed, distorted and oxidized within nodules than in adjacent clayey paléosol matrix, and so predate lithostatic compaction. Nodules also contain more weatherable minerals, such as volcanic glass, hypersthene and olivine, than the surrounding claystone. Finally, some nodules are penetrated by drab-haloed root traces, thought to have formed during the early burial decomposition of the last crop of roots before burial of the paleosols (Retallack, 1983a, 1991a).

Although there are some clearly pedogenic nodules in Oligocene paleosols of these sequence, there also are carbonate layers and a widely dispersed carbonate cement. This latter cement is most conspicuous within paleochannel sandstones and in C horizons of paleosols in the silty upper part of the sequence, and is evidence of a component of carbonate added to the sequence presumably from groundwater during burial. The carbonate layers found mainly within the late Eocene clayey paleosols are more problematic. They formed within the soil zone, because they include replacive micritic matrix,

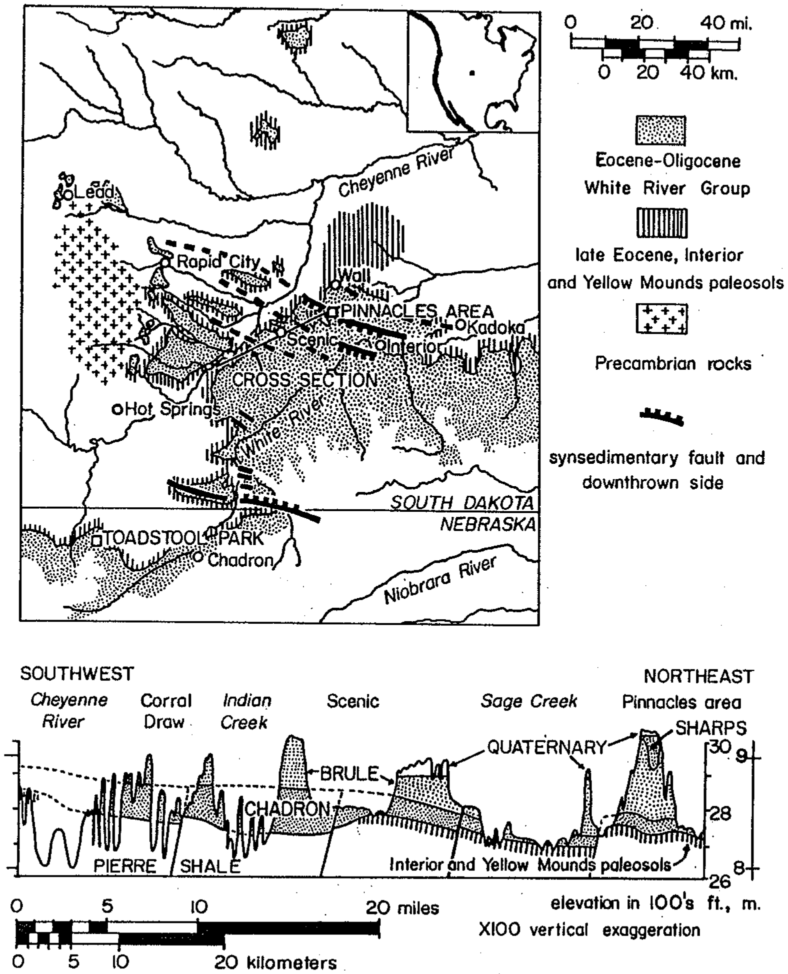

Fig. 3-3. Location (above) and geological cross section (below) of a measured section of Eocene and Oligocene paleosols in the Pinnacles area of Badlands National Park, South Dakota (from Retallack, 1988a).

displacive brecciated textures, cellularly preserved root traces, calcite crystal tubes after roots, fossil larval cells of dung beetles (*Pallichnus dakotensis*) and bees (*Celliforma ficoides*), and displacive sparry calcite veins. They also are disrupted by fossil root traces (Retallack, 1983a, 1984b). However, their morphology is unusual for soils. There may be up to five of these calcareous stringers within a single profile, each stringer with sharp upper and lower contact. Although roughly stratabound, the stringers curve up and down across bedding and thicken and thin irregularly. I have long been uncom-

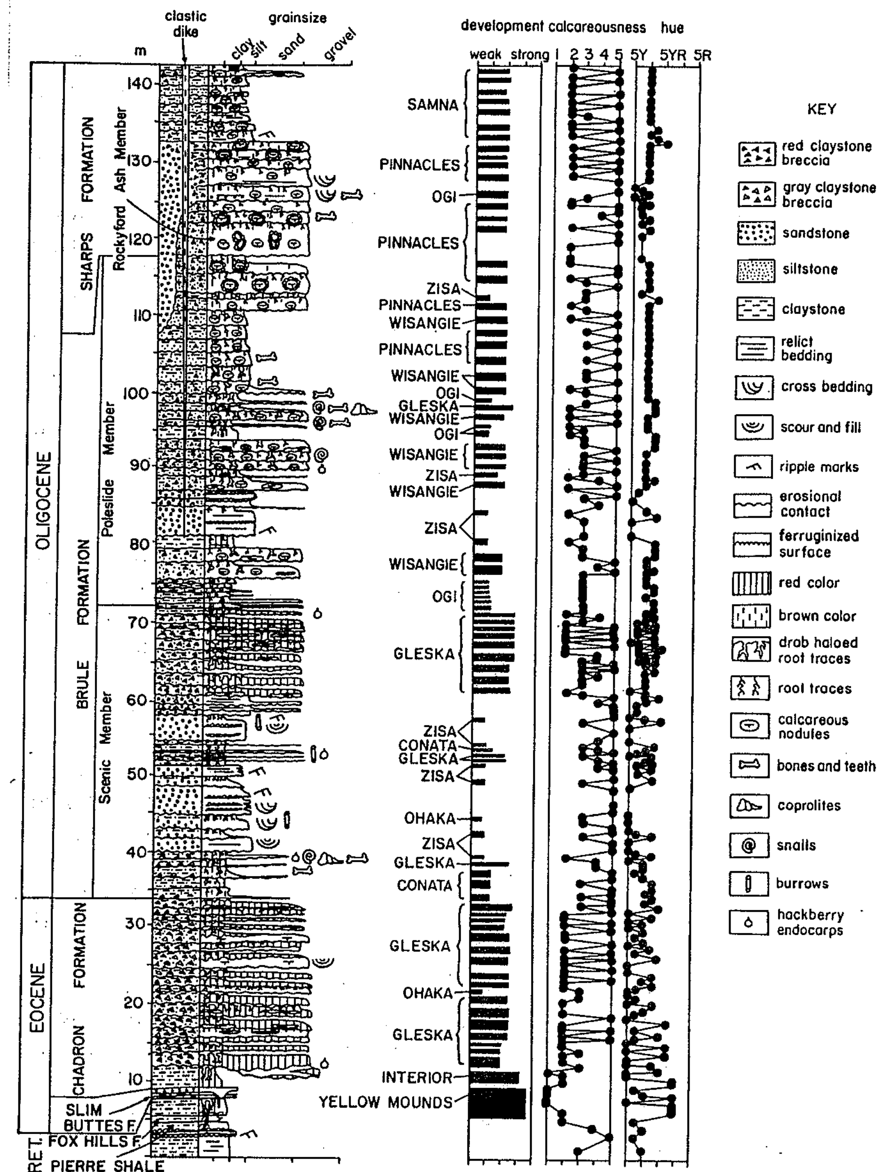

Fig. 3-4. A measured section of Eocene and Oligocene paleosols from the Pinnacles area of Badlands National Park, South Dakota (from Retallack, 1990). Individual paleosols are marked by heavy bars, the width of which corresponds to relative degree of soil development. Calcareousness is from a nominal scale of Retallack (1988b) and hue from a Munsell color chart compared with freshly excavated rock.

fortable interpreting these as strongly developed calcic (K) horizons (Retallack, 1983a), and new isotopic results obtained by Lander (1990) have convinced me that these are near-surface examples of valley calcretes, similar to those now found in the deserts of Western Australia (Mann & Horwitz, 1979) and Namibia (Carlisle, 1983). Also, like these valley calcretes the Badlands petrocalcic stringers are most prolific near paleochannel sandstones and include patches of opal, chalcedony and carnotite. There is little theoretical reason to suspect that carbonates derived in part from groundwater evaporation within the soil zone should show the same relationship to mean annual rainfall as carbonate nodules of well-drained aridland soils; however, the two modern cases are very near the surface of the soil in very dry regions. The paleoclimatic significance of valley calcretes remains to be established.

Paleoclimatic interpretation of calcareous paleosols of Badlands National Park is not severely compromised by changing atmospheric levels of CO_2. The Eocene–Oligocene boundary was a time of global climatic cooling, the most significant of several abrupt climatic deteriorations which marked the transition from a middle Eocene greenhouse climate to the Plio–Pleistocene Ice Age (Miller, 1992; Retallack, 1992a). Calcareous paleosols of Badlands National Park are principally of Oligocene age and probably formed under atmospheric CO_2 levels much less than 700 mg L^{-1} postulated for the Eocene by Cerling (1991) and closer to interglacial levels of about 300 to 400 mg L^{-1} experienced by many surface soils of the new compilation (Fig. 3–2).

Erosion of the surface of paleosols in the Badlands before their burial also was minimal. Few of the paleosols in this sequence of wind and water redeposited volcanic ashes (Evanoff et al., 1992) are sharply overlain by sediments, and those that are sharply truncated also have preserved the platy to granular structure and fine root traces of their surface horizons (Retallack, 1983a, 1990).

Compaction of the paleosols also was not a problem for interpretation of the depth of their carbonate nodules. The maximum thickness of the Eocene–Oligocene White River Group is 195 m, and the cumulative thicknesses of other Oligocene and geologically younger formations in this region total 361 m (Martin, 1983). If a marine, noncalcareous shale was buried to a depth of 500 m, it would be compacted to about 90% of its former thickness (Baldwin & Butler, 1985). This seems a very unlikely maximal compaction even at the base of the sequence, because the geologically younger sediments did not form a single layer, but filled local paleovalleys excavated as the nearby Black Hills and Great Plains continued to rise (Schultz & Stout, 1980; Swinehart et al., 1985; Angevine & Flanagan, 1987). This together with the increasingly pervasive calcareous cement and silty texture higher in the sequence makes it unlikely that compaction was significant.

The depths to calcareous nodular horizons within paleosols, recorded in the field during section measuring, show a decline in paleosols stratigraphically higher in the section that, from the relationship demonstrated for surface soils (Fig. 3–2), can be interpreted as evidence of Oligocene climatic drying (Fig. 3–5). The petrocalcic horizons, for which paleoclimatic significance has yet to be demonstrated convincingly, also continue the trend of wetter climate further back into the Eocene where the paleosols are largely noncalcareous. The basal paleosol of the sequence is leached of carbonate in its calcareous marine sedimentary parent material to a depth of 5 m. This deeply leached Eocene paleosol may have formed in a climate with annual rainfall well in excess of 1000 mm. Rainfall enjoyed by overlying late Eocene paleosols is uncertain, but was probably intermediate between this and early Oligocene paleosols for which mean annual rainfall was some 450 mm ± 141. Additional drying during mid-Oligocene time brought rainfall down to semiarid levels of about 350 mm ± 141. There is great fluctuation in the individual data points, much of which is not significant given the large standard error from the compilation of modern soils. However, the especially marked breaks at 72 and 118 m coincide with both local periods of gully erosion and nondeposition (Retallack, 1986a) and global episodes of climatic deterioration (Retallack, 1992). Within each depositional unit punctuated by erosional and climatic events the trend revealed by linear regression through the fluctuation is in each case a climatic drying (Retallack, 1986a, 1992).

Such evidence from paleosols can be important additions to other paleoclimatic indicators, such as eolian sediments, fossil leaves, and vertebrates. Drier climate of later Oligocene time also may explain the decreased

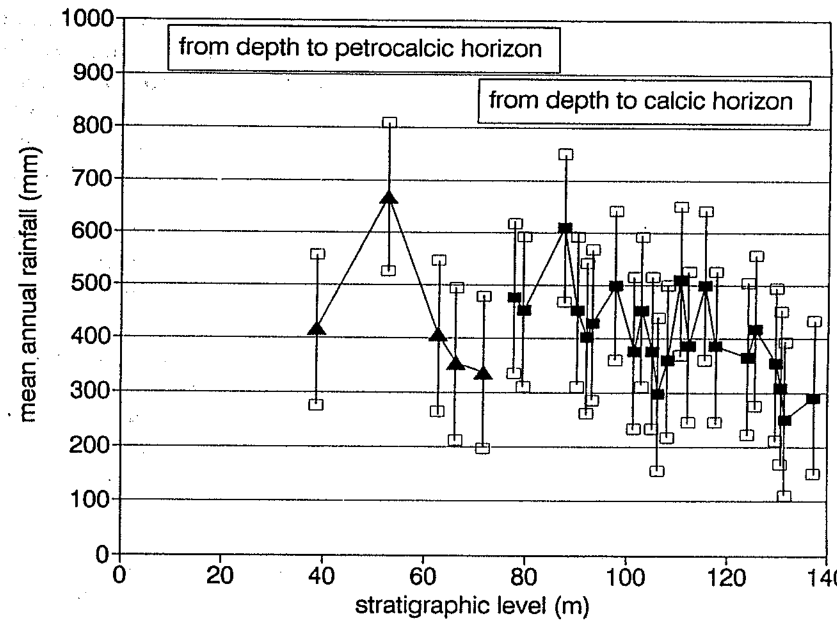

Fig. 3–5. Higher, and thus geologically younger, within the measured section of Eocene and Oligocene paleosols from South Dakota (Fig. 3–2), calcic horizons become shallower within the profile. Using the relationship compiled here (Fig. 3–1), this can be interpreted as a result of declining mean annual rainfall during Oligocene time. Closed triangles and boxes are calculated rainfall using regression and open boxes are error (± 1σ).

abundance of kaolinite and increased abundance of smectite and then illite in paleosols higher in the sequence, as well as changes in hue from red to brown and then yellow, and changes in texture from clayey to silty (Retallack, 1986a). Climatic drying and possibly also cooling is prominently displayed in the appearance of these colorfully banded badlands. Although few fossil plants are preserved in the Badlands of South Dakota, the paleosols contain abundant drab-haloed root traces, and their profile form and soil structures also are evidence of changing vegetation; moist forests of the Eocene [37 millions of years before the present (M.Y.B.P.)] giving way to dry forests by late Eocene (35 M.Y.B.P.), to dry woodland by Oligocene (33 M.Y.B.P.), to wooded grasslands with streamside gallery woodland by 32 M.Y.B.P. and large areas of open grassland by 30 M.Y.B.P. (Retallack, 1983a, 1986a). Some changes in assemblages of fossil vertebrates of Badlands National Park can be related to progressively drier conditions and more open vegetation. Climatic drying at the Eocene–Oligocene boundary correlates in time with local extinction of large titanotheres, alligators and a wide variety of turtles and amphibians (Hutchison, 1982; Emry et al., 1987). The long-term evolutionary trend toward the kinds of cursorial limbs and high-crowned teeth well displayed in modern mammalian faunas of grasslands was initiated during this time, but in these respects Eocene and Oligocene faunas remained more like modern faunas of woodland than those of grassland (Retallack, 1990). Mid-Tertiary paleoclimatic cooling and drying was an early impetus for evolution of the grassland biome in continental interiors (Retallack, 1992a).

OTHER FACTOR-FUNCTIONS USEFUL FOR PALEOPEDOLOGY

Numerous other relationships between soil features and environmental factors also are useful for intrepreting paleoenvironments from paleosols. Only a brief tabular summary (Tables 3–1, 3–2, 3–3, 3–4, 3–5) can be offered here, but a more extended textbook treatment of the application to paleosols of these other relationships is available elsewhere (Retallack, 1990). Documentation of these relationships in surface soils can be found in a vast literature, as well as in widely available textbooks (Buol et al., 1989; Birkeland, 1984). The references to studies of surface soils cited here (Tables 3–1 to 3–5) have been chosen with deliberate preference for studies using micromorphological and bulk geochemical approaches, because these approaches have proven especially useful in the study of paleosols (Retallack, 1983a, 1991b).

Several widely discussed relationships for surface soils have been omitted from this tabular summary for use with paleosols, because their application is severely compromised by alteration of paleosols after burial, or because the measured soil features are controlled by too many factors to be teased apart. Soils buried in sedimentary successions suffer many of the same alterations as have been well documented for their enclosing sediments (Scholle & Schluger, 1979; Tissot & Welte, 1984; Surdam & Crossey, 1987). Many

Table 3–1. Paleoclimatic indicators in paleosols.

Climatic variable	Paleosol feature	Effective range	Studies of surface soils	Application to paleosols
Mean annual temperature	Tower karst	More than 12°C	Jennings, 1985	Leary, 1981
	Cavernous subsoil weathering	More than 12°C	Jennings, 1985	Wright, 1981
	Black phytokarst	More than 12°C	Folk et al., 1973	Folk & McBride, 1976
	Spherical micropeds	More than 8°C	Stoops, 1983	Retallack, 1991b
	Pingos	Less than −1°C	Washburn, 1980	Williams, 1986
	Ice-wedge polygons	Less than −4°C	Washburn, 1980	Williams, 1986
	Sand-wedge polygons	Less than −12°C	Washburn, 1980	Williams, 1986
Mean annual precipitation	Presence of gibbsite	More than 800 mm	Sherman, 1952	Keller et al., 1954
	Proportion of smectite to kaolinitic clay	±250 mm over range 100–3000 mm	Barshad, 1966	Retallack, 1983a
	Presence of carbonate	Less than 1000 mm	Birkeland, 1984	Retallack, 1983a
	Depth to calcareous nodular (Bk) horizon	±141 mm over range 100–1000 mm	Jenny, 1941; Arkley, 1963; Retallack, 1994 (this paper)	Retallack, 1983a, 1991b
	Presence of palygorskite	Less than 400 mm	Singer & Galan, 1984	Watts, 1976
	Presence of gypsum	Less than 300 mm	Birkeland, 1984	West, 1975
	Depth to gypsum crystal (By) horizon	±100 mm over range 0–300 mm	Dan & Yaalon, 1982	West, 1975
Seasonality of rainfall	Intergrown ferric concretions and calcareous nodules	Monsoonal wet–dry seasonality	Sehgal et al., 1968; Courty & Féderoff, 1985	Retallack, 1991b
	Dikes and festooned shear planes (mukkara)	Pronounced dry season, semiarid to subhumid	Paton, 1974	Allen, 1986a; Retallack, 1986b
	Common charcoal	Pronounced dry season, semiarid to humid	de Castri et al., 1981	Harris, 1957; Retallack & Dilcher, 1981
	Surface root mat and very deep sinker roots	Pronounced dry season, arid to subhumid	van Donselaar-ten Bokkel Huinink, 1966; Rutherford, 1982	Retallack, 1983a, 1991b

Table 3-2. Paleosol features of different vegetation types.

Plant formation	Root traces	Soil horizon sequence	Other soil features	Soil type	Studies of surface soils	Application to paleosols
Rain forest	Large and small in tabular mat	A-(E)-Bt-C	Spherical micropeds, few weatherable minerals, bases/alumina ratio near zero	Oxisol, Ultisol	Sanchez & Buol, 1974	Retallack, 1991c
Oligotrophic forest	Large and small, deeply penetrating	A-(E)-Bs-C	Quartz-rich, few weatherable minerals	Spodosol, Dystrochrept	Mokma & Vance, 1989	Retallack, 1990
Forest and woodland	Large and small, deeply penetrating	A-(E)-Bt-C	Blocky peds, clay skins, some weatherable minerals	Alfisol, Ultisol	Ciolkosz et al., 1990	Retallack, 1983a, 1991b
Dry woodland	Large and small deeply penetrating	A-Bt-Bk-C	Blocky peds, clay skins, common weatherable minerals and carbonate	Alfisol, Mollisol, Aridisol	Murthy et al., 1982	Retallack, 1983a, 1991b
Wooded grassland	Abundant fine (<2 mm), few large, deeply penetrating	A-(Bt)-Bk-C	Granular peds at surface, shallow calcic horizon	Mollisol, Inceptisol	de Wit, 1978	Retallack, 1991b
Open grassland	Abundant fine (<2 mm) near surface	A-(Bt)-Bk-(By)-C	Granular peds at surface, shallow calcic horizon	Mollisol, Inceptisol	Aandahl, 1982; Bronger & Heinkele, 1989	Retallack, 1983a, 1991b
Desert scrub	Sparse large, deeply penetrating	A-(Bt)-Bk-(By)-C	Vesicular, platy or blocky structure at surface, very shallow calcic horizon	Aridisol, Inceptisol, Entisol	Dan & Yaalon, 1982	Loope, 1988
Desert shrubland	Sparse, medium, woody, deeply penetrating	A-(Bt)-Bk-(By)-C	Vesicular, platy or blocky surface, very shallow calcic horizon, stone pavement	Aridisol, Inceptisol, Entisol	Stace et al., 1968	Loope, 1988

(continued on next page)

Plant formation	Root traces	Soil horizon sequence	Other soil features	Soil type	Studies of surface soils	Application to paleosols
Microbial earth	None	A-C	Ministromatolites, laminar crusts, claystone breccias	Entisol	Friedmann et al., 1967	Retallack, 1986b, 1990
Microbial rockland	None	A-C	Rock varnish, endolithic microrelief	Entisol	Folk et al., 1973; Friedmann & Weed, 1987; Viles, 1987	Beeunas & Knauth, 1985
Brakeland	Sparse rhizomes and fine roots	A-(Bk)-C	Platy and blocky peds at surface	Entisol, Inceptisol	Walker & Peters, 1977	Retallack, 1990
Polsterland	None	A-(Bk)-C	Platy peds, surface erosion mounds and swales	Entisol, Inceptisol	Brown & Veum, 1974	Feakes & Retallack, 1988; Retallack, 1990
Swamp	Large, in tabular mat	O-A-(Bg)-C	Woody peat or coal, noncalcareous	Histosol, Aqualf, Aquult	Ho & Coleman, 1967; Lytle, 1968	Retallack & Dilcher, 1981
Marsh	Fine, in tabuler mat	O-A-(Bg)-C	Herbaceous peat or coal, noncalcareous	Histosol, Aquoll	Gore, 1983	Dimichele et al., 1987
Salt marsh	Fine, in tabular mat	O-A-(Bg)-C	Herbaceous peat or coal, pyrite nodules, oysters and other marine shells	Histosol, Aquent, Aquept	Rabenhorst & Haering, 1989	Retallack, 1990
Carr	Large, in tabular mat	O-A-(Bg)-C	Woody peat or coal, calcareous	Histosol, Aquent, Aquept	Gore, 1983	Retallack & Dilcher, 1988
Fen	Fine, in tabular mat	O-A-(Bg)-C	Herbaceous peat or coal, calcareous	Histosol, Aquent, Aquoll	Gore, 1983	Retallack, 1990
Mangal	Large, in tabular mat	O-A-(Bg)-C	Woody peat or coal, pyritic nodules, oysters or other marine shells	Histosol, Aquent, Aquept	Chapman, 1977	Retallack & Dilcher, 1981

Table 3–3. Paleosol features of different topographic settings.

Topographic position	Root traces	Burrows	Soil structure	Microfabric	Other soil features	Studies of surface soils	Applications to paleosols
Hillslope	Deeply penetrating	Deeply penetrating	Oxidized peds (soil clods) and cutans	Sepic plasmic fabrics	Argillic (Bt) and calcic (Bk) horizons, soil creep of quartz veins or dipping beds, stone lines, colluvial mantles	Ciolkosz et al., 1990; Simon et al., 1990	Williams, 1968; Zbinden et al., 1988; Holland & Zbinden, 1988
Plateau	Deeply penetrating	Deeply penetrating	Oxidized peds and cutans	Sepic plasmic fabrics	Argillic (Bt) and calcic (Bk) horizons, stone lines, truncated profiles, thick laterites and other duricrusts	Goudie, 1973; McFarlane, 1976; Johnson & Watson-Stegner, 1987	Retallack, 1991b
Well-drained lowland	Deeply penetrating	Deeply penetrating	Oxidized peds and cutans	Sepic plasmic fabrics	Argillic (Bt) and calcic (Bk) horizons, mukkarra structure, sand wedges, ice wedges, fossil bones and land snail shells, hematite and other oxidized minerals	Gile et al., 1980; de Wit, 1978	Williams, 1986; Retallack, 1983a, 1986b, 1991b
Intermittently waterlogged lowland	Tabular & deeply penetrating	Mixed terrestrial and aquatic	Oxidized and reduced soil structures	Sepic and asepic microfabrics	Cumulic horizons, salt crusts, valley calcretes, mix of oxidized minerals such as hematite and gley minerals such as siderite	Harden, 1982; Walker & Butler, 1983; Cremeens & Mokma, 1986	Joeckel, 1988; Smith, 1990; Retallack, 1991b
Waterlogged lowland	Tabular mat	Of mainly aquatic creatures	Few relict peds and cutans, bedding	Undulic and other asepic fabrics	Coal, peat, carbonaceous shale, plant fossils, cumulic horizons, siderite or pyrite nodules	Ho & Coleman, 1967; Lytle, 1968	Retallack & Dilcher, 1981; Retallack et al., 1987; Gardner et al., 1988

Table 3–4. Paleosol features favored by different kinds of parent materials.

Parent material	Grain size	Minerals	pH	Soil types	Studies of surface soils	Application to paleosols
Till	Clayey, rocky	Smectite, calcite	Alkaline	Inceptisol, Mollisol, Alfisol	Birkeland, 1984	Retallack, 1980
Loess	Clayey, silty	Smectite, calcite	Alkaline	Inceptisol, Mollisol, Alfisol	Fehrenbacher et al., 1986	Retallack, 1980
Volcanic sand	Clayey, sandy	Smectite, magnetite	Alkaline	Inceptisol, Alfisol	Dethier, 1988	Retallack, 1983a, 1991b
Quartz sand	Sandy	Quartz, hematite	Acidic	Spodosol, Psamment, Dystrochrept	Thompson & Bowman, 1984; Schwartz, 1988	Batten, 1973; Retallack, 1977; Percival, 1986
Alluvium	Clayey	Smectite, kaolinite	Neutral	Inceptisol, Alfisol, Aridisol	Lepsch & Buol, 1974; Walker & Butler, 1983; Busacca & Singer, 1989	Feakes & Retallack, 1988
Marine shale	Clayey	Smectite, illite	Neutral	Inceptisol, Alfisol, Vertisol	Aandahl, 1978	Retallack, 1983a; Holland & Beukes, 1990
Schist	Clayey	Illite, chlorite	Neutral	Alfisol, Ultisol	England & Perkins, 1959; Marron & Popenoe, 1986	Holland, 1984
Limestone	Clayey, rocky	Calcite, kaolinite	Neutral	Entisol, Inceptisol, Oxisol	Ahmad & Jones, 1969; Scholten & Andriesse, 1986	James & Choquette, 1987; Retallack, 1991b
Granitic rocks	Sandy	Quartz	Acidic	Spodosol, Ultisol, Oxisol	Dixon & Young, 1981; Rutherford, 1987	Grandstaff et al., 1986
Basaltic rocks	Clayey	Smectite, magnetite	Neutral	Inceptisol, Alfisol, Vertisol	Lepsch & Buol, 1974; Craig & Loughnan, 1964	Holland et al., 1989; Retallack, 1991b
Ultramatic rocks	Rocky	Serpentine, pyroxene	Neutral	Entisol, Inceptisol	Garcia et al., 1974; Alexander, 1988	Williams, 1968
Volcanic ash	Clayey	Allophane, halloysite	Neutral	Andisol	Neall, 1977; Tan, 1984	Retallack, 1983a, 1991b

Table 3-5. Paleosol features related to time for formation.

Features	Degree of development: Very weak 10^2 yr	Weak 10^3 yr	Moderate 10^4 yr	Strong 10^5 yr	Very strong 10^6 yr	Studies of surface soils	Paleosols
Subsurface calcareous (Bk) horizon morphology	None	Wisps	Nodules	Layer	Layer with pisolites, laminae	Gile et al., 1966; Hay & Reeder, 1978; Machette, 1985	Leeder, 1975; Retallack, 1991b
Subsurface clayey (Bt) horizon morphology	Relict bedding or other rock structure	Some clay skins	Clay skins, peds (soil clods), sepic fabric	Pervasive sepic microfabric within peds	Thick (>2 mm) well-structured, omnisepic	Harden, 1982; Birkeland, 1990	Retallack, 1983a, 1986a, 1990
Surface peaty (O) horizon thickness	0–4 cm	4–40 cm	40 cm–4 m	4–40 m	>40 m	Moore & Bellamy, 1973; Falini, 1965	Retallack, 1990
Weathering rind thickness (mm)							
on basalt	0	<0.05	0–1.4	0.6–2	>2	Colman, 1986	–
on andesite	0	<0.1	0–1.2	0.8–2	>2	Colman, 1986	–
on granite	0	<0.2	0.1–1.7	0.3–3	>3	Birkeland, 1984; Hall & Michaud, 1988	–
Grain morphology							
of pyroxene	Pitted	Etched	Very etched	Gone	Gone	Birkeland, 1984	Retallack, 1991b
of hornblende	Fresh	Pitted	Pitted	Etched	Very etched	Hall & Michaud, 1988	–
of quartz	Fresh	Fresh	Fresh	Pitted	Etched	Cleary & Conolly, 1971	–
Harden index (SDI)	0–5	3–12	8–120	20–250	40–250	Harden, 1982, 1990	

paleosols are disfigured by three widespread alterations that occur very early during burial; decompositional loss of soil organic matter (Stevenson, 1969), formation of drab-colored gleyed haloes around remnant organic matter (de Villiers, 1965; Allen, 1986b) and dehydration reddening of ferric hydroxides (Simonson, 1941; Ruhe, 1969). These alterations may transform a gray/brown soil to a gaudy green/red mottled paleosol with minimal amounts of organic matter (Retallack, 1991a). As a consequence, such soil features as degree of reddening and amounts of organic matter and soil N cannot easily be used to interpret paleoenvironment of paleosols, even though they correlate with climatic variables in surface soils (Jenny, 1941; Birkeland, 1984).

The degree of clayeyness of soils and their depth of weathering also are favorites for the development of climofunctions from modern soils, and these are variables that can be measured in paleosols. In this case, however, their application to paleosols is defeated by the unknown relative contributions of climatic temperature and rainfall, and of time for formation, all of which conspire to create clayey and deeply weathered soils.

ALTERNATIVES TO THE FACTOR-FUNCTION APPROACH

Taxonomic Uniformitarianism

One alternative to the factor-function approach that has proven useful in paleoenvironmental interpretation of paleosols is an approach similar to that widely known among paleontologists as taxonomic uniformitarianism (Dodd & Stanton, 1990). If, for example, fossil bones are identified as those of a large fossil alligator or a fossil snail is identified as that of a large helminthoglyptid, then assuming that the fossil creatures had ecological tolerances similar to their living relatives, a warm, moist, frost-free paleoclimate is indicated. Similarly, identification of a paleosol within a modern soil taxonomy may be taken to imply past conditions similar to those enjoyed by such soils today. In my own research of this kind (Retallack, 1983a, 1991b), I have simultaneously tried to employ several different soil classifications, such as those of the U.S. Soil Conservation Service (Soil Survey Staff, 1975), the Food and Agriculture Organization of UNESCO (U.N.FAO, 1971–1981), the South African Department of Agricultural Technical Services (MacVicar et al., 1977), and the Australian CSIRO (Stace et al., 1968). The approach also has been strengthened by the simultaneous comparison of genetically related suites of paleosols, such as the individual rock units of Fig. 3-4. These can be compared as a landscape unit to soil mapping units in local soil surveys or regional soil maps such as those of U.N.FAO (1971–1981). Taxonomic study of paleosols is a very effective way of locating analogous modern profiles and landscapes within the vast descriptive literature on soils for more detailed comparison with specific paleosols. It also is a useful check on inferences based on the factor-function approach.

It can be argued that paleosols should not be identified in classifications of surface soils, which were not designed for this kind of use (Fastovsky & McSweeney, 1989) and are fundamentally different from biological classifications (Fastovsky, 1991). However, these do not seem to be the most serious problems associated with taxonomic uniformitarian interpretation of paleosols. Classification of paleosols should be a better guide to paleoenvironments than biological classification of fossils, because soil taxonomy is based primarily on environmentally significant features. Organisms on the other hand, are classified by a mix of characteristics, some apparently are adaptations to environment, and others primarily reflecting evolutionary ancestry. As a paleontologist by training, I am quite comfortable with the identification of fossil alligators and snails, even though soft part anatomy would be necessary for a zoologist to feel comfortable with an identification. The identification of paleosols in soil classifications demands comparable assumptions that some soil scientists may find uncomfortable. For example, Aridisols cannot be identified among paleosols from pre-existing paleoclimatic data, because that would make circular the interpretation of paleoclimate from them. In view of the relationships here presented (Figs. 3–1, 3–2) a suitable paleopedological definition of Aridisols would include paleosols with carbonate nodules at a reconstructed depth of less than about 1 m. Such a focus on soil features, rather than climatic variables, would eliminate a persistent element of circular reasoning remaining in "Soil Taxonomy" itself (Soil Survey Staff, 1975). A start has been made in modifying the criteria of soil classification to criteria that can reasonably be expected in paleosols (Retallack, 1988b, 1990), and a consensus view of reasonable criteria will emerge, just as paleontologists have developed for fossil identification.

Although soil classifications were designed in part for agricultural and other human uses, they have fundamental underpinnings in genetic concepts from the accumulated experience of several generations of scientists that have thought very deeply about soils (Buol et al., 1989). My use of a variety of soil classifications has impressed on me their similarity in stressing such features as degree of development, and the importance of such materials as peat, carbonate and clay (Retallack, 1990). They do organize data in a way that is useful for trying to understand a paleosol.

The main problem with taxonomic study of paleosols is not these philosophical difficulties but the practical result that the paleoenvironmental implications of taxonomic study are imprecise and difficult to quantify. Even units deep within the taxonomic hierarchy of soil classification or of landscape assemblages of soils commonly have wide environmental range. If the classification is followed out to find specific surface soils analogous with a paleosol, they are seldom identical in all respects (Retallack, 1991b). Each soil and landscape has endured an individual history.

Another problem is the jargon of soil classification. Just as there was resistance among soil scientists to the introduction of "Soil Taxonomy" (Soil Survey Staff, 1975), some geologists feel that it is an unreasonable imposition to have to master the language of soil science (Fastovsky, 1991). Some familiarity with these terms is very useful for publishable research in paleopedology, but for teaching the uninitiated it may be necessary to abridge this formidable lexicon and use equivalent common English terms (some are suggested by Retallack, 1990).

Numerical Models

There has long been a conceptual framework for mathematical modeling of soil formation. Soils can be viewed as energy transformers, that is a body of material changed by the continuing efforts of natural processes (Runge, 1973). They also can be envisaged as open systems to the extent that they represent a boundary between earth and air through which materials move and are transformed (Simonson, 1978). Recent numerical models of the formation of carbonate horizons within soils provide examples of both these flux (Machette, 1985) and process models (McFadden & Tinsley, 1985; Mayer et al., 1988). These particular models have not to my knowledge been applied to buried paleosols, although there is potential to do so. Other models have been usefully employed in interpreting paleoenvironment from paleosols, such as one specifying atmospheric control of C isotopic systematics of paleosol carbonate (Cerling et al., 1989; Quade et al., 1989; Cerling, 1991; McFadden et al., 1991) and another estimating atmospheric oxidation from chemical weathering of paleosol silicate minerals (Holland, 1984; Holland & Zbinden, 1988; Pinto & Holland, 1988; Zbinden et al., 1988; Holland et al., 1989; Holland & Beukes, 1990). Research continues on the conceptual framework of such models (Johnson & Watson-Stegner, 1987; Johnson et al., 1987; Anderson, 1988; Johnson, 1990) and several excellent textbooks also are available on numerical modeling of soil formation (Richter, 1987, 1990; Ross, 1989). Their application is bound to become more widespread as personal computers proliferate.

The principal problem with numerical modeling of paleosols is the assumptions on which they are based. The two examples already mentioned (Cerling, 1991; Holland, 1984) are typical in requiring data on such conditions as original soil porosity, moisture content and biologically respired CO_2. Such information may not be obtained directly from paleosols. Reasonable values and limits can be applied from modern soils to force the models to perform, but how these values resonate through complex equations modeling a paleosol is not always straightforward. Many other models for modern soil formation are rate models, of the general form $\delta x/\delta t$, where x is some measured feature of the soil, and t is the time in years for its formation. Unfortunately, time for formation can only be estimated to an order of magnitude for paleosols (Table 3–5), and such wide error limits are spread even wider from the denominator of complex equations. In contrast, Jenny's (1941) formulation of time as an independent variable obviates this problem so that it is well suited to paleopedological studies, although Jenny's work has been criticized on this basis (Yaalon, 1975). Because of their assumptions, the results of modeling can only be cautiously accepted. Both the models of Cerling (1991) and Holland (1984) have failed in application to specific paleosols because some assumptions were not tenable (Retallack, 1986b, 1992b). This problem is likely to become more severe as models evolve into animated computer games, and the seductive delight of playing obscures the underlying assumptions.

A second problem for the widespread use of numerical modeling is the advanced mathematical and computer skills that they currently demand. Gaining a working understanding of even such simple models as those of Cerling (1991) and Holland (1984) requires an intellectual effort at least comparable to that of learning the language of soil taxonomy. This is not an effort many soil scientists and geologists are going to make until the models become more user friendly. It will also restrict the teaching of numerical modeling to advanced university classes.

SUMMARY

After 50 yr, the factor-function approach to the study of soils is still thriving, and as indicated here, finding new applications to the paleoenvironmental interpretation of paleosols. The relationship between mean annual rainfall and depth to carbonate nodules in soils popularized in Hans Jenny's (1941) book is a good example of a climofunction that can be used to interpret paleoclimate from calcareous paleosols. A new compilation offered here has shown that this is a surprisingly robust and widespread relationship within aridland soils.

Ongoing research on a variety of other relationships between environmental variables and soil features is providing a vast fund of other information of use for the paleoenvironmental interpretation of paleosols. The principal limitation on the use of the factor-function approach to paleosols are the profound alteration of some paleoenvironmentally sensitive features of soils after their burial and the control of some soil features by several environmental variables.

Alternatives to the factor-function approach include identifying a paleosol within a soil classification in order to infer a paleoenvironment similar to modern analogs, or numerically simulating soil-forming processes of a paleosol. Paleosol identification commonly lacks precision and numerical models often founder on assumptions. Both approaches require a considerable effort to master either terminology or mathematics. Much of our understanding of soil classification and of numerical modeling of soils derives from studies of soil-forming factors, so the factor-function approach is unlikely to be supplanted by these other approaches in the foreseeable future.

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APPENDIX 3-1

Data for relation between mean annual rainfall and depth of calcic horizon in format: source, rainfall (mm), and depth of carbonate (cm).

1,400,-85	7,115,-17	8,615,-106	8,505,-62	17,315,-43	28,995,-190
1,250,-30	7,117,-22	8,755,-106	8,520,-62	17,259,-35	28,953,-245
1,220,-25	7,240,-22	8,895,-106	8,570,-62	17,305,-45	28,680,-60
1,180,-12	7,130,-24	8,900,-110	8,585,-62	17,298,-17	28,600,-20
1,100,-35	7,180,-24	8,635,-120	8,625,-62	17,298,-20	29,700,-100
2,380,-25	7,245,-24	8,732,-120	8,665,-62	17,240,-60	29,700,-70
2,410,-30	7,210,-25	8,840,-125	8,770,-62	17,305,-25	29,365,-30
2,635,-90	7,360,-27	8,855,-125	8,560,-71	17,259,-35	29,160,-5
2,430,-20	7,180,-33	8,765,-142	8,355,-77	17,305,-30	30,635,-120
2,430,-90	7,120,-35	8,840,-145	8,515,-77	17,298,-83	31,890,-180
2,660,-60	7,210,-35	8,980,-149	8,600,-77	18,257,-44	31,814,-330
2,660,-30	7,270,-38	8,715,-152	8,625,-77	18,276,-30	31,807,-210
2,685,-97	7,165,-41	8,745,-152	8,690,-77	18,257,-60	31,682,-175
2,635,-127	7,280,-43	8,875,-152	8,570,-81	18,276,-50	31,693,-140
2,535,-90	7,455,-43	8,895,-152	8,597,-81	18,276,-56	31,594,-80
2,610,-51	7,260,-51	8,910,-152	8,690,-81	18,276,-60	31,559,-30
2,510,-90	7,240,-53	8,380,0	8,490,-84	18,276,-48	31,519,-40
2,510,-40	7,260,-53	8,410,0	8,707,-84	18,257,-46	31,429,-10
2,305,-42	7,530,-53	8,370,-22	8,720,-84	19,380,-30	32,750,-70
2,305,-60	7,315,-60	8,380,-22	8,760,-84	19,440,-45	32,750,-60
2,330,-40	7,380,-67	8,395,-22	8,355,-92	19,610,-85	32,750,-76
2,380,-70	7,430,-83	8,385,-25	8,540,-92	19,620,-80	33,82,-21
2,510,-90	7,530,-97	8,375,-25	8,585,-92	19,580,-30	33,178,-27
2,510,-80	7,435,-118	8,390,-27	8,625,-92	19,580,-50	33,178,-33
3,650,-85	8,762,-153	8,420,-30	9,500,-110	19,580,-45	33,178,-44
3,820,-134	8,785,-153	8,680,-30	10,350,-65	20,188,0	33,180,-46
3,775,-106	8,990,-153	8,400,-31	10,350,-55	20,45,0	33,200,-41
3,675,-94	8,730,-154	8,410,-31	11,430,-97	20,188,0	33,200,-58
3,300,-76	8,770,-167	8,455,-31	11,325,-35	21,127,-6	33,225,-46
3,350,-44	8,820,-182	8,472,-31	11,97,-34	22,333,-71	33,225,-49
4,400,-81	8,850,-182	8,540,-31	12,664,-191	22,333,-66	33,280,-36
5,520,-70	8,870,-182	8,560,-34	12,730,-270	23,377,-31	33,280,-48
5,530,-75	8,980,-182	8,425,-38	13,500,-147	23,377,-38	33,310,-64
5,550,-60	8,920,-213	8,410,-40	14,400,-40	23,455,-51	33,310,-69
5,770,-110	8,985,-243	8,445,-40	14,400,-130	23,455,-43	33,333,-67
5,770,-80	8,920,-254	8,460,-40	14,400,-110	23,377,-46	33,420,-72
5,710,-95	8,950,-275	8,580,-40	15,483,-60	23,377,-18	34,600,-70
5,685,-114	8,895,-298	8,475,-43	15,483,-95	23,377,-23	35,730,-114
5,650,-70	8,910,-300	8,510,-45	15,500,-105	23,377,-38	35,730,-95
5,650,-80	8,960,-300	8,390,-48	15,213,-23	23,455,-19	35,750,-99
5,770,-120	8,640,-92	8,405,-48	15,177,-28	24,333,-61	35,830,-107
6,760,-60	8,655,-92	8,460,-48	15,177,-25	24,607,-51	36,580,-59
6,760,-81	8,670,-92	8,540,-48	15,177,-60	24,420,-38	37,325,-45
6,760,-100	8,705,-92	8,550,-48	15,177,-41	25,546,-97	38,476,-64
6,760,-161	8,730,-92	8,430,-49	16,513,-40	25,419,-91	38,476,-45
6,760,-60	8,660,-94	8,570,-49	16,443,-76	26,399,-46	38,476,-67
6,750,-58	8,690,-94	8,440,-52	16,570,-76	26,399,-23	38,476,-80
6,730,-80	8,775,-94	8,670,-52	16,389,-91	26,399,-54	38,476,-67
6,730,-87	8,820,-94	8,410,-62	16,435,-76	27,706,-92	38,176,-26
6,620,-50	8,850,-95	8,425,-62	17,298,-80	27,522,-45	39,183,-15
6,620,-30	8,765,-102	8,440,-62	17,298,-80	27,427,-50	39,178,-20
7,110,-12	8,795,-102	8,465,-62	17,305,-63	27,526,-40	39,183,-24
7,140,-13	8,730,-105	8,480,-62	17,305,-30	27,425,-70	39,170,-24

(continued on next page)

APPENDIX 3-1. Continued.

39,173,-40	39,170,-15	40,189,-3	41,480,-44	41,940,-187	41,770,-120
39,180,-18	39,173,-45	40,324,-38	41,360,-65	41,918,-135	41,820,-75
39,183,-14	39,180,-12	40,184,-40	41,290,-50	41,715,-57	41,750,-110
39,195,-32	39,174,-15	40,91,-22	41,330,-50	41,747,-52	41,830,-125
39,193,-19	39,176,-20	41,705,-77	41,290,-58	41,740,-65	41,725,-100
39,185,-28	39,176,-20	41,745,-60	41,320,-50	41,680,-46	41,330,-60
39,176,-10	39,190,-10	41,580,-45	41,320,-34	41,770,-85	41,400,-60
39,176,-22	40,300,-115	41,490,-72	41,320,-52	41,890,-136	41,750,-85
39,183,-25	40,225,-22	41,380,-45	41,330,-45	41,815,-100	41,720,-75
39,183,-23	40,200,-15	41,490,-60	41,953,-154	41,770,-92	
39,183,-31	40,105,-18	41,380,-33	41,950,-168	41,770,-98	

† Sources: 1 = Dan & Yaalon, 1982; 2 = Stace et al., 1968; 3 = Murthy et al., 1982; 4 = Sehgal et al., 1968; 5 = de Wit, 1978; 6 = Jager, 1982; 7 = Arkley, 1963; 8 = Jenny & Leonard, 1935; 9 = U.N.FAO, 1971; 10 = U.N.FAO, 1981; 11 = U.N.FAO, 1977; 12 = Sidhu et al., 1977; 13 = Ahmad et al., 1977; 14 = Courty & Féderoff, 1985; 15 = Glinka, 1931; 16 = Dokuchaev, 1883; 17 = Fedorin, 1960; 18 = Nogina, 1976; 19 = Thorp, 1936; 20 = Campbell & Claridge, 1987; 21 = Tedrow, 1970, 1977; 22 = McCraw, 1964; 23 = Raeside & Cutler, 1966; 24 = Soil Bureau Staff, 1968; 25 = Leamy & Sanders, 1967; 26 = Orbell, 1974; 27 = del Villar, 1957; 28 = Soil Correlation Committee for South America, 1967; 29 = Al Taie et al., 1969; 30 = Bronger & Heinkele, 1989; 31 = Ruhe, 1984; 32 = Hussain et al., 1984; 33 = Marion et al., 1985; 34 = Fadda, 1968; 35 = Bhargava et al., 1981; 36 = Vinayak et al., 1981; 37 = Bal & Buursink, 1976; 38 = Siderius, 1973; 39 = Mulders, 1969; 40 = Dan et al., 1981; 41 = Plaza & Moscatelli, 1989.

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