Earth-Science Reviews 47 (1999) 41–70

EARTH-SCIENCE REVIEWS

www.elsevier.com/locate/earscirev

Paleosols in clastic sedimentary rocks: their geologic applications

Mary J. Kraus *

Department of Geological Sciences, University of Colorado, Boulder, CO, 80309-0399, USA

Received 22 July 1998; accepted 24 March 1999

Abstract

Interest in pre-Quaternary paleosols has increased over the past decade, in large part, because they have proved to be beneficial in solving diverse geological problems. The majority of paleosols are described from continental deposits, most commonly from alluvial strata. Criteria for recognizing these paleosols have been extensively described; however, classifying them has proved more complicated. Pre-Quaternary paleosols are generally classified according to one or more modern soil classification systems, although one new classification has been proposed exclusively for paleosols to avoid problems using the modern soil classifications. In addition to taxonomic classification, paleosols can be categorized according to the interplay among deposition, erosion, and the rate of pedogenesis when they formed. Paleosols can be solitary if they formed during a period of landscape stability following the development of an unconformity. Such paleosols are commonly thick and extremely well developed. More commonly, paleosols are vertically stacked or multistory because they formed in sedimentary systems undergoing net aggradation. If erosion was insignificant and sedimentation was rapid and unsteady, compound paleosols generally formed. If the rate of pedogenesis exceeded the rate of deposition, composite paleosols developed. Thick, cumulative paleosols indicate that erosion was insignificant and that sedimentation was relatively steady. Both autogenic and allogenic processes can influence depositional and erosion patterns and, thus, affect the kinds of soils that form. Consequently, paleosols can help to interpret the history of sediment deposition and the autogenic and allogenic processes that influenced a sedimentary basin. Paleosols are also helpful in stratigraphic studies, including sequence stratigraphic analyses. They are used for stratigraphic correlations at the local and basinal scale, and some workers have calculated sediment accumulation rates based on the degree of paleosol development. In addition to their stratigraphic applications, paleosols can be used to interpret landscapes of the past by analyzing paleosol–landscape associations at different spatial scales, ranging from local to basin-wide in scope. At the local scale, lateral changes in paleosol properties are largely the result of variations in grain size and topography. At the scale of the sedimentary basin, paleosols in different locations differ because of basinal variations in topography, grain size, climate, and subsidence rate. Paleosols are used to reconstruct ancient climates, even to estimate ancient mean annual precipitation (MAP) and mean annual temperature (MAT). Ancient climatic conditions can be interpreted from modern soil analogs or by identifying particular pedogenic properties that modern studies show to have climatic significance. Stable carbon and oxygen isotopes are also used to interpret ancient climate, and some effort has been made to estimate MAT from isotopic composition. On the basis of

* Tel.: +1-303-492-7251; Fax: +1-303-492-2606; E-mail: mary.kraus@colorado.edu

PII: S0012-8252(99)00026-4

modern soil analogs, paleo-precipitation has been estimated from the depth at which calcic horizons originally formed. Finally, paleosol carbonates have been used to estimate ancient atmospheric CO_2 values.

Keywords: paleosols; clastic rocks; paleoclimatology; lithostratigraphy; terrestrial environment

1. Introduction

A paleosol or fossil soil is a soil that formed on a landscape of the past. Soils form because of the physical, biological, and chemical modification of sediment or rock exposed at the earth surface. Most paleosols are found in sedimentary rocks, and, although first studied in the Quaternary record, they are now commonly recognized in strata as old as Precambrian. Soils and paleosols can form because of lengthy episodes of landscape stability, in which case they may eventually mark a stratigraphic diastem or unconformity; they can also form in terrestrial depositional systems that are aggrading as long as the rate of sedimentation does not overwhelm the rate of pedogenesis. Soils and paleosols thus reflect a complex interplay among sedimentation, erosion, and non-deposition.

Paleosols have been described from a variety of continental depositional settings including eolian (e.g., Soreghan et al., 1997), palustrine (e.g., Tandon et al., 1995; Wright and Platt, 1995; Tandon and Gibling, 1997), and deltaic (e.g., Fastovsky and McSweeney, 1987; Arndorff, 1993). Paleosols are also found in marginal marine strata (Lander et al., 1991; Wright, 1994) and can appear in marine strata if sea-level fell to expose marine sediment (e.g., Driese et al., 1994; Webb, 1994). Paleosols are most commonly described from the alluvial record; consequently, they provide many of the examples discussed in this paper. To constrain the scope of this review article, discussion centers on, but is not limited to, paleosols that formed in clastic sedimentary successions that are Phanerozoic but pre-Quaternary in age. Furthermore, because paleosol studies expanded during the 1990's in terms of their number, sophistication, and kinds of geologic applications, most of the examples discussed here are taken from these recent investigations.

The current interest in pre-Quaternary paleosols has arisen because paleosols have proven remarkably useful for examining an array of geological problems, and the major goal of this paper is to examine some of the exciting areas of geological research that benefit from analysis of paleosols. First, deposition is central to the development of sedimentary paleosols, and processes that control the rate and continuity of sediment accumulation influence the degree of pedogenic development. Consequently, paleosols can help to interpret the history of sediment deposition and the allogenic processes that influenced the sedimentary basin. Paleosols are also used in stratigraphic analyses, and one of their more recent applications is to sequence stratigraphy. Some workers have also used paleosols to calculate short-term sediment accumulation rates, both qualitatively and quantitatively.

Other geologic applications of paleosols range from landscape reconstruction to paleoclimatic studies. Paleosols provide detailed insight into ancient landscapes and landscape evolution because the spatial distribution of different paleosols reflects the particular landforms on which they formed and the geomorphic processes operating in the ancient landscape. The paleoclimatic importance of paleosols stems from studies of Quaternary soil development, which have shown that some pedogenic features can be quantitatively related to soil-forming factors such as climate. Paleosols are being used not only to interpret ancient climatic regimes but, in some cases, to estimate paleo-precipitation and paleo-temperature. Paleosols are also proving helpful in better understanding the composition of the ancient atmosphere and atmospheric changes over geologic time. Finally, paleosols have importance for studying the advent and evolution of terrestrial plants and animals.

Because criteria for recognizing paleosols and for distinguishing pedogenesis from diagenesis have been described in detail, the reader is referred to those sources (e.g., Retallack, 1991, 1997b; Wright, 1992a; Pimentel et al., 1996) rather than providing a synopsis here. One of the noteworthy advances of the last decade is that paleosols are increasingly recognized in cores based on various morphological and geochemical features (e.g., Leckie et al., 1989; Lander et al., 1991; Platt and Keller, 1992; Caudill et al., 1997). Paleosols even have been identified on the basis of wireline log signatures although this approach is difficult (Ye, 1995). Classifying paleosols has received far less attention than recognizing them, and this paper begins by discussing the classifications commonly used for paleosols and the strengths and weaknesses of these classifications.

2. Classifying paleosols

Most paleosol workers use one or more modern classification systems including United States Soil Taxonomy (Soil Survey Staff, 1975, 1998) and the FAO (1974) classification. The classification of Duchaufour (1982) has been used by some (e.g., Besly and Fielding, 1989; Kraus, 1997). One classification, that of Mack et al. (1993), is specific to paleosols, although it is based on modern soil classifications.

The U.S. classification and the FAO classification are taxonomic systems that use profile characteristics to classify the soils. The U.S. system relies on diagnostic horizons that are identified on the basis of properties such as texture, color, amount of organic matter, presence of particular minerals, cation exchange capacity, and pH. The FAO classification resembles the U.S. system in using diagnostic horizons; however, as Duchaufour (1982) pointed out, the FAO system is less complicated. Furthermore, it recognizes hydromorphic soils as a major group, the Gleysols, whereas soil saturation is only considered at the sub-order level in the U.S. system (Table 1). This difference is important with paleosol classification because floodplain paleosols are common and, like their modern counterparts, many are hydromorphic (e.g., Fastovsky and McSweeney, 1987; Besly and Fielding, 1989; Arndorff, 1993; Kraus and Aslan, 1993). Duchaufour also argued that the U.S. and FAO systems suffer by not considering the soil environment more heavily. He proposed an environmental classification in which soil properties are considered in terms of the particular processes of soil formation that operate under particular environmental conditions. Although this classification uses the diagnostic horizons found in the U.S. and FAO systems, it differs from those in that it emphasizes that classification cannot rely on individual horizons. Rather, all the horizons in a particular paleosol are genetically related and must be used together in classification. The Duchaufour classification is interpretative in that the attributes of the soil horizons are used to interpret the processes and environmental conditions of soil formation. The processes are then used to classify the soil. For example, his Division II soils are characterized by the formation of sesquioxides. Within this division, the kind of weathering and the degree of weathering serve to distinguish three classes: Fersiallitic soils, Ferruginous soils, and Ferrallitic soils (Table 1).

Table 1
Comparison of the Mack et al. (1993) classification of paleosols and Soil Taxonomy (Soil Survey Staff, 1975, 1998). Some groups in the Mack classification have equivalents in Soil Taxonomy; others, such as Calcisols and Gypisols, do not. The Duchaufour (1982) classification is also shown but not compared to the other two

Mack et al.	Soil survey staff	Duchaufour
Protosol	Entisol Inceptisol	I. Slightly developed soils
Vertisol	Vertisol	II. Desaturated humic soils
Histosol	Histosol	III. Calcimagnesian soils
Gleysol	not a great order	IV. Isohumic (steppe) soils
Excluded	Andisol	V. Vertisols
Oxisol	Oxisol	VI. Brunified soils
Spodosol	Spodosol	VII. Podzolised soils
Argillisol	Alfisol Ultisol	VIII. Hydromorphic soils
Calcisol	no direct equivalent	IX. Ferisiallitic soils
Gypisol	no direct equivalent	X. Ferruginous soils
No direct equivalent	Aridisol	XI. Ferrallitic soils
Excluded	Mollisol	XII. Salsodic soils
No direct equivalent	Gelisols	

One drawback to using the U.S. or FAO systems is their dependence on soil properties, such as cation exchange capacity or amount of organic matter, that are not preserved in paleosols. Modern classifications also rely on knowledge of the climatic conditions under which a soil formed. Primarily because of these problems, Mack et al. (1993) proposed a classification just for paleosols. This system relies on the presence of stable minerals and morphological properties that tend to be preserved as a soil is transformed to a paleosol. Some of the major soil groups in the Mack classification are identical to those in U.S. classification (Table 1). Some categories in the U.S. scheme were excluded because of difficulties in recognizing them in the ancient record. For example Aridisols were excluded because this taxonomic class relies on knowledge of the soil moisture regime. Several new categories were proposed, some of which have approximate equivalents in the U.S. scheme (e.g., Protosol). Others, such as Calcisols, which are distinguished on the basis of a calcic horizon, have no direct equivalent. Gelisols were added to Soil Taxonomy after development of the Mack et al. classification (Soil Survey Staff, 1998). These are soils that have permafrost within a particular depth or have gelic materials, which are the result of cryopedogenic processes.

The Mack et al. (1993) system has been received with some favor because it is designed for field identification of paleosols and because it makes paleosol classification more objective and simpler and, thus, achievable to a broader group of geologists. Despite these advantages, the system has flaws, which Retallack (1993) outlined. One concern is that, because of its restriction to paleosols, the use of this classification will weaken the communication between soil scientists and paleopedologists. A second concern is that, because paleopedologists rely on modern soil analogs to interpret ancient environmental conditions, the environmental value of paleosols will be diminished by a classification that is specific to paleosols.

Why classify paleosols? In general, scientific classifications serve to organize information and to foster effective communication about a particular subject. Classifications also provide guidelines for future studies in a particular subject by emphasizing what factors or properties are important. With paleosols, we want a classification system or systems that accomplish these goals. But, because paleosols are studied in order to help interpret past conditions and events including climates, depositional conditions, and paleoecological changes, Retallack (1993) is right that the classification must be firmly tied to modern soil systems. The U.S. and FAO classifications do have drawbacks for paleosol classification, primarily the fact that these are taxonomic classifications that rely on too many features that are commonly absent in the ancient record. In many ways, the Duchaufour (1982) classification is more easily applied to paleosols because of its focus on **process** rather than modern soil properties. Morphological and geochemical properties that are preserved in the rock record are usually sufficient to determine the soil forming processes that occurred and then to classify the paleosol. Another advantage of the Duchaufour classification is that it recognizes soil intergrades, which are soils that occur between two classes. For many paleopedologists, the major deterrent to using the Duchaufour classification is probably that many of the soil categories are so dissimilar from the commonly used U.S. categories and the names themselves are so different (Table 1). Yet, if a researcher can get beyond the unfamiliar terminology, the process-oriented approach of this system can be effective for classifying paleosols.

3. Paleosols and time

The kind of paleosol that forms in the sedimentary record depends on how rapidly the sediment accumulated, whether that accumulation was steady or discontinuous, and, if pauses occurred, their duration. Sediment accumulation varies through time, producing different kinds of paleosols upward through a vertical succession (Fig. 1). Sedimentary paleosols form a continuum, at one end of which are multiple paleosols, which formed in relatively thick and conformable stratigraphic successions because aggradation was relatively continuous (Fig. 1B). But various autogenic processes (those that are inherent to the depositional system) and allogenic processes (those that are external to the depositional system) produce episodes of landscape stability or erosion. Depending on the particular process and the timescale over which it operated, stratigraphic gaps of various magnitudes (e.g., diastems and unconformities) will develop in the stratigraphic succession. Many of these surfaces will be marked by thick and very strongly developed paleosols, which are the other end member in the paleosol continuum (Fig. 1A). The following sections provide a more detailed

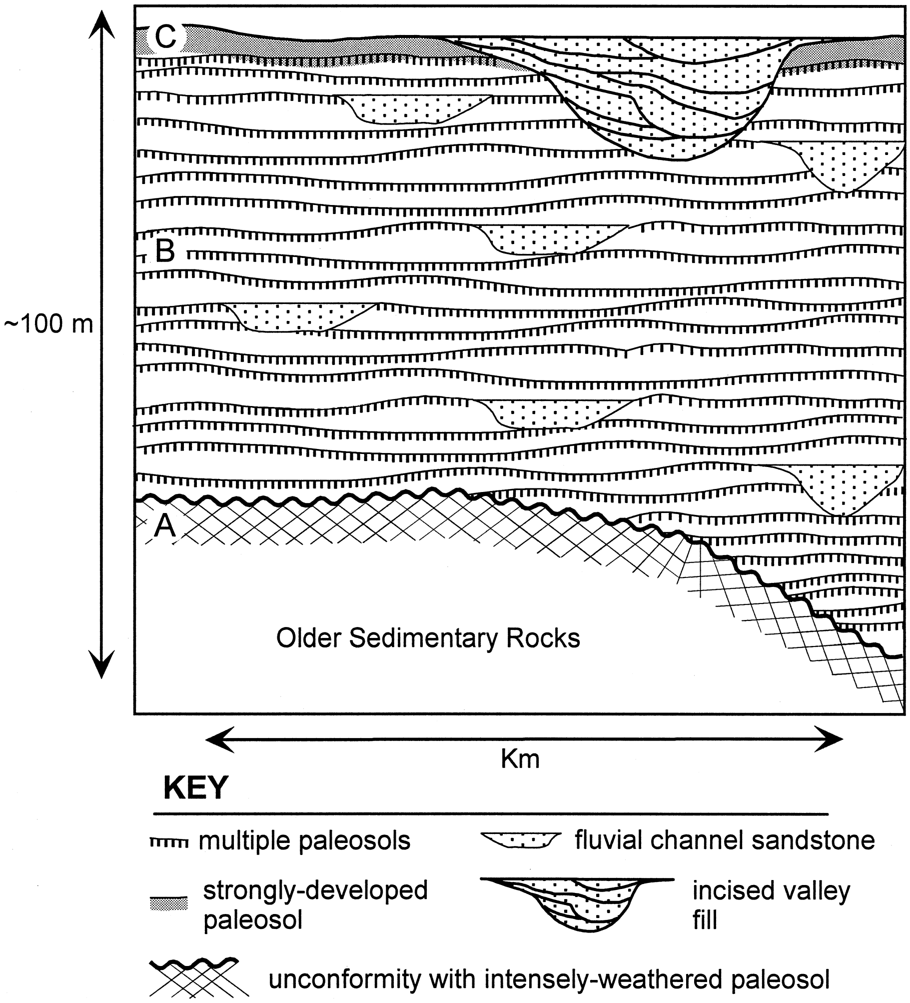


Fig. 1. Schematic diagram showing the range of paleosols that can form in a thick vertical section depending on whether sediment accumulation was steady or discontinuous, and, if pauses occurred, their duration. (A) A thick and strongly weathered paleosol formed on an unconformable surface because of a lengthy period of landscape stability and soil development. (B) A thick sequence of multiple paleosols formed on floodplain deposits because erosion was insignificant and sedimentation was steady. (C) A moderately long pause in sedimentation related to valley incision produced a paleosol that is more strongly developed than the multiple paleosols but not as intensely weathered as the paleosol at the unconformity. This paleosol has partly overlapped one of the underlying multiple paleosols.

description of paleosols of different temporal magnitudes and the processes by which they form.

3.1. Paleosols in aggradational systems

Paleosols can be classified according to the balance between sediment accumulation and the rate of pedogenesis (e.g., Morrison, 1978; Marriott and Wright, 1993; Wright and Marriott, 1996) (Fig. 2). If erosion is insignificant and sedimentation is rapid and unsteady, compound paleosols usually form (Fig. 2A). These are weakly developed, vertically stacked profiles that are separated by minimally weathered sediment. If the rate of pedogenesis exceeds the rate of deposition, vertically successive profiles may partly overlap, giving rise to composite paleosols. In contrast, if erosion is insignificant and sedimentation is steady, thick cumulative soils can form (Fig. 2B).

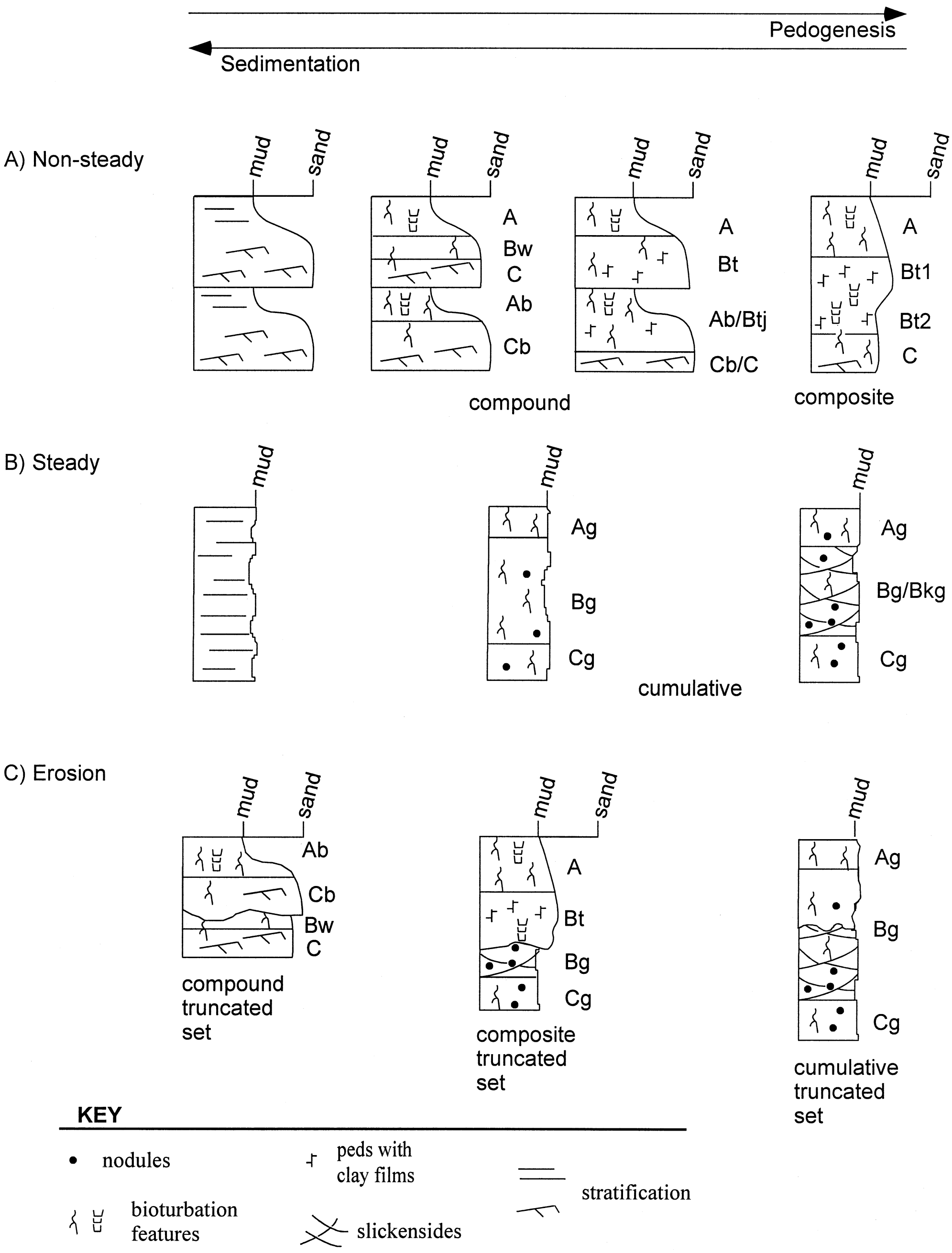


Fig. 2. Vertical profiles of sediments and soils (paleosols) reflecting varying rates of pedogenesis and sedimentation for (A) non-steady and (B) steady depositional conditions. In (C), sedimentation was interrupted by a period of erosion. Compound paleosols are likely when sedimentation is non-steady. Weakly developed cumulative profiles form when sedimentation is steady but rapid; better developed cumulative profiles form when sedimentation rates are slow relative to rates of pedogenesis. With erosion, a scour surface either separates two distinct paleosols or is incorporated into the paleosol. See text for more details. Ag = gleyed A horizon; Bg = gleyed B horizon; Bw = B horizon showing color or structure development but little if any illuvial accumulation; Bt = B horizon showing accumulation of clays; Btj = incipient development of a Bt horizon; Cg = gleyed C horizon (after Morrison, 1978; Bown and Kraus, 1981; Marriott and Wright, 1993).

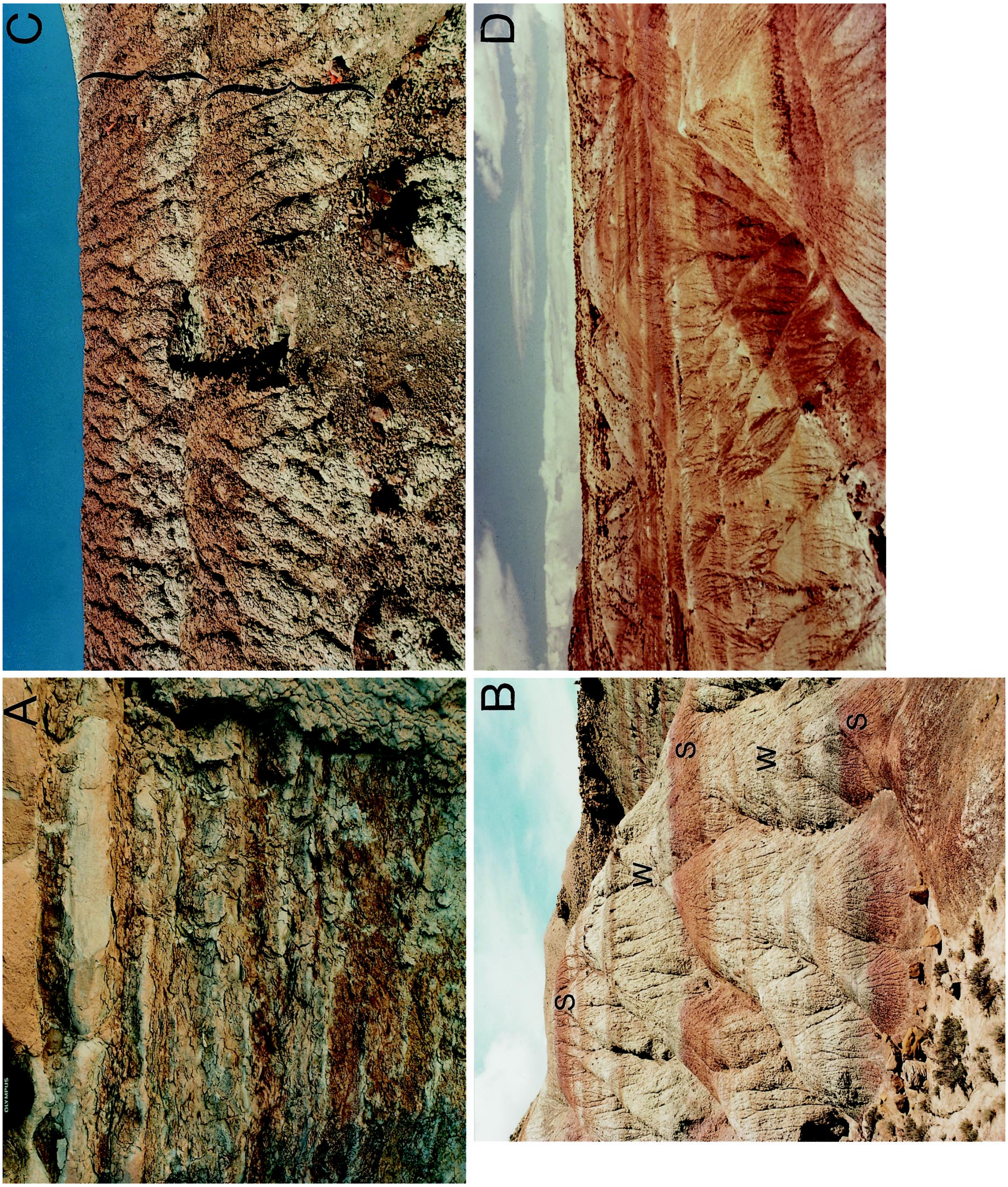
These profiles reflect the deposition of successive, thin increments of sediment accompanied by pedogenesis.

Using fluvial systems as an example, channel deposits may show little evidence of pedogenesis or may contain compound soils because sedimentation is so rapid (e.g., Marriott and Wright, 1993; Kraus and Aslan, 1999). As the channel migrates laterally over time, composite or well-expressed soils with Bt horizons may form in the upper deposits. Overbank deposition is usually steady but slow, commonly on the order of 1–10 mm/year (e.g., Walling et al., 1992), and overbank deposits tend to thin and decrease in grain size away from the active channel (Guccione, 1993; Marriott, 1996). As a result, compound soils with weakly expressed profiles tend to form on natural levees (Fig. 3A), whereas cumulative profiles develop in floodbasin areas distal to the active channel (Fig. 3B). Compound soils can also form in crevasse deposits because of rapid and unsteady deposition (Fig. 3C).

Autogenic and allogenic processes that operate over intermediate time-scales (10^3 to 10^4 years) can produce relatively lengthy pauses in deposition as well as local erosion, both of which can influence soil development. An example of the effects of an autogenic mechanism is provided by avulsion in a fluvial system. By causing channels to shift their positions on the floodplain, avulsion can terminate sedimentation on a particular part of the floodplain for periods of time that are probably on the order of 10^3 years, which is the periodicity of avulsion (e.g., Bridge and Leeder, 1979). A well-developed paleosol can form in floodplain areas cut off from sedimentation by this autogenic process. In the Paleogene Willwood Formation, episodic avulsions have produced stratigraphic intervals that are meters thick and that have two kinds of paleosols: (1) relatively well developed cumulative paleosols, which formed on floodbasin deposits and (2) very weakly developed compound or cumulative paleosols, which formed on the sediment deposited by the avulsion of the trunk channel (Kraus and Aslan, 1993; Kraus, 1996; Kraus and Gwinn, 1997) (Fig. 3B). Vertical paleosol sequences show alternations of the weakly developed and more strongly developed paleosols, and no allogenic controls need be invoked to generate such an alternation.

Turning to extrinsic mechanisms, climatic changes can also influence soil development at an intermediate time scale by initiating and terminating sedimentation. Many Quaternary loess successions show a vertical alternation between pedogenically-unmodified sediment and relatively well developed paleosols (e.g., Kukla and An, 1989; Frakes and Sun, 1994; Pecsi, 1995). In fact, these alternations provide a loess-paleosol stratigraphy (e.g., Pecsi, 1995). The paleosols are believed to have formed during periods of reduced sediment input associated with more humid climates (Fig. 4). During drier times, sediment input overwhelmed pedogenesis, and no soil formed. Although few pre-Quaternary examples of loessite and loess-paleosols have been described, an exception is provided by Soreghan et al. (1997) who observed alternations between paleosols and unmodified loessite horizons in the Late Paleozoic Maroon Formation. In this example, the paleosols show significantly higher magnetic susceptibilities than do the loessite horizons, which aids in their recognition. As with the Quaternary cases, the paleosols were linked to more humid climatic conditions and the unaltered loessite layers to drier climates (Fig. 4).

Although some processes merely halt sedimentation, other intermediate-scale mechanisms cause incision. Mechanisms like climate change can produce truncated soil profiles due to erosion of the upper part of a developing soil (e.g., Marriott and Wright, 1993) (Fig. 2C). In the sedimentary record, episodes of truncation are eventually followed by renewed sedimentation. If that sedimentation is slow and steady, a cumulative-truncated profile can result; however, if sedimentation is so rapid that the truncated soil becomes buried, a compound-truncated or composite-truncated set can develop (Marriott and Wright, 1993). In the former, the truncated paleosol is separated by unweathered sediment from an overlying paleosol; in the latter, the older, truncated paleosol and the younger paleosol partially overlap. In a study of upper Silurian and lower Devonian paleosols, Marriott and Wright showed how these different kinds of paleosols can be used to understand the complex processes that operate in ancient fluvial systems. They compared and contrasted two stratigraphic sections, one dominated by cumulative paleosols and one with numerous truncated paleosols. They inferred that the first section had a stable


C
D
A
B
S
W
S
W
S

floodplain that underwent slow but relatively continuous deposition. The other section had an unstable floodplain subject to periods of erosion, which might have been triggered by changes in climatic conditions or vegetation cover.

In fluvial systems, climatic change can also cause floodplain incision and terrace development. Although common in Quaternary fluvial systems, pre-Quaternary incised floodplains and terraces are often inferred, rather than directly observed (e.g., Wright, 1992b). This is presumably a problem of scale. Many outcrops are probably too laterally restricted to allow large cut-and-fill features to be recognized. Paleosols may be especially useful for recognizing ancient episodes of incision and terracing (Figs. 1C and 3D). For example, in a study of Eocene–Oligocene alluvial paleosols, Bestland (1997) recognized several episodes of terrace formation, because each is marked by a strongly developed paleosol indicating lengthy pauses in sedimentation. As discussed in a later section, paleosols are also of value for recognizing incised valleys and terraces that resulted from eustatic changes.

Finally, Retallack (1998) has recognized sawtooth patterns of development in sequences of paleosols. The paleosols in the sequence show a gradual upward increase in their degree of development until a major decline in development occurs. He provides an excellent discussion of the mechanisms that could produce this kind of pattern including: tectonic activity, sea-level fluctuations, and climatic changes.

3.2. Paleosols associated with unconformities

Unconformities usually represent significant hiatuses that may last millions or even tens of millions of years. They form when a period of landscape degradation and/or an episode of landscape stability is followed by sediment deposition. At least some period of landscape stability is needed for a soil to develop on the unconformable surface. In some examples, paleosols marking regional unconformities are exceptionally thick and well developed, indicating lengthy periods of soil development and landscape stability (Fig. 1A). For example, a Paleogene Oxisol described by Abbott et al. (1976) is 30 m thick and this profile thickness was regarded as a minimum because the upper contact is an erosional surface.

The formation of unconformities is controlled by allogenic factors such as sea level fluctuations, global or regional climate change, and regional tectonics, processes that influence geomorphic systems over time intervals of 10^5–10^7 years (e.g., Summerfield, 1991). For example, the thick Paleogene paleosol described above formed on a tectonically-generated unconformable surface (Abbott et al., 1976), and Driese et al. (1994) and Webb (1994) documented paleosols that mark unconformities developed on limestones exposed during regressions in the Late Paleozoic. Although Driese et al. found that the subsequent transgression partially eroded the soil that developed, enough was preserved to distinguish the unconformity. As described in a later section, paleosols are being increasingly used to identify sequence boundaries and to subdivide a particular stratigraphic succession into sequences.

Unconformities are commonly regional in scale, and they can be highly irregular surfaces along which the amount of missing time varies considerably (e.g., Wheeler, 1958). Consequently, the paleosol associated with an unconformity can show lateral changes on a regional scale, and those changes can be used to interpret lateral variations in topography and missing

Fig. 3. (A) Levee deposits with weakly developed paleosols. Although the rock shows reddening and root traces, relict bedding is apparent and indicates that sedimentation was so rapid that the parent material did not undergo mixing or homogenization. Lens cap at top of photo is 6 cm in diameter. (B) Alternations of strongly developed cumulative paleosols (S) and weakly developed compound paleosols (W). The cumulative paleosols are distinguished by gray A horizons overlying thick red B horizons; they formed on floodbasin deposits. The weak paleosols have pale colors and formed on what Kraus (1996) interpreted to be avulsion deposits. Ridge containing the paleosols is 18 m high. (C) Two vertically stacked compound paleosols formed on what are interpreted as distal crevasse splay deposits. The paleosols show reddening but are separated by rock that shows little pedogenic modification and is represented by white bands. The upper compound paleosol (orange and underlying white band) is 47 cm thick and the lower is 50 cm thick. (D) Floodplain incision. Well-developed red paleosols have been scoured and the scour filled with more weakly developed grayish paleosols. The scour cuts approximately 10 m down into the red paleosols. All examples from the lower Eocene Willwood Formation, Bighorn Basin, Wyoming.

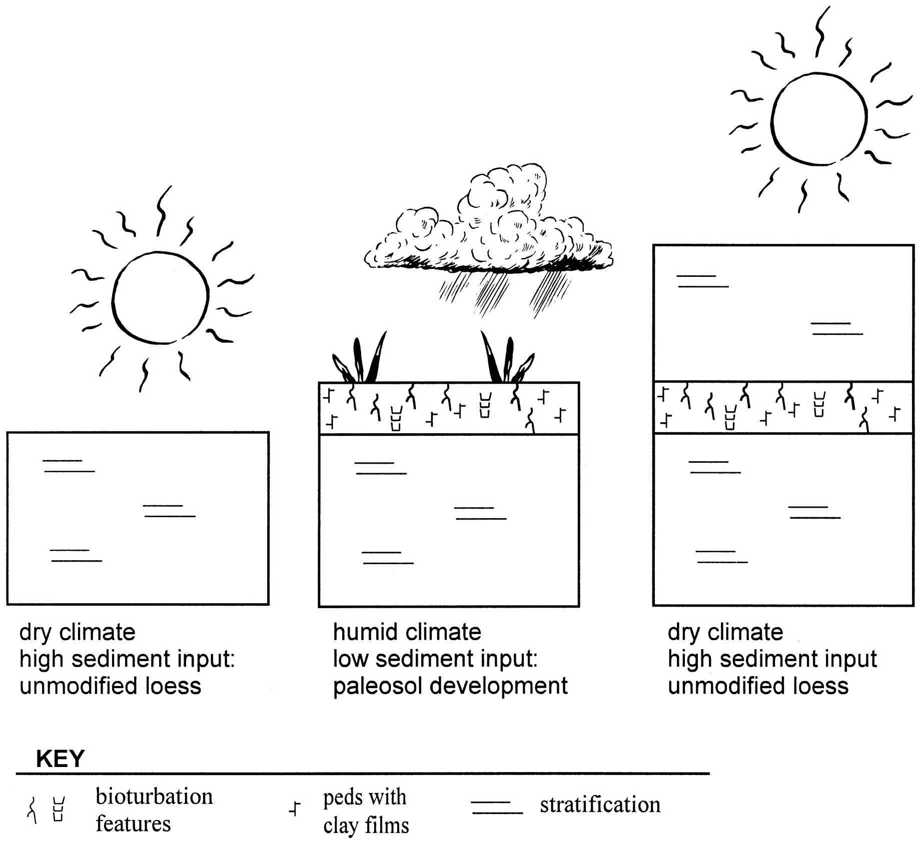


Fig. 4. Schematic diagram showing succession of steps that produce alternations of unmodified loess and paleosols in thick sequences of loessite. Formation of the paleosols has been attributed to reduced sediment input related to more humid climates. During drier periods, sediment input is sufficient to inhibit pedogenic development.

time along the ancient landscape. A good example, described by Joeckel (1995), is a Pennsylvanian paleosol that changes laterally from a thick and well-developed profile to a thinner profile with weaker pedogenic development. Joeckel used the paleosol changes to understand topographic variations and differences in exposure time across the Pennsylvanian landscape.

3.3. Stratigraphic analysis

Recognizing that many individual paleosols are morphologically distinctive and are really extensive, Quaternary geologists have long used them as stratigraphic markers for subdividing Quaternary deposits and for local and regional correlations (e.g., Richmond, 1962; Morrison, 1967; Pecsi, 1995). Although geologists working in Tertiary and older rocks also have a relatively long record of using paleosols to solve stratigraphic relationships, prior to about 1980, the focus was on the stratigraphic significance of strongly developed paleosols that marked unconformities (e.g., Ritzma, 1955; Schultz et al., 1955; Abbott et al., 1976). More recent studies demonstrate the utility of paleosols for subdividing thick continental successions into genetic sequences and solving local and even global correlation problems. They also show that pedogenic carbonates are a potentially important new means of absolute dating continental rocks and that floodplain paleosols can be a useful component of sequence stratigraphic models and field studies.

The Bozeman Group is a thick continental succession in MT, USA, which spans the Cenozoic. Hanneman and Wideman (1991) and Hanneman et al. (1994) showed that problems with the stratigraphic relations of this group could be clarified using calcretes to delimit basin-wide unconformities and thus to subdivide the succession into five sequences. Of particular interest is that Hanneman et al. (1994) integrated surface geologic, seismic, and well log data and showed that the calcretes produced bright seismic reflections that could be traced in the subsurface. This study not only highlights the potential of paleosols for solving stratigraphic problems, but also shows that, in some cases, paleosols can be recognized in the subsurface using data other than core data and the paleosols can then be used for regional correlations. Similarly, Ye (1995) used paleosols to correlate Miocene strata in the subsurface and to constrain the position of a petroleum reservoir. The paleosols were identified in cores primarily on the basis of color, the presence of roots and calcrete nodules, and mineralogy. Although Ye was also able to use log characteristics to distinguish paleosols from rock that was not pedogenically modified, in contrast to Hanneman et al., he found it very difficult to identify, and thus correlate, individual paleosols from log characteristics alone.

Going beyond local or regional correlation using paleosols, a study by Koch et al. (1992) showed that some paleosols have the potential to make global correlations. A sharp decrease in the $^{13}C/^{12}C$ ratios of biogenic marine carbonates has been well documented at approximately the Paleocene/Eocene boundary and has been attributed to global marine warming (e.g., Shackleton, 1986; Zachos et al., 1993). Koch et al. were able to use pedogenic carbonates to identify this sharp isotopic change in Paleogene alluvial rocks in the Bighorn Basin of Wyoming and, thus, to correlate the continental record to the marine isotopic record. Through this correlation, they showed that the P/E boundary in the marine record corresponds to a particular biostratigraphic boundary in the Bighorn Basin and were then able to relate changes in mammalian faunas at this boundary to global climate change.

As anyone working in the stratigraphic record is aware, direct absolute dating of the sedimentary record is difficult. Yet, recent work indicates that U/Pb dating of pedogenic carbonates offers a new approach to obtaining absolute dates for continental rocks (Rasbury et al., 1998). Those authors estimated the U–Pb ages of pedogenic carbonates from Pennsylvanian–Permian strata. Because they argued that those strata were deposited rapidly and show no evidence of later diagenetic modification, they concluded that the time of formation of the calcretes is a proxy for the time of deposition of the strata that enclose them. This technique is potentially powerful because pedogenic carbonates are found in numerous stratigraphic successions throughout the Phanerozoic.

Finally, various authors have emphasized the sequence stratigraphic importance of paleosols, both from a modeling approach (e.g., Wright and Marriott, 1993) and, more commonly, through field studies (e.g., Aitken and Flint, 1996; McCarthy and Plint, 1998). Wright and Marriott (1993) developed a model that predicts how the degree of pedogenic maturity and soil drainage of coastal plain paleosols varies at different times in the sea level cycle. Their model suggests that, during sea-level lowstands, mature and well-drained soils should form on terraces that are produced by channel incision because the water table drops and the floodplains receive no sediment (Fig. 5). When sea level first begins to rise, hydromorphic soils develop because baselevel is rising. As sea level continues to rise, accommodation space is created and floodplain sedimentation is rapid, leading to weakly developed paleosols. Later, during early highstand, the rate of aggradation begins to decrease and more strongly developed soils should form. Finally, when highstand is fully achieved, accommodation space is low. Mature soils form; however, their preservation potential is probably low because aggradation rates are so low that sediment reworking by channel combing is intense.

A key feature of the Wright and Marriott model is that, contrary to other models of sequence stratigraphy, it predicts that the highest rates of aggradation are characteristic of the transgressive systems tract (TST), and that, because of low accommodation space, the highstand systems tract (HST) had low rates of aggradation. Consequently, abundant overbank deposits with weakly developed paleosols should characterize the TST, whereas lower amounts of overbank deposits with well-developed paleosols are typical of the HST. In contrast some field studies

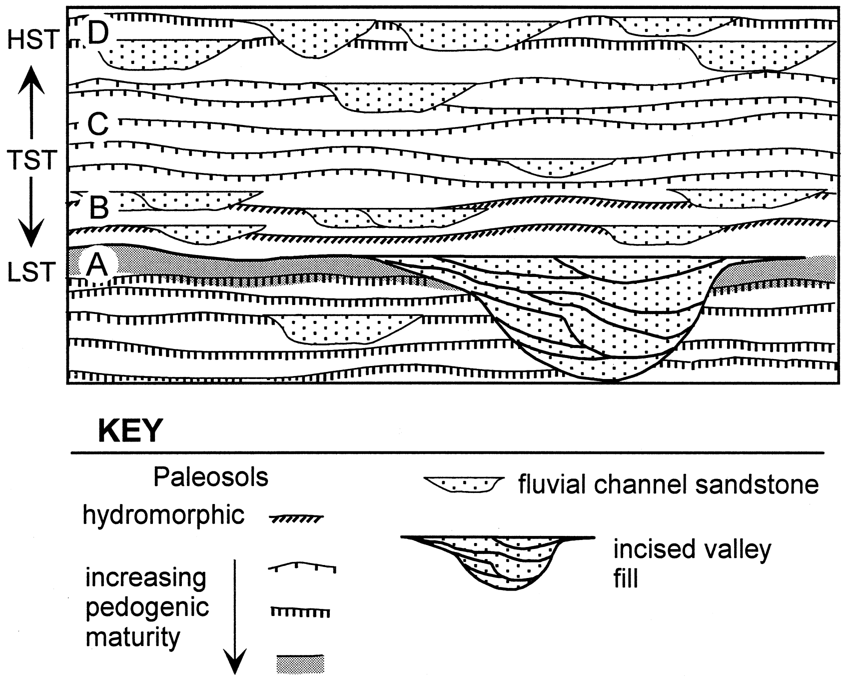


Fig. 5. Schematic diagram showing Wright and Marriott (1993) model of pedogenic development related to the sea-level cycle. (A) The lowstand systems tract (LST) is characterized by channel incision and strongly developed and well-drained paleosols that form on terraces. (B) Early in the TST, hydromorphic paleosols may form because base level is rising; channel sandstones may overlap because accommodation space is still low. (C) Increased accommodation space in the TST produces rapid sediment accumulation and weakly developed soils. (D) The HST is characterized by low accommodation space, thus, well-developed paleosols (modified from Wright and Marriott, 1993).

(e.g., figure supplied by M. Uliana and L. Legarreta, cited in Shanley and McCabe, 1994) have shown that the HST, rather than the TST, is dominated by overbank deposits and that paleosols are most common in the HST. Shanley and McCabe (1994, p. 562), suggested that this discrepancy may have arisen because the Wright and Marriott model ''takes no account of the basinal position of systems tracts nor the importance of recognition of the landward equivalent of marine condensed intervals''. Nonetheless, the model shows that paleosols are an important area of future research in sequence stratigraphy.

Sequence stratigraphic field studies generally stress the importance of paleosols for identifying interfluvial sequence boundaries and for distinguishing incised valley fills from major channel sandstones. For example, Aitken and Flint (1996) found paleosols to be critical for identifying interfluvial sequence boundaries in Pennsylvanian rocks, because, beyond the incised valley fills, the boundaries are subtle features. The paleosols, which developed on delta plain deposits, are typically hydromorphic or gleyed paleosols; however, the presence of composite paleosols indicates that many of the soils were initially well drained when water tables were low, and then they underwent a secondary gleyed stage when the water table rose in response to transgression. Despite their potential value in identifying sequence boundaries, paleosols like these probably have low preservation potential because the ensuing transgression can obliterate them (Aitken and Flint, 1996). Interfluvial sequence boundaries with associated paleosols have also been described from entirely fluvial successions (e.g., Leckie et al., 1989). The presence of numerous, closely spaced paleosols in the Cretaceous strata suggested to Leckie et al. that sediment supply to the floodplain was cut off and they attributed this to channel incision caused by lowered base level.

Despite the progress using paleosols to recognize interfluvial sequence boundaries, McCarthy and Plint (1998) noted that more attention should be paid to details of the paleosols. In particular, those authors showed how examining the micromorphologic attributes of the paleosols in conjunction with field morphology and stratigraphic relationships leads to a more complete understanding of the sequence of events that produced paleosols at a sequence boundary. In a study of Cretaceous paleosols that formed on coastal plain deposits, they found that grey soil colors and the presence of siderite and preserved organic matter indicated that the soils first formed under poorly drained conditions (Fig. 6A). The presence of clay coatings and various ferruginous features suggest that drainage then improved, and papules (fragmented clay coatings) and pedorelicts indicate that the interfluves underwent an episode of erosion (Fig. 6B). They attributed these features to lowered base level, which caused the channels to incise and water tables to drop. Silty layers directly above the sequence boundary within the paleosols reflect renewed sedimentation on the interfluves because base level rose again. Finally, wet soil indicators, present above the sequence boundary, show that water tables also rose, although multiple organic-rich beds indicate that water tables fluctuated somewhat (Fig. 6C).

Although sea level fluctuations are an important control on coastal plain rivers and the development and morphology of paleosols associated with those rivers, the current opinion is that sea-level effects probably do not extend inland more than about 100–150 km from the shoreline (e.g., Shanley and McCabe, 1994; North, 1996). In alluvial systems more distant from the sea or in closed basins, climate and regional tectonic activity are the major controls on

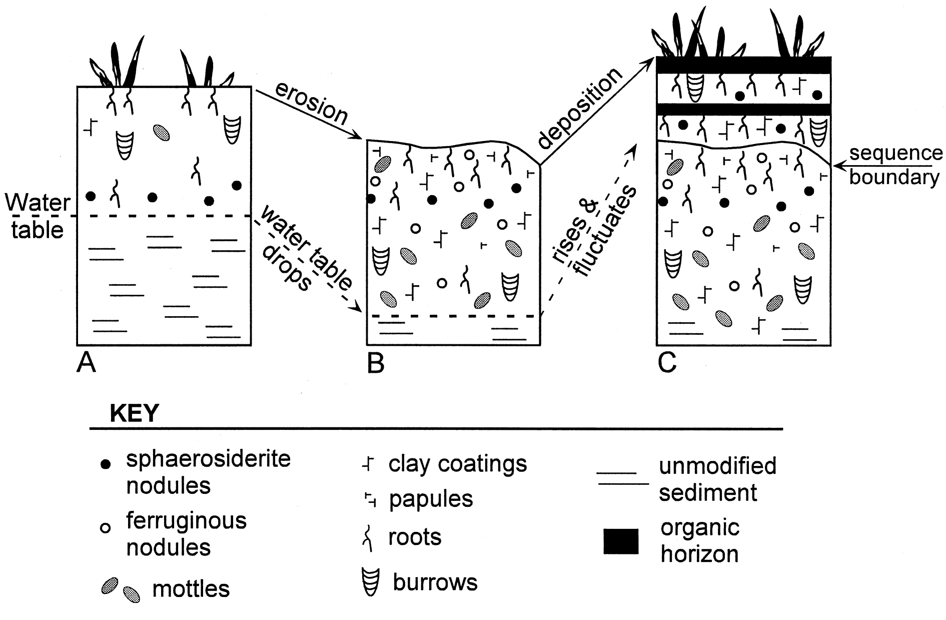


Fig. 6. Schematic diagram showing sequential development of an interfluve paleosol that marks a sequence boundary. (A) Siderite and grey soil colors show that initial soil drainage was poor. (B) Base level dropped, the floodplain was incised, and the water table dropped. Clay coatings and ferruginous nodules formed in the better-drained soil. (C) Base level rose and new sediment was deposited. The water table also rose, producing hydromorphic features including multiple organic horizons. See text for more details (modified from McCarthy and Plint, 1998).

rivers (e.g., Blum, 1994; Blum and Valastro, 1994), and paleosol development reflects those effects rather than sealevel fluctuations.

3.4. Sediment accumulation rates

Sediment accumulation rates are generally calculated by dividing the thickness of a particular stratigraphic section by its known or estimated time span. Sediment accumulation rates estimated for continental stratigraphic successions are usually severely time-averaged because radiometric dates are not available or widely spaced in time and because the temporal resolution of paleomagnetic dating or continental biostratigraphy is generally relatively coarse (e.g., Kraus and Bown, 1993). Because the developmental history of a sedimentary paleosol reflects the rate of pedogenesis relative to the total time of soil development and how steady or unsteady that deposition was (Fig. 2), the kind of paleosol that develops is a good indicator of sediment accumulation rates for thin stratigraphic intervals. The relative degree of pedogenic development is used as a proxy for the relative rate of sediment accumulation, and, as discussed in a later section, paleosols have been used to compare and contrast sediment accumulation rates, in a qualitative sense, in different parts of a depositional basin (e.g., Atkinson, 1986; Platt and Keller, 1992) and through time (e.g., Kraus and Aslan, 1993; Kraus, 1997; Soreghan et al., 1997; examples in Kraus and Aslan, 1999).

In addition to this qualitative approach, several workers have attempted to use paleosols to quantitatively estimate short-term sediment accumulation rates, using somewhat different approaches. In a seminal paper on this topic, Leeder (1975) compiled ages of Quaternary calcretes to establish ages of formation for particular stages of calcrete development. The ages were then incorporated into a model for estimating floodplain accretion rates. Retallack (1983, 1984) also took a quantitative approach to accumulation rates by estimating development times for different kinds of paleosols through analogy to modern soils.

Bown and Kraus (1993) and Kraus and Bown (1993) took a different approach. On the basis of morphologic criteria including the combined thickness of the A and B horizons, geochemical trends within a paleosol profile, and degree of clay translocation, those workers assigned floodplain paleosols in the Paleogene Willwood Formation to seven stages of pedogenic development. Bown and Kraus (1993) constructed a composite stratigraphic section extending from the base to the 600-m level of the formation from numerous measured sections and then subdivided the composite section into 25-m-thick intervals. The paleosols found within a particular 25-m interval were assigned to a stage of development and then averaged to yield a relative degree of pedogenic maturity for each interval (Fig. 7A). The relative amount of time represented by each interval was then estimated from the relative degree of develop-

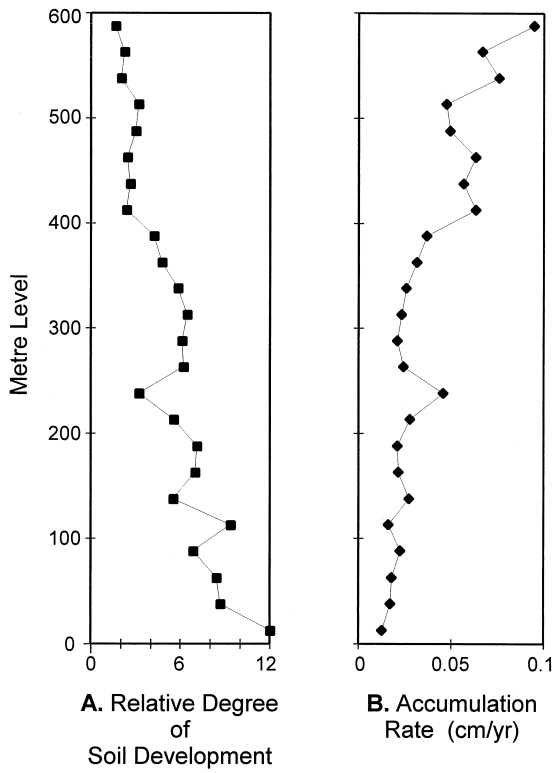


Fig. 7. (A) Plot of relative degree of pedogenesis vs. meter level in composite section of the lower Tertiary Willwood Formation. Relative degree of pedogenic development was calculated for 25-m-thick intervals by assigning paleosols in each interval to a stage of pedogenic development outlined by Bown and Kraus (1993). (B) That information was combined with the absolute age of the 600-m-thick section to estimate sediment accumulation rates for each 25-m-thick interval. See text for more details (modified from Bown and Kraus, 1993; Kraus and Bown, 1993).

ment for each interval. For example, the lowest 25-m interval has the highest relative degree of maturity at 12.0 and occupied 9% of the total time represented by the total section (Fig. 7A). Because absolute ages had been established for the lower and upper boundaries of the 600-m-thick section, the time span for each 25-m interval could be calculated. From that calculation, the sediment accumulation rate for each 25-m-thick interval was estimated (Fig. 7B). Kraus and Bown (1993) then used these short-term accumulation rates to examine up-section changes in sedimentology and paleontology.

Using paleosols to estimate sediment accumulation rates in a quantitative fashion suffers from several limitations. The Leeder (1975) and Retallack (1983, 1984) approach assumes that times of formation of modern soil analogs can be reliably determined. The pitfalls of this assumption for calcretes have been thoroughly discussed by Wright (1990). He concluded that this approach is not justified because Quaternary calcretes actually show considerable variability in their rates of formation depending on local rainfall and the availability of airborne carbonate dust.

A second potential problem is erosion or slow sedimentation rates, which can generate composite paleosols. With erosion, accumulation rates can be underestimated. If the paleosols are composite, as they are in numerous stratigraphic successions, accumulation rates can be overestimated. Finally, Retallack (1998) has pointed out that studies like those of Bown and Kraus (1993) and Kraus and Bown (1993) assume that there are no major discontinuities in the paleosol section and that all of the time contained in a vertical section is represented by paleosols. He noted that this is probably an unrealistic assumption for stratigraphic sections that span several millions of years.

4. Paleolandscape reconstruction

In the Quaternary record, soil–landscape studies are important for interpreting landscape evolution (e.g., McFadden and Knuepfer, 1990). Similarly, in pre-Quaternary strata, paleosol–landscape studies can help in reconstructing ancient landscapes (e.g., Bown and Kraus, 1987; Besly and Fielding, 1989; Platt and Keller, 1992). Paleosol/landscape studies focus on the spatial distribution of different kinds of paleosols and the different landscape elements and processes that produced those paleosols. The study of Joeckel (1995), described above, is a good example of using lateral changes in a paleosol associated with an unconformity to analyze regional topographic variations along a specific degradational landscape. Furthermore, thick sedimentary sequences with multiple paleosols provide a record of land surfaces over time so that landscape changes and evolution can be examined.

Paleosol–landscape associations can be studied at different spatial scales that range from local (e.g., a channel and associated floodplain, an eolian dune and associated interdune) to basin-wide in scope (e.g., a drainage system, a sand sea). At the local scale, lateral changes in paleosol properties are largely the result of local variations in grain size and topography. At the scale of the sedimentary basin, paleosols in different locations differ because of basinal variations in topography, grain size, climate, and subsidence rate. The following discussion provides examples of how paleosols have been used to interpret landscapes of the past, starting at the local scale and building to the basin-wide scale. For a more thorough discussion of paleosol–landscape associations at different spatial and temporal scales, see Kraus and Aslan (1999).

4.1. Local landscape reconstruction

Paleosol–landscape associations at this scale have been studied primarily in alluvial rocks (e.g., Bown and Kraus, 1987; Kraus, 1987, 1997; Besly and Fielding, 1989; Platt and Keller, 1992). Catenas and pedofacies are common paleosol–landscape associations that develop at the scale of the channel and associated floodplain. The two are not mutually exclusive, and floodplain paleosols can show a combination of the two.

In a catena, better-drained soils form on channel-marginal deposits (levees, crevasse splays) because the deposits are elevated relative to the surrounding floodbasin and usually consist of relatively permeable fine sands and silts (Fig. 8). The soils may show evidence of oxidized conditions, including brown A and Bw (weathered or structural B) horizons and

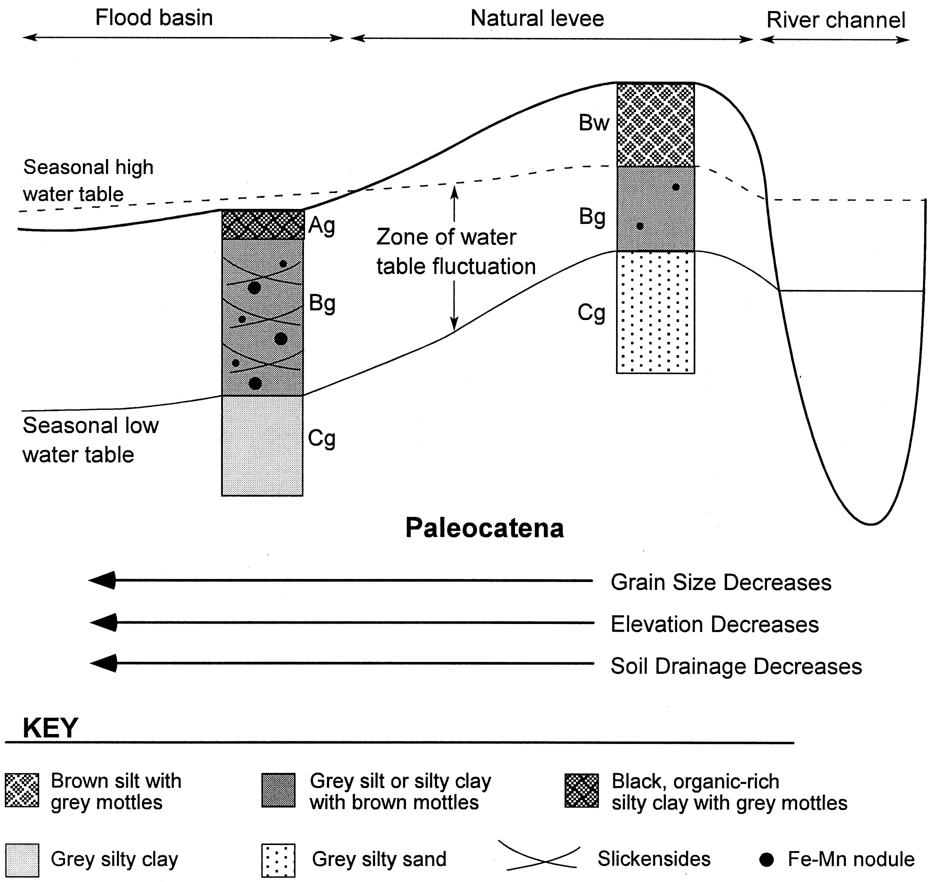


Fig. 8. Schematic diagram showing a paleocatena, which developed because of changes in grain size and topography away from the ancient channel. A moderately or well-drained soil generally forms on the levee, although subsurface horizons can be gleyed (Bg horizon) because of proximity to the ground water table. Poorly to very poorly drained soils are more typical of the floodbasin (modified from Kraus and Aslan, 1999).

sparse organic matter. Grey soil colors are more abundant lower in the profile (Bg and Cg horizons, which are gleyed B and C horizons) because it is closer to the ground water table and more prone to reduced conditions (e.g., Duchaufour, 1982). Away from the channel, the soils become progressively more poorly drained because the topographic position is lower and the sediment is finer-grained and less permeable. Waterlogged, reduced conditions favor the accumulation and preservation of organic matter in an Ag (gleyed A) horizon and the development of a thick, grey Bg horizon with many redoximorphic features. Fluctuations of the water table also produce intersecting slickensides in the clayey sediments.

Ancient catenas have been described from Paleogene fluvial rocks (e.g., Fastovsky and McSweeney, 1987) and from Jurassic deltaic plain deposits (e.g., Arndorff, 1993). Those catenary relationships have been used to reconstruct the ancient floodplain landscape. Arndorff, for example, found that the distribution of different paleosols reflected their position on the deltaic plain. Relatively well drained, light brown paleosols with a yellowish to orange subsurface horizon formed on sandy natural levees and crevasse splays. In contrast, dark-colored silty claystones interpreted as gleyed alluvial soils formed in the backswamps.

Pedofacies focus on lateral changes in the degree of pedogenic development with increased distance from the ancient channel (Bown and Kraus, 1987). Bown and Kraus observed weakly developed paleosols on channel-marginal deposits and increasingly well developed paleosols away from the channel sandstone (Fig. 9). These changes were primarily attributed to sediment accumulation rates, which tend to decrease away from an active channel (e.g., Guccione, 1993). Pedofacies have also been inferred from cores on the basis of the degree of paleosol development and the proportion of sandstone present (Platt and Keller, 1992) (Fig. 10). Weakly developed paleosols (stage 1 of Platt and Keller) were found in cores dominated by meander belt sandstones (25% of total core thickness) and crevasse-splay sandstones (30% of core thickness). Cores where sandstones of any kind were less common were interpreted to be distal to major channels. These were dominated by more strongly developed paleosols (stages 2–3), as predicted by the pedofacies model.

Although pedofacies relationships have been recognized in these and other ancient alluvial sequences (e.g., Wright and Robinson, 1988; Smith, 1990), the pedofacies model, as it is currently understood, does not satisfactorily explain lateral relationships in all floodplain paleosol successions (e.g., Wright, 1992b). Kraus (1997) and Kraus and Aslan (1999) have discussed limitations of the pedofacies model. In addition, North (1996) pointed out that the pedofacies model may be appropriate only for thick, aggradational successions that are dominated by composite and cumulative paleosol profiles.

Paleosols in other sedimentary environments also show lateral changes that reflect different locations in the depositional landscape. Although studies of pre-Quaternary paleosols associated with eolian environments are sparse, Quaternary deposits provide numerous examples. Loess deposits become thinner and finer-grained away from source areas, and these

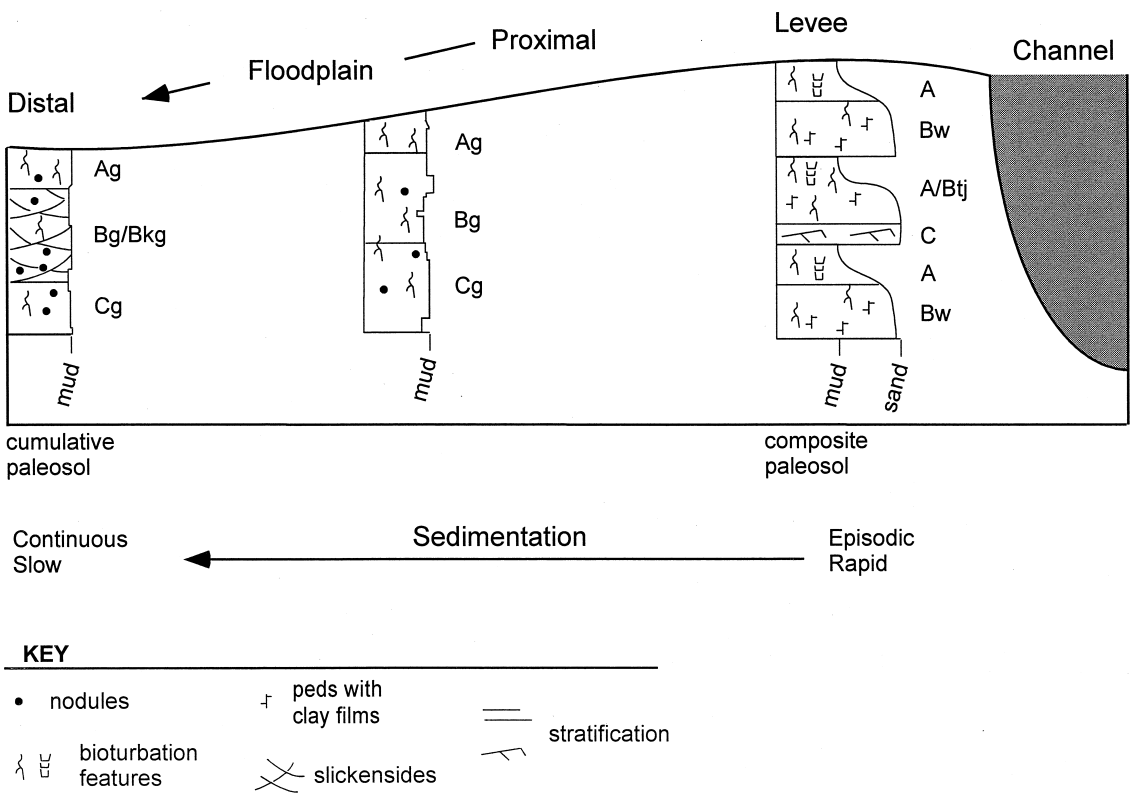


Fig. 9. Schematic diagram showing pedofacies relationships. Compound paleosols tend to form adjacent to the channel because sedimentation was rapid and episodic. Farther from the active channel, more strongly developed, cumulative paleosols formed because sedimentation was slow and steady (modified from Bown and Kraus, 1987).

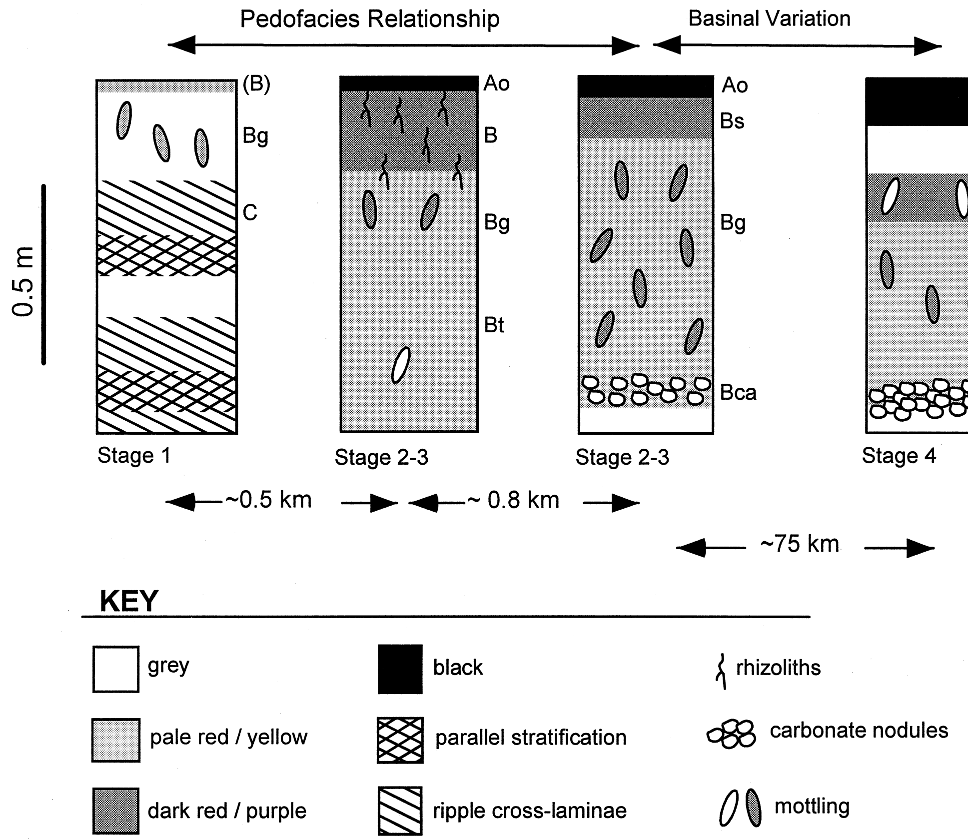


Fig. 10. Paleosol–landscape associations in alluvial rocks at both the local and basinal scales. Pedofacies developed over a distance of <2 km, with weakly developed paleosols (Entisols or Inceptisols) grading into better-developed paleosols (Alfisols and Vertisols). Over a distance of ~ 75 km, paleosols became even more strongly developed (Stage 4) in response to basinal changes in the rate of aggradation. See text for more details (modified from Platt and Keller, 1992).

lateral changes in parent material produce lateral changes in the cumulative soils that form on the loess (e.g., Smith, 1942; Ruhe, 1983). For example McDonald and Busacca (1990) found that, with increasing proximity to the sediment source, a single soil bifurcated into two well-developed soils. Holliday (1990) also documented lateral variations in soils developed on eolian deposits and attributed these primarily to variations in the grain size of the sediment, although local thickness variations also played a role.

Catenas and pedofacies are produced by variations in topography and grain size related to landscape position. Modern soil studies show that compositional differences are linked to grain size differences and suggest that those compositional differences also influence soil development. In fluvial systems, for example, sands and coarse silts that accumulate in channel-marginal environments (levees, splays) tend to be dominated by quartz, feldspar, and lithic fragments (e.g., Schumacher et al., 1988). In contrast, clay and fine silt, which typically accumulate in distal floodbasins, consist primarily of clay minerals such as smectite, illite, kaolinite, and chlorite. Aslan and Autin (1998) concluded that, in Mississippi River floodplains, these depositionally controlled compositional differences had a greater impact on the chemistry of alluvial soils than did weathering processes. Similarly, with Quaternary loess deposits, compositional changes, including an increase in clay minerals, are found with increasing distance from source areas, (e.g., Smith, 1942; Kleiss and Fehrenbacher, 1973). These studies indicate that compositional variations associated with different parts of the depositional landscape should also be considered when analyzing and interpreting paleosols.

4.2. Basinal landscape reconstruction

At the basin scale, analyzing paleosol/landscape analysis can provide information about global or regional climate change, sea level fluctuations, and regional tectonics. At a truly basinal scale, Alonso Zarza et al. (1992) described Miocene paleosols that developed in a spectrum of depositional settings, including alluvial fans, lake margins, and river floodplains. Stratigraphic intervals approximately 100 m thick were examined in two contrasting areas of the Madrid Basin. Paleosols in the two areas differ, and those differences were related to local climatic conditions and sediment accumulation rates. Consequently, the authors could use the paleosols to interpret basinal variations in climate and basin subsidence, which controlled sediment accumulation rates.

In fluvial basins, changes in the kinds of paleosols have been documented in the direction of regional paleoslope. For example, in the Eocene Capella Formation, Atkinson (1986) observed more mature and better-drained paleosols proximal to the source area. Over a distance of nearly 30 km down paleoslope, progressively less mature and more poorly drained paleosols were found. Atkinson attributed these changes to a decline in topographic relief and increased accumulation rates away from the source. Over an even greater downslope distance ($\sim$ 75 km), Platt and Keller (1992) also documented an increase in the maturity and hydromorphy of floodplain paleosols (Fig. 10). Consistent with the increased maturity, the stratigraphic interval thins downslope, indicating slower sediment accumulation in more distal parts of the alluvial basin.

McCarthy et al. (1997) recognized floodplain paleosols of varying degrees of pedogenic development and related the paleosols to landscape position on a basinal scale. In this Lower Cretaceous example, the paleosols formed in a foreland basin adjacent to the epicontinental seaway that developed in North America during that time. Not surprisingly, the authors asserted that pedogenic development should be directly dependent on floodplain stability and that the tectonic hinge zone is the most stable area of the landscape. Consequently, they argued that ancient hinge zones can be located by examining regional patterns of paleosol development.

Finally, a study of early Eocene fluvial rocks shows that basinal variations in grain size, related to the basin geography, can also produce different kinds of paleosols (Kraus and Gwinn, 1997). Paleosols in the northern part of the Bighorn Basin are better drained than paleosols 75 km to the south, in the central part of the basin. The differences were attributed to grain size differences, and those, in turn, were linked to basin position, in particular, distance from a local sediment source. Because the northern area is directly adjacent to a local sediment source, the sediment is coarser and more permeable, producing better-drained paleosols. The southern area is more distal to any of the sediment sources around the basin, resulting in clay-rich sediment that produced waterlogged soils.

Despite their potential for helping interpret the history of alluvial basins, studies of paleosol variability at the basin scale are few, in part, because they are time and labor intensive. Kraus (1992) demonstrated that these difficulties can be overcome by including remote sensing analysis of satellite or airborne spectral data in a field and laboratory study of paleosols. Remote sensing is potentially valuable because it can provide data not easily obtained on the ground and gathers some data more efficiently than field study. Yet, remote sensing is rarely used to solve basic stratigraphic or sedimentologic problems. Because remote sensing imagery is becoming more readily available and because data analysis now can be done on personal computers, this tool should be more widely used in future paleosol studies, where exposures are good.

5. Interpreting paleoclimates

Paleosols are used both to interpret the paleoclimatic regime and to quantitatively estimate ancient mean annual precipitation (MAP) and mean annual temperature (MAT). Ancient climatic conditions can be interpreted by classifying the paleosols and using modern analogs to infer the paleoclimatic regime or by identifying particular pedogenic properties that modern studies show to have climatic significance. Stable carbon and oxygen isotopes are also used to interpret ancient climate and some effort has been made to use isotopic composition to estimate MAT (e.g., Mack et al., 1991; Koch et al., 1995). Finally, the depth at which calcic horizons originally formed has been used to estimate paleo-precipitation. The following sections provide overviews of these approaches as well as their limitations.

5.1. Modern soil analogs

The paleosol literature provides numerous examples of the classification approach to paleoclimatic interpretation. For example, Mack (1992) used alluvial paleosols to interpret a climate change across the Lower Cretaceous–Upper Cretaceous boundary in New Mexico. Older red, calcic paleosols, interpreted as Aridisols, were attributed to semi-arid or arid conditions. These are overlain by Inceptisols and Alfisols that have somber colors and lack carbonate, suggesting a subhumid or humid climate. Similarly, Bestland (1997) used paleosols to interpret a climate change across the Eocene–Oligocene boundary. The older paleosols were interpreted as 'Ultisol-like' paleosols that formed under humid, subtropical conditions, whereas the younger smectitic paleosols were judged to be 'Alfisol-like' and indicative of humid, temperate conditions.

In a more unifying approach to paleoclimatic interpretation, Mack and James (1994) generated a paleosol analog to the Soil Map of the World (e.g., FAO, 1974), which links modern soils to particular climatic zones. Using the Mack et al. (1993) paleosol classification (Table 1), they suggested that highly weathered paleosols, such as Oxisols and Argillisols, should be characteristic of wet equatorial paleoclimates in which MAP and MAT are high and have little seasonal variation. Argillisols, Spodosols, and Gleysols are more apt to form in humid (MAP $>$ 1000 mm) midlatitude climates, whereas Calcisols are indicative of a dry (MAP $<$ 1000 mm) subtropical paleoclimatic zone. One caveat to this global approach is that it is not appropriate for paleosols that formed prior to the advent of vascular plants.

The second approach to climate reconstruction—focusing on a particular soil property or group of properties—is based on Quaternary soil studies, which have shown that particular features, including clay and carbonate accumulations and the depth of soil oxidation can be quantitatively related to soil-forming factors, including climate (e.g., Bockheim, 1980; Harden and Taylor, 1983; Birkeland, 1984). Molecular weathering ratios (e.g., silica/sesquioxides or bases/alumina) provide a good illustration. In some cases, the ratios indicate intense leaching of base cations and the loss of silica, which are characteristics of modern soils that formed in humid, tropical climates (e.g., Arndorff, 1993; Retallack and German-Heins, 1994; Gill and Yemane, 1996). Clay mineralogy has also been used to detect and interpret a climatic change. Robert and Kennett (1994), for example, found that clay minerals on the Antarctic continent showed a dramatic increase in smectite (and corresponding decrease in illite) during the latest Paleocene. The clay change corresponds to a well-documented isotopic shift that marks the latest Paleocene thermal maximum (e.g., Zachos et al., 1993). Robert and Kennett concluded that the increase in smectite was the result of increased chemical weathering as a result of warmer temperatures and greater rainfall at that time.

In some cases, the paleosols are parts of cyclothems in which alternations of particular kinds of paleosols are linked to climatic cycles. Good examples are provided by the wet/dry cycles linked to pedogenically-unmodified sediment and paleosols in loess and loessite. In some, the cyclic climatic changes are linked to either sea-level changes or tectonic activity. For example, working in Lower Permian cyclothems of the U.S. mid-continent, Miller et al. (1996) described cyclic changes from vertic paleosols to Alfisols with calcic horizons and attributed these to fluctuations between semi-arid/subhumid conditions and humid monsoonal conditions. In a subsequent study (McCahon and Miller, 1997), these paleosol/climate cycles were linked to glacially controlled sea-level changes. The change to increased precipitation followed by a rapid change back to arid conditions was attributed to gradual regression caused by glacial activity followed by a

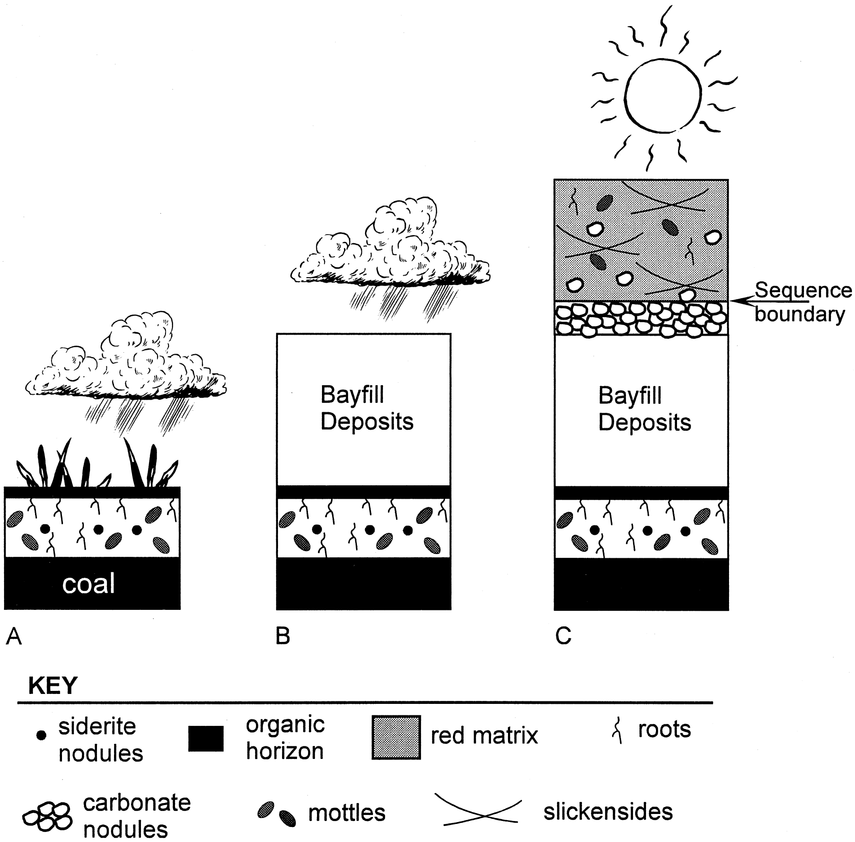


Fig. 11. Schematic diagram showing development of paleosol cycles described by Tandon and Gibling (1994). (A) Hydromorphic paleosols formed during marine highstand when climates were wet. (B) Climates remained humid as bayfill deposits accumulated. (C) Calcareous paleosols formed because of change to dry conditions during lowstand and early transgression.

rapid transgression. Tandon and Gibling (1994) described Carboniferous cyclothems that are 20 to 30 m thick and that show an alternation between hydromorphic paleosols, including Histosols, and calcareous paleosols. They invoked a climatic change from humid to strongly seasonal conditions to explain the paleosol cycles, and they too concluded that the climatic cycles were linked to sea-level fluctuations (Fig. 11). Relatively humid climates accompanied a marine highstand, and conditions became drier during the lowstand and early transgression.

5.2. Depth to calcic horizon

A particular soil property that has been used to estimate MAP is depth to a calcic horizon in soils containing a zone of calcareous nodules. This technique has been developed by Retallack (1994) who expanded on earlier efforts by Jenny (1941) to link MAP (mean annual temperature) to depth to the calcic horizon. Jenny produced a scatter plot of data obtained from soils on the Great Plains of the US, to which Retallack added data from calcareous soils found in different soil-forming environments all over the world. The empirical relationship he determined from the modern soils was then applied to calcareous paleosols of Eocene and Oligocene age to estimate paleo-precipitation. The advantages to this approach are obvious. Many paleosols contain calcareous nodules, and this approach provides an expedient means of estimating paleo-precipitation.

Nonetheless, as Retallack (1994) indicated, this approach has several potential problems. First, depth to the calcic horizon varies with atmospheric CO_2

(Cerling, 1991, 1992). Cerling (1991) and others (e.g., Driese and Mora, 1993) have shown that levels of CO_2 in the atmosphere have varied over time and have differed significantly from the modern levels under which the empirical curve was constructed (see Section 6). A second potential problem is that MAP can be underestimated if the upper part of a paleosol has been truncated. Retallack (1994) has suggested ways to avoid these two problems, and the reader is referred to his discussion.

A more contentious problem is the effects of sediment compaction, especially if the rocks have undergone significant burial. Retallack (1994) has tended to calculate depth to the carbonate horizon after decompacting the strata using the compaction curves of Baldwin and Butler (1985). Yet, several studies suggest that these curves are not appropriate for paleosols and, in fact, that many paleosols are little affected by compaction. In an innovative study, Caudill et al. (1996) used vertical micro-cracks in Paleozoic Vertisols to show that the paleosols underwent no more than 10% compaction despite the depth of burial. Caudill et al. concluded that, because Vertisols have initial bulk densities that are quite high, they undergo little compaction with burial. In their case, they calculated 650 mm of precipitation on the basis of depth to the carbonate horizon in the Vertisols, which is comparable to the precipitation under which modern Vertisols form. Evidence from both modern and ancient floodplain deposits also indicates that, because paleosols require surface exposure, they underwent significant dewatering and compaction prior to burial (Nadon and Issler, 1997). These studies suggest that depth to the calcic horizon can be used directly to estimate MAP as long as problems with truncation and paleoatmospheric CO_2 can be reconciled.

5.3. Isotopic analysis

For a thorough discussion of the factors controlling the isotopic composition of soil minerals and the problems encountered in interpreting isotopic compositions, the reader is referred to Cerling and Quade (1993). In brief, the $\delta^{18}O$ value of soil carbonate depends on the isotopic composition of the meteoric waters from which it precipitated. The isotopic composition of the meteoric waters is, in turn, influenced by the mean annual temperature (MAT) (Cerling, 1984; Cerling and Quade, 1993). Furthermore, the carbon isotopic composition of modern soil carbonates tends to be higher in warmer areas because the kind of vegetation (C_3 vs. C_4 plants) is temperature dependent. C_3 plants, which include most trees, shrubs, and cool-season grasses, produce soil carbonate $\delta^{13}C$ values around -27% whereas C_4 plants, which include most of the summer grasses and sedges, produce a value around -12%. On the basis of studies that show a major change from C_3 to C_4 vegetation between 8 and 7 Ma (Cerling, 1992; Cerling et al., 1993; Quade et al., 1994, 1995; Latorre et al., 1997), most paleoclimatic interpretations of pre-Miocene paleosols assume that the vegetation consisted entirely of C_3 plants. Because of these relationships between isotopic composition and temperature, paleosol carbonates have been analyzed to reconstruct paleoclimates (e.g., Mack et al., 1991; Koch et al., 1995). Other soil minerals, particularly clay minerals, are also potentially useful (e.g., Stern et al., 1997).

Reconstructing paleoclimates using the isotopic composition of paleosol minerals suffers from several problems. First is climatic overprinting, which can occur if sediment accumulates very slowly and the pedogenic carbonate precipitates under more than one climatic regime (Cerling, 1984). In those circumstances, the carbonate will contain an isotopic mix. Second, the oxygen isotopic composition of soil minerals is more prone to diagenetic modification than is the carbon isotope composition. Studies of Late Paleozoic pedogenic carbonates by Driese and Mora (1993) and Eocene carbonates by Cerling (1991) showed little if any diagenetic alteration of the carbon isotopes; however, the $\delta^{18}O$ in both examples was depleted as a result of recrystallization. To understand more thoroughly the effects of diagenesis on oxygen isotopes, Mora et al. (1998) analyzed the isotopic compositions of pedogenic calcite and illite in three vertic paleosols that underwent different burial depths and burial temperatures. Their results clearly demonstrate that the oxygen isotopes in this example were affected by burial diagenesis and are not appropriate for paleoenvironmental and paleoclimatic interpretations.

Cognizant of the limitations to oxygen isotopes, Mack et al. (1991) used the isotopic composition of

Permian pedogenic carbonates to interpret a climatic change over the 18 m.y. that the stratigraphic interval spanned. In particular, Mack et al. modified the approach of Cerling (1984), which was devised for environments dominated by C_4 plants, to examine Permian carbonates, which formed in a C_3 plant world. They found that $\delta^{18}O$ values increase up-section, from which they concluded that MAT increased from 15°C to 30°C. They also concluded that precipitation decreased as the temperature rose. $\delta^{13}C$ also increased up-section, an isotopic change that the authors attributed to a decrease in plant productivity, and a change that they suggested is consistent with increased temperatures and decreased precipitation. An important aspect of this study is that Mack et al. did not analyze the isotopes in isolation; rather they evaluated the reliability of their isotopic conclusions in light of paleomagnetic data, which provided the paleolatitudinal position of the study area, and macro- and micromorphological properties of the paleosols.

Koch et al. (1995) developed a different approach that is based on empirical equations, developed by Friedman and O'Neil (1977), that relate soil temperature to the $\delta^{18}O$ value of pedogenic carbonate and the $\delta^{18}O$ value of the soil water from which the carbonate precipitated. The Paleogene Willwood Formation was used to demonstrate this method. To estimate the $\delta^{18}O$ values of the Paleogene meteoric waters, Koch et al. measured the $\delta^{18}O$ values of apatite from fossil teeth (which are abundant in the Willwood paleosols) and aragonite in fossil freshwater bivalves, because the oxygen in biogenic minerals comes primarily from meteoric waters. Then, they combined the estimated $\delta^{18}O$ values of meteoric waters with measured $\delta^{18}O$ values of pedogenic carbonates from the same stratigraphic levels to determine ancient soil temperatures. Because Brady (1990) showed that, at depths more than about 30 cm in the soil profile, air temperatures are similar to soil temperatures, the calculated soil temperatures were used to estimate MAT. The MAT values ranged between about 0°C and 35°C, temperatures that are reasonable when compared to MAT values estimated for the Willwood Formation on the basis of plant fossils. The authors, however, acknowledged that the results were mixed and that this technique suffers from significant problems, which they discussed in detail. Because the Willwood Formation accumulated rapidly (e.g., Kraus and Bown, 1993), climatic overprinting should not have been a problem. More problematic with the Willwood paleosols is diagenetic overprinting because, as mentioned above, Cerling (1991) found the Willwood carbonates to be depleted in $\delta^{18}O$ due to recrystallization.

Because pedogenic carbonate are not found in all soils, Stern et al. (1997) explored the possibility of reconstructing paleoclimates using $\delta^{18}O$ values from clay minerals, which, like carbonates, have been shown to reflect the isotopic composition of meteoric waters (see Refs. in Stern et al., 1997). They measured the isotopic compositions of smectite and kaolinite from strata deposited between ~12 Ma and 2 Ma as part of the Siwalik succession in Pakistan. When plotted against time, $\delta^{18}O$ values for smectite showed similar trends to those obtained from pedogenic carbonates in the same strata by Quade et al. (1989). Both data sets show increases in $\delta^{18}O$ at ~7.5 Ma, and Stern et al. attributed the isotopic shift to a climatic change, probably either the result of increased evaporation caused by drier climates or the development of a rainshadow caused by uplift of the Tibetan plateau. In contrast, the kaolinite isotopes showed no significant trend upward through the section, and the authors suggested this may reflect contamination with detrital kaolinite. Nonetheless, their results indicate that the isotopic composition of pedogenic clay minerals shows considerable promise for paleoclimatic reconstruction with the same cautions that apply to pedogenic carbonate.

5.4. Limitations

Some workers have put forth caveats to the interpretation of paleoclimates from paleosols. The major caution is the impact of hydrologic conditions in the depositional area. In a study of Paleogene alluvial rocks in Portugal, Pimentel et al. (1996) found early diagenetic groundwater alteration features that mimic hydromorphic soil features such as pseudogley mottling and calcic horizons. Pimentel et al. provided criteria for distinguishing between the diagenetic and pedogenic features; however, they cautioned that the diagenetic features could be mistaken for pedogenic features and lead to incorrect paleoclimatic interpretations. The features do not have climatic significance in this case.

Although Caudill et al. (1996) concluded that calcareous Vertisols might provide an excellent tool for estimating ancient MAP, Aslan and Autin (1998) have suggested that caution should be used when estimating paleo-precipitation from the depth of calcic horizons, even in Vertisols. They observed calcic zones (Bk and Ck horizons) in Mississippi River floodplain soils, many of which are Vertisols. Applying the Retallack equation to both natural levee soils and backswamp soils, they calculated MAP between 545 and 908 mm, figures that seriously underestimate the actual MAP of 1500 mm. Aslan and Autin concluded that the position of the calcite was not controlled by the climate. Rather the depth to calcite in these soils is controlled by water-table levels and that calcite precipitates from ascending, not descending, groundwaters during the wet season. Consequently, they suggested that paleosols that formed in aggradational floodplain settings are more likely to preserve information on floodplain sedimentation and paleohydrology than on paleoclimates.

A similar concern about carbonates formed from ascending vs. descending waters was sounded by Slate et al. (1996). Those authors compared and contrasted the morphology and isotopic compositions of carbonates that formed in well-drained paleosols and hydromorphic paleosols. They emphasized that the Cerling model should be restricted to soil carbonate that formed in the vadose zone because only here are all oxidized carbon species in isotopic equilibrium. Carbonates from hydromorphic paleosols are unsuitable for isotopic analysis because ascending ground waters probably contaminated their isotopic composition. This paper is also important because it provides criteria for distinguishing between soil carbonates that formed in well-drained paleosols from those that formed in hydromorphic paleosols. Similarly, Pimentel et al. (1996) concluded that carbonates in Paleogene alluvial rocks were the product of groundwater diagenesis, although the carbonates resembled those resulting from pedogenesis. The authors warned that such carbonates have no particular climatic interpretation, nor should they be used to estimate sediment accumulation rates on the basis of their degree of development.

Finally, depending on the nature of the exposure and amount of data available, it may be difficult to distinguish climatic controls from other mechanisms that can produce vertical changes in paleosol properties. The uncertainty that may exist is demonstrated by a study of Lower Carboniferous paleosols by Wright et al. (1991). An upward change from calcrete-bearing Vertisols to paleosols interpreted as ferrolytic Vertisols indicates a change in soil moisture from better drained to more poorly drained conditions. The ferrolytic paleosols also contain an unusual clay assemblage that was attributed to intense leaching. The authors suggested that either a change in local drainage conditions (an intrinsic mechanism) or a climate change to more prolonged precipitation could have been responsible. Although they favored the climatic interpretation, they admitted that this interpretation was tenuous because it was based on only two paleosol profiles from only two localities.

6. Interpreting ancient atmospheres

6.1. Isotopic analysis

Cerling (1991, 1992) provided a model for estimating ancient atmospheric $\rho(CO_2)$ by showing that the isotopic composition of soil carbonate is directly related to soil CO_2, which depends on the concentration of CO_2 in the atmosphere. As noted above, the carbon isotopic composition of soil carbonates is also sensitive to the kind of vegetation (C_3 vs. C_4 plants). The model assumes that the pre-Miocene vegetation consisted entirely of C_3 plants, because various isotopic studies (Cerling, 1992; Cerling et al., 1993; Quade et al., 1994, 1995; Latorre et al., 1997) have shown a major change from C_3 to C_4 vegetation between 8 and 7 Ma. Also of importance for paleosols studies is the depth of the pedogenic carbonate because $\delta^{13}CO_2$ values become more negative downward in a soil profile to a depth of ~ 20 cm.

Using paleosol carbonates and the Cerling model, atmospheric CO_2 values have been estimated for various times in the Phanerozoic (Fig. 12). In a study of pedogenic carbonates of Late Silurian through Pennsylvanian age, Mora et al. (1996) concluded that atmospheric CO_2 levels were high for the Late Silurian (3200–5200 ppm V), then steadily declined

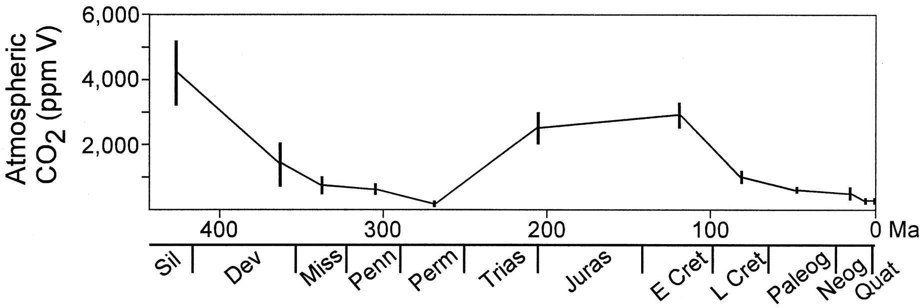


Fig. 12. Changes in atmospheric CO_2 based on isotopic analyses of paleosol carbonates. Paleozoic values are ranges from Mora et al. (1996); Late Cretaceous value from Ghosh et al. (1995); other values from Cerling et al. (1992).

through the Paleozoic (700–2050 ppm V for the Devonian; 450–1000 ppm V for the Mississippian; 450–800 ppm V for the Pennsylvanian). By Permian time, atmospheric CO_2 levels dropped to 150–200 ppm V (Mora et al., 1996). Retallack (1997a) interpreted a similar drop in atmospheric CO_2 from paleosols. This significant decline was linked to the expansion of land plants and to the global climate change that produced extensive late Paleozoic glaciation (Mora et al., 1996).

Cerling (1991) determined that atmospheric $\rho(CO_2)$ then rose during the Late Triassic to Early Jurassic (2000–3000 ppm V). Following high Early Cretaceous atmospheric CO_2 (2500–3300 ppm V), values then fell through the Cenozoic. In a study of pedogenic carbonates from central India, Ghosh et al. (1995) concluded that $\rho(CO_2)$ in the Late Cretaceous atmosphere was 800–1200 ppm V. Levels fell further in the Cenozoic with 600 ppm V estimated from Eocene paleosols, and 400–700 ppm V determined from Miocene paleosols (Cerling, 1991). The current value is approximately 300 ppm V.

6.2. Limitations

To yield reliable $p(CO_2)$ estimates, the carbonate collected from paleosols must be of indisputable pedogenic origin and cannot have undergone post-pedogenic modification. Wright and Vanstone (1991) emphasized that groundwater carbonates pose a potential problem to this method. Groundwater carbonates have different isotopic compositions than overlying pedogenic carbonates but they can be difficult to distinguish from true pedogenic carbonates. Another problem associated with groundwaters is that, with continued sedimentation, a soil becomes buried and moved below the water table. Consequently, carbonate that first formed in the rooted zone can be overprinted by groundwater precipitation. Burial diagenesis is also a potential problem, although studies of lower Eocene paleosols (Cerling, 1991) and Devonian pedogenic carbonates (Driese and Mora, 1993) both show that the $\delta^{13}C$ of paleosol carbonates was only minimally affected by diagenesis.

The particular soil environment in which the carbonate precipitated can also affect its isotopic composition and, thus, the $p(CO_2)$ value it yields. In a study of Devonian Vertisols, Driese and Mora (1993) examined carbonate from two different sources in the paleosols: rhizoliths and pedogenic nodules. They found that the nodule carbonates were isotopically heavier than the rhizolith carbonates. The authors attributed this to the depth at which the carbonates precipitated. The nodules precipitated in the zone of soil cracking, and, because Vertisols commonly develop cracks of 1 m in depth, atmospheric CO_2 may have penetrated deep into the developing soil, resulting in nodules with isotopic values that overestimate atmospheric CO_2. The rhizoliths formed deeper in the soil profile and apparently below the zone of cracking. Consequently, the authors concluded that they provided a more accurate estimate of paleoatmospheric CO_2 than the nodules.

Finally, as noted above, the proportion of C_3 or C_4 vegetation influences the $\delta^{13}C$ values of paleosol carbonates. The Cerling model (1991, 1992) assumes that C_4 vegetation did not appear until late Miocene time. This assumption has been questioned by Wright and Vanstone (1991) who argued that both paleobotanical studies (e.g., Spicer, 1989) and heavy isotopic values obtained from pre-Miocene pedogenic carbonates (e.g., those of Mora et al., 1991) indicated that the early vegetation contained C_4 as well as C_3 plants. The reader is referred to discussion and reply on this topic (Cerling et al., 1992).

And what of the change to vascular plants in the early Paleozoic? As mentioned earlier, Mora et al. (1996) suggested that atmospheric CO_2 levels dropped markedly during mid to late Paleozoic time in response to the expansion of terrestrial vegetation. On the basis of an isotopic study of goethite taken from Ordovician ironstones, Yapp and Poths (1994) suggested that the productivity of pre-vascular plants was similar to that of modern plants. Thus, they believe that, when using the isotopic analysis of paleosol carbonates to estimate paleoatmospheric levels of CO_2, no assumptions or corrections need to be made.

7. Summary

It now appears that soil development, due to the episodic nature of sediment accumulation, is a normal part of the continental sedimentary regime and that many ancient continental deposits contain vertically stacked or multistory paleosols. Because sedimentary rocks comprise 75% of the rocks exposed at the Earth surface (Pettijohn, 1975), and because many of those rocks formed in terrestrial environments subject to pedogenic modification, paleosols are abundant in the geologic record. Their geologic history is also lengthy, extending well back into the Precambrian (e.g., Driese et al., 1995).

The study of pre-Quaternary paleosols and the ways in which the paleosols are being used to help solve various geologic problems have grown in the past decade. Because many ancient soil profiles are easily recognized, exposed over broad areas, and formed almost instantaneously in terms of geologic time (in the range of 2000–30,000 years), they offer a nearly ideal method of correlating deposits in the continental realm, both at a local and at a regional or basinal scale. Paleosols are still an under-utilized aspect of sequence stratigraphy, and this is an area in which future research should focus. Continental successions with multistory paleosols can also provide a continuous record of ancient climatic conditions and climatic changes through time. Additionally, paleosol–landscape analysis can produce a clearer, more complete picture of the environmental conditions and processes operating in ancient continental basins. In particular, recognizing and analyzing paleosol variability at different spatial and temporal scales is important for evaluating how landscapes evolved over time and for assessing the relative significance of autogenic and allogenic controls on landscape evolution.

Acknowledgements

Research contributing to this paper was supported by National Science Foundation Grants EAR-9303959 and EAR-9706115 to MJK. P.D. Gingerich provided invaluable logistical support for fieldwork in the Bighorn Basin. Field assistance during studies of paleosols in the Bighorn Basin was provided by Maureen McHugh, Brian Gwinn, Roberta Yuhas. Constructive reviews for the journal were provided by Greg Retallack and Paul Wright.

References

Abbott, P.L., Minch, J.A., Peterson, G.L., 1976. Pre-Eocene paleosol south of Tijuana, Baja California, Mexico. J. Sediment. Petrol. 46, 355–361.

Aitken, J.F., Flint, S.S., 1996. Variable expressions of interfluvial sequence boundaries in the Brathitt Group (Pennsylvanian), eastern Kentucky, USA. In: Howell, J.A., Aitken, J.F. (Eds.), High Resolution Sequence Stratigraphy: Innovations and Applications. Geol. Soc. London, Spec. Publ. 104, 193–206.

Alonso Zarza, A.M., Wright, V.P., Calvo, J.P., Garcia del Cura, M.A., 1992. Soil–landscape and climatic relationships in the middle Miocene of the Madrid Basin. Sedimentology 39, 17–35.

Arndorff, L., 1993. Lateral relations of deltaic palaeosols from the Lower Jurassic Ronne Formation on the Island of Bornholm, Denmark. Palaeogeogr., Palaeoclimatol., Palaeoecol. 100, 235–250.

Aslan, A., Autin, W.J., 1998. Holocene flood–plain soil formation in the southern lower Mississippi Valley: implications for interpreting alluvial paleosols. Geol. Soc. Am. Bull. 110, 433–449.

Atkinson, C.D., 1986. Tectonic control on alluvial sedimentation as revealed by an ancient catena in the Capella Formation (Eocene) of northern Spain. In: Wright, V.P. (Ed.), Paleosols: Their Recognition and Interpretation. Blackwell Scientific Publications, Oxford, pp. 139–179.

Baldwin, B., Butler, C.O., 1985. Compaction curves. Am. Assoc. Pet. Geol. Bull. 69, 622–626.

Besly, B.M., Fielding, C.R., 1989. Palaeosols in Westphalian coal-bearing and red-bed sequences, central and northern England. Palaeogeogr., Palaeoclimatol., Palaeoecol. 70, 303–330.

Bestland, E.A., 1997. Alluvial terraces and paleosols as indicators of early Oligocene climate change (John Day Formation, Oregon). J. Sediment. Res. 67A, 840–855.

Birkeland, P.W., 1984. Soils and Geomorphology, Oxford University Press, New York, 372 pp.

Blum, M.D., 1994. Genesis and architecture of incised valley fill sequences: a Late Quaternary example from the Colorado River, Gulf Coastal Plain of Texas. In: Weimer, P. Posamentier, H. (Eds.), Siliciclastic Sequence Stratigraphy: Recent Developments and Applications, Am. Ass. Petrol. Geol. Mem. 58, 259–283.

Blum, M.D., Valastro, S. Jr., 1994. Late Quaternary sedimentation, lower Colorado River, Gulf Coastal Plain of Texas. Geol. Soc. Am. Bull. 106, 1002–1016.

Bockheim, J.G., 1980. Solution and use of chronofunctions in studying soil development. Geoderma 24, 71–85.

Bown, T.M., Kraus, M.J., 1981. Lower Eocene alluvial paleosols (Willwood Formation, northwest Wyoming, USA) and their significance for paleoecology paleoclimatology, and basin analysis. Palaeogeogr., Palaeoclimatol., Palaeoecol. 34, 1–30.

Bown, T.M., Kraus, M.J., 1987. Integration of channel and floodplain suites: I. Developmental sequence and lateral relations of alluvial paleosols. J. Sediment. Petrol. 57, 587–601.

Bown, T.M., Kraus, M.J., 1993. Time-stratigraphic reconstruction and integration of paleopedologic, sedimentologic, and biotic events (Willwood Formation, Lower Eocene, northwest Wyoming, USA). Palaios 8, 68–80.

Brady, N.C., 1990. The Nature and Properties of Soils. Macmillan, New York, 621 pp.

Bridge, J.S., Leeder, M.R., 1979. A simulation model of alluvial stratigraphy. Sedimentology 26, 599–623.

Caudill, M.R., Driese, S.G., Mora, C.I., 1996. Preservation of a paleo-Vertisol and an estimate of late Mississippian paleoprecipitation. J. Sediment. Res. 66A, 58–70.

Caudill, M.R., Driese, S.G., Mora, C.I., 1997. Physical compaction of vertic palaeosols: implications for burial diagenesis and palaeo-precipitation estimates. Sedimentology 44, 673–685.

Cerling, T.E., 1984. The stable isotopic composition of modern soil carbonate and its relationship to climate. Earth Planet. Sci. Lett. 71, 229–240.

Cerling, T.E., 1991. Carbon dioxide in the atmosphere: evidence from Cenozoic and Mesozoic paleosols. Am. J. Sci. 291, 377–400.

Cerling, T.E., 1992. Use of carbon isotopes in paleosols as an indicator of the $\rho(CO_2)$ of the paleoatmosphere. Global Biochem. Cycles 6, 307–314.

Cerling, T.E., Quade, J., 1993. Stable carbon and oxygen isotopes in soil carbonates. Climate Change in Continental Isotopic Records. American Geophysical Union, pp. 217–231.

Cerling, T.E., Wright, V.P., Vanstone, S.D., 1992. Further comments on using carbon isotopes in palaeosols to estimate the CO_2 content of the palaeo-atmosphere. J. Geol. Soc. London 149, 673–676.

Cerling, T.E., Wang, Y., Quade, J., 1993. Expansion of C4 ecosystems as an indicator of global ecological change in the late Miocene. Nature 361, 344–345.

Driese, S.G., Mora, C.I., 1993. Physico-chemical environment of pedogenic carbonate formation in Devonian vertic palaeosols, central Appalachians, USA. Sedimentology 40, 199–216.

Driese, S.G., Srinivasan, K., Mora, C.I., Stapor, F.W., 1994. Paleoweathering of Mississippian Monteagle Limestone preceding development of a lower Chesterian transgressive systems tract and sequence boundary, middle Tennessee and northern Alabama. Geol. Soc. Am. Bull. 106, 866–878.

Driese, S.G., Simpson, E.L., Eriksson, K.A., 1995. Redoximorphic paleosols in alluvial and lacustrine deposits, 1.8 Ga Lochness Formation, Mount Isa, Australia; pedogenic processes and implications for paleoclimate. J. Sediment. Res. 65A, 675–689.

Duchaufour, P., 1982. Pedology. Allen and Unwin, London, 448 pp.

Fastovsky, D.E., McSweeney, K., 1987. Paleosols spanning the Cretaceous–Paleogene transition, eastern Montana and western North Dakota. Geol. Soc. Am. Bull. 99, 66–77.

Food and Agriculture Organisation / United Nations Scientific and Cultural Organization, 1974. Soil Map of the World. Vol. I, Legend. Unesco, Paris.

Frakes, L.A., Sun, J., 1994. A carbon isotope record of the upper Chinese loess sequence; estimates of plant types during stadials and interstadials. Palaeogeogr., Palaeoclimatol., Palaeoecol. 108, 183–189.

Friedman, I., O'Neil, J.R., 1977. Compilation of stable isotope fractional factors of geochemical interest. In: Fleischer, M. (Ed.), Data of Geochemistry. U.S. Geol. Surv. Prof. Pap. 440-KK, 12 pp.

Ghosh, P., Bhattacharya, S.K., Jani, R.A., 1995. Palaeoclimate and palaeovegetation in Central India during the Upper Cretaceous based on stable isotope composition of the palaeosol carbonates. Palaeogeogr., Palaeoclimatol., Palaeoecol. 114, 285–296.

Gill, S., Yemane, K., 1996. Implications of a Lower Pennsylvanian Ultisol for equatorial Pangean climates and early, oligotrophic, forest ecosystems. Geology 24, 905–908.

Guccione, M.J., 1993. Grain-size distribution of overbank sediment and its use to locate channel positions. In: Marzo, M., Puigdefabregas, C. (Eds.), Alluvial Sedimentation. Int. Ass. Sediment., Spec. Publ. 17, 185–194.

Hanneman, D.L., Wideman, C.J., 1991. Sequence stratigraphy of Cenozoic continental rocks, southwestern Montana. Geol. Soc. Am. Bull. 103, 1335–1345.

Hanneman, D.L., Wideman, C.J., Halvorson, J.W., 1994. Calcic paleosols: their use in subsurface stratigraphy. Am. Assoc. Pet. Geol. Bull. 78, 1360–1371.

Harden, J.W., Taylor, E.M., 1983. A quantitative comparison of soil development in four climatic regimes. Quat. Res. 20, 342–359.

Holliday, V.T., 1990. Soils and landscape evolution of eolian plains: the Southern High Plains of Texas and New Mexico. In: Knuepfer, P.L.K., McFadden, L.D. (Eds), Soils and Landscape Evolution. Geomorphology 3, 489–515.

Jenny, H., 1941. Factors of Soil Formation. McGraw-Hill, New York, 281 pp.

Joeckel, R.M., 1995. Paleosols below the Ames Marine Unit (Upper Pennsylvanian, Conemaugh Group) in the Appalachian Basin, U.S.A.; variability on an ancient depositional landscape. J. Sediment. Res. 65A, 393–407.

Kleiss, H.J., Fehrenbacher, J.B., 1973. Loess distribution as revealed by mineral variation. Soil Sci. Soc. Am. Proc. 37, 291–295.

Koch, P.L., Zachos, J.C., Gingerich, P.D., 1992. Correlation between isotope records in marine and continental carbon reservoirs near the Paleocene/Eocene boundary. Nature 358, 319–322.

Koch, P.L., Zachos, J.C., Dettman, D.L., 1995. Stable isotope stratigraphy and paleoclimatology of the Paleogene Bighorn Basin (Wyoming, USA). Palaeogeogr., Palaeoclimatol., Palaeoecol. 115, 61–89.

Kraus, M.J., 1987. Integration of channel and floodplain suites: II. Lateral relations of alluvial paleosols. J. Sediment. Petrol. 57, 602–612.

Kraus, M.J., 1992. Alluvial response to differential subsidence: sedimentological analysis aided by remote sensing, Willwood Formation (Eocene), Bighorn Basin, Wyoming, USA. Sedimentology 39, 455–470.

Kraus, M.J., 1996. Avulsion deposits in lower Eocene alluvial rocks, Bighorn Basin, Wyoming. J. Sediment. Res. 66B, 354–363.

Kraus, M.J., 1997. Lower Eocene alluvial Paleosols; pedogenic development, stratigraphic relationships, and Paleosol/landscape associations. Palaeogeogr., Palaeoclimatol., Palaeoecol. 129, 387–406.

Kraus, M.J., Aslan, A., 1993. Eocene hydromorphic paleosols: significance for interpreting ancient floodplain processes. J. Sediment. Petrol. 63, 453–463.

Kraus, M.J., Aslan, A., 1999. Paleosol sequences in floodplain environments: a hierarchical approach. In: Thiry, M. (Ed.), Palaeoweathering, Palaeosurfaces and Related Continental Deposits. Int. Assoc. Sedimentol., Spec. Publ. 27, 303–321.

Kraus, M.J., Bown, T.M., 1993. Short-term sediment accumulation rates determined from Eocene alluvial paleosols. Geology 21, 743–746.

Kraus, M.J., Gwinn, B., 1997. Facies and facies architecture of Paleogene floodplain deposits, Willwood Formation, Bighorn Basin, Wyoming, USA. Sediment. Geol. 114, 33–54.

Kukla, G.J., An, Z., 1989. Loess stratigraphy in Central China. Palaeogeogr., Palaeoclimatol., Palaeoecol. 72, 203–225.

Lander, R.H., Bloch, S., Mehta, S., Atkinson, C.D., 1991. Burial diagenesis of paleosols in the giant Yacheng gas field, People's Republic of China: bearing on illite reaction pathways. J. Sediment. Petrol. 61, 256–268.

Latorre, C., Quade, J., McIntosh, W.C., 1997. The expansion of C_4 grasses and global change in the late Miocene; stable isotope evidence from the Americas. Earth Planet. Sci. Lett. 146, 83–96.

Leckie, D., Fox, C., Tarnocai, C., 1989. Multiple paleosols of the late Albian Boulder Creek formation, British Columbia, Canada. Sedimentology 36, 307–323.

Leeder, M.R., 1975. Pedogenic carbonates and flood sediment accretion rates: a quantitative model for alluvial arid-zone lithofacies. Geol. Mag. 112, 257–270.

Mack, G.H., 1992. Paleosols as an indicator of climatic change at the Early–Late Cretaceous boundary, southwestern New Mexico. J. Sediment. Petrol. 62, 483–494.

Mack, G.H., James, W.C., 1994. Paleoclimate and the global distribution of paleosols. J. Geol. 102, 360–366.

Mack, G.H., Cole, D.R., Giordano, T.H., Schaal, W.C., Barcelos, J.H., 1991. Paleoclimatic controls on stable oxygen and carbon isotopes in caliche of the Abo Formation (Permian), south-central New Mexico, U.S.A. J. Sediment. Petrol. 61, 458–472.

Mack, G.H., James, W.C., Monger, H.C., 1993. Classification of paleosols. Geol. Soc. Am. Bull. 105, 129–136.

Marriott, S.B., 1996. Analysis and modelling of overbank deposits. In: Anderson, M.G., Walling, D.E., Bates, P.D. (Eds.), Floodplain Processes. Wiley, Chichester, pp. 63–93.

Marriott, S.B., Wright, V.P., 1993. Paleosols as indicators of geomorphic stability in two Old Red Sandstone alluvial suites, South Wales. J. Geol. Soc. London 150, 1109–1120.

McCahon, T.J., Miller, K.B., 1997. Climatic significance of natric horizons in Permian (Asselian) palaeosols of north-central Kansas, USA. Sedimentology 44, 113–125.

McCarthy, P.J., Plint, A.G., 1998. Recognition of interfluve sequence boundaries: integrating paleopedology and sequence stratigraphy. Geology 26, 387–390.

McCarthy, P.J., Martini, I.P., Leckie, D.A., 1997. Anatomy and evolution of a Lower Cretaceous alluvial plain; sedimentology and palaeosols in the upper Blairmore Group, south-western Alberta, Canada. Sedimentology 44, 197–220.

McDonald, E.V., Busacca, A.J., 1990. Interaction between aggrading geomorphic surfaces and the formation of a Late Pleistocene paleosol in the Palouse loess of eastern Washington state. In: Knuepfer, P.L.K., McFadden, L.D. (Eds), Soils and Landscape Evolution. Geomorphology 3, 449–470.

McFadden, L.D., Knuepfer, P.L.K., 1990, Soil geomorphology; the linkage of pedology and surficial processes. In: Knuepfer, P.L.K., McFadden, L.D. (Eds), Soils and Landscape Evolution. Geomorphology 3, 197–205.

Miller, K.B., McCahon, T.J., West, R.R., 1996. Lower Permian (Wolfcampian) Paleosol-bearing cycles of the U.S. Midcontinent; evidence of climatic cyclicity. J. Sediment. Res. 66A, 71–84.

Mora, C.I., Driese, S.G., Seager, P.G., 1991. Carbon dioxide in the Paleozoic atmosphere: evidence from carbon-isotope compositions of pedogenic carbonate. Geology 19, 1017–1020.

Mora, C.I., Driese, S.G., Colarusso, L.A., 1996. Middle to late Paleozoic atmospheric CO_2 levels from soil carbonate and organic matter. Science 271, 1105–1107.

Mora, C.I., Sheldon, B.T., Elliott, W.C., Driese, S.G., 1998. An oxygen isotope study of illite and calcite in three Appalachian Paleozoic vertic paleosols. J. Sediment. Res. 68, 456–464.

Morrison, R.B., 1967. Principles of Quaternary soil stratigraphy. In: Morrison, R.B., Wright, H.E. (Eds.), Means of Correlation of Quaternary Successions, Int. Assoc. Quat. Res. (INQUA), VII Congr., Proc. 9, 1–69.

Morrison, R.B., 1978. Quaternary soil stratigraphy—concepts, methods, and problems. In: Mahaney, W.C. (Ed.), Quaternary Soils, Geo Abstracts, Norwich. pp. 77–108.

Nadon, G.C., Issler, D.R., 1997. The compaction of floodplain sediments: timing, magnitude and implications. Geosci. Can. 24, 37–43.

North, C.P., 1996. The prediction and modelling of subsurface fluvial stratigraphy. In: Carling, P.A., Dawson, M.R. (Ed.), Advances in Fluvial Dynamics and Stratigraphy. Wiley, Chichester, pp. 395–508.

Pecsi, M., 1995. The role of principles and methods in loess-paleosol investigations. GeoJournal 36, 117–131.

Pettijohn, F.J., 1975. Sedimentary Rocks, Third edn. New York, Harper and Row, 628 pp.

Pimentel, N.L., Wright, V.P., Azevedo, T.M., 1996. Distinguishing early groundwater alteration effects from pedogenesis in ancient alluvial basins: examples from the Palaeogene of southern Portugal. Sediment. Geol. 105, 1–10.

Platt, N.H., Keller, B., 1992. Distal alluvial deposits in a foreland basin setting—the lower freshwater Molasse (lower Miocene), Switzerland: sedimentology, architecture and palaeosols. Sedimentology 39, 545–565.

Quade, J., Cerling, T.E., Bowman, J.R., 1989. Development of Asian monsoon revealed by marked ecological shift during the latest Miocene in northern Pakistan. Nature 342, 163–165.

Quade, J., Solounias, N., Cerling, T.E., 1994. Stable isotopic evidence from Paleosol carbonates and fossil teeth in Greece for forest or woodlands over the past 11 Ma. Palaeogeogr., Palaeoclimatol., Palaeoecol. 108, 41–53.

Quade, J., Cater, J.M.L., Ojha, T.P., Adam, J., Harrison, T.M., 1995. Late Miocene environmental change in Nepal and the northern Indian subcontinent; stable isotopic evidence from paleosols. Geol. Soc. Am. Bull. 107, 1381–1397.

Rasbury, E.T., Hanson, G.N., Meyers, W.J., Holt, W.E., Goldstein, R.H., Saller, A.H., 1998. U–Pb dates of paleosols: constraints on Late Paleozoic cycle durations and boundary ages. Geology 26, 403–406.

Retallack, G.J., 1983. A paleopedological approach to the interpretation of terrestrial sedimentary rocks: the mid Tertiary fossil soils of Badlands National Park, South Dakota. Geol. Soc. Am. Bull. 94, 823–840.

Retallack, G.J., 1984. Completeness of the rock and fossil record: some estimates using fossil soils. Paleobiology 10, 59–78.

Retallack, G.J., 1991. Untangling the effects of burial alteration and ancient soil formation. Ann. Rev. Earth Planet. Sci. 19, 183–206.

Retallack, G.J., 1993. Classification of paleosols: discussion and reply. Discussion Geol. Soc. Am. Bull. 105, 1635–1636.

Retallack, G.J., 1994. The environmental factor approach to the interpretation of paleosols. In: Amundson, R., Harden, J., Singer, M. (Eds.), Factors of Soil Formation: a Fiftieth Anniversary Retrospective. Soil Sci. Soc. Am., Madison, WI, pp. 31–64.

Retallack, G.J., 1997a. Early forest soils and their role in Devonian global change. Science 276, 583–585.

Retallack, G.J., 1997b. A Colour Guide to Paleosols. Wiley, Chichester, 175 pp.

Retallack, G.J., 1998. Fossil soils and completeness of the rock and fossil records. In: Donovan, S.K., Paul, C.R.C. (Eds.), The Adequacy of the Fossil Record. Wiley, Chichester, pp. 133–163.

Retallack, G.J., German-Heins, J., 1994. Evidence from paleosols for the geological antiquity of rain-forest. Science 265, 499–502.

Richmond, G.M., 1962. Quaternary stratigraphy of the La Sal Mountains, Utah. U.S. Geol. Surv., Prof. Pap. 324, 135 pp.

Ritzma, H.R., 1955. Late Cretaceous and Early Cenozoic structural pattern, southern Rock Springs Uplift Wyoming. Wyo. Geol. Assoc., 10th Ann. Field Conf. Guidebook, 135–137.

Robert, C., Kennett, J.P., 1994. Antarctic subtropical humid episode at the Paleocene–Eocene boundary: clay–mineral evidence. Geology 22, 211–214.

Ruhe, R.V., 1983. Aspects of Holocene pedology in the United States. In: Wright, H.E., Jr. (Ed.), Late-Quaternary Environments of the United States, Vol. 2. The Holocene. Univ. of Minnesota Press, Minneapolis,

Schultz, C.B., Tanner, L.G., Harvey, C., 1955. Paleosols of the Oligocene of Nebraska. Bull. Univ. Neb. State Mus. 4, 1–16.

Schumacher, B.A., Day, W.J., Amacher, M.C., Miller, B.J., 1988. Soils of the Mississippi River alluvial plain in Louisiana. Louisiana Agri. Exp. Sta. Bull. 796, 275 pp.

Shackleton, N.J., 1986. Paleogene stable isotope events. Palaeogeogr., Palaeoclimatol., Palaeoecol. 57, 91–102.

Shanley, K.W., McCabe, P.J., 1994. Perspectives on the sequence stratigraphy of continental strata. Am. Assoc. Pet. Geol. Bull. 78, 544–568.

Slate, J.L., Smith, G.A., Wang, Y., Cerling, T.E., 1996. Carbonate-paleosol genesis in the Plio-Pleistocene St. David Formation, southeastern Arizona. J. Sediment. Res. 66A, 85–94.

Smith, G.D., 1942. Illinois loess—variations in its properties and distribution: a pedologic interpretation. Univ. Illinois Agri. Exp. Sta. Bull. 490, 137–184.

Smith, R.M.H., 1990. Alluvial paleosols and pedofacies sequences in the Permian Lower Beaufort of the southwestern Karoo Basin, South Africa. J. Sediment. Petrol. 60, 258–276.

Soil Survey Staff, 1975. Soil taxonomy. U.S. Dept. Agri. Handbook 436, 754 pp.

Soil Survey Staff, 1998. Keys to Soil Taxonomy, 8th edn. U.S. Dept. Agri. Nat. Res. Conserv. Serv., 327 p.

Soreghan, G.S., Elmore, R.D., Katz, B., Cogoini, M., Banerjee, S., 1997. Pedogenically enhanced magnetic susceptibility variations preserved in Paleozoic loessite. Geology 25, 1003–1006.

Spicer, R.A., 1989. Physiological characteristics of land plants in relation environment through time. Trans. R. Soc. Edinburgh: Earth Sci. 44, 321–329.

Stern, L.A., Chamberlain, C.P., Reynolds, R.C., Johnson, G.D.,

1997. Oxygen isotope evidence of climate change from pedogenic clay minerals in the Himalayan molasse. Geochim. Cosmochim. Acta. 61, 731–744.

Summerfield, M.A., 1991. Global Geomorphology: an Introduction to the Study of Landforms. Wiley, New York, 537 pp.

Tandon, S.K., Gibling, M.R., 1994. Calcrete and coal in late Carboniferous cyclothems of Nova Scotia, Canada: climate and sea-level changes linked. Geology 22, 755–758.

Tandon, S.K., Gibling, M.R., 1997. Calcretes at sequence boundaries in upper Carboniferous cyclothems of the Sydney Basin, Atlantic Canada. Sediment. Geol. 112, 43–67.

Tandon, S.K., Good, A., Andrews, J.E., Dennis, P.F., 1995. Palaeoenvironments of the dinosaur-bearing Lameta Beds (Maastrichtian), Narmada Valley, Central India. Palaeogeogr., Palaeoclimatol., Palaeoecol. 117, 153–184.

Walling, D.E., Quine, T.A., He, O., 1992. Investigating contemporary rates of floodplain sedimentation. In: Carling, P.A., Petts, G.E. (Eds.), Lowland Floodplain Rivers: Geomorphological Perspectives. Wiley, Chichester, pp. 165–184.

Webb, G.E., 1994. Paleokarst, paleosol, and rocky-shore deposits at the Mississipian–Pennsylvanian unconformity, northwestern Arkansas. Geol. Soc. Am. Bull. 106, 634–648.

Wheeler, H.E., 1958. Time-stratigraphy. Am. Assoc. Pet. Geol. Bull. 42, 1047–1063.

Wright, V.P., 1990. Estimating rates of calcrete formation and sediment accretion in ancient alluvial deposits. Geol. Mag. 127, 273–276.

Wright, V.P., 1992a. Paleosol recognition; a guide to early diagenesis in terrestrial settings. In: Wolf, K.H., Chilingarian, G.V. (Eds.), Diagenesis, III. Elsevier., Amsterdam, pp. 591–619.

Wright, V.P., 1992b. Paleopedology: stratigraphic relationships and empirical models. In: Martini, I.P., Chesworth, W. (Eds.), Weathering, Soils and Paleosols. Elsevier, Amsterdam, pp. 475–499.

Wright, V.P., 1994. Paleosols in shallow marine carbonate sequences. Earth-Sci. Rev. 35, 367–395.

Wright, V.P., Marriott, S.B., 1993. The sequence stratigraphy of fluvial depositional systems: the role of floodplain storage. Sediment. Geol. 86, 203–210.

Wright, V.P., Marriott, S.B., 1996. A quantitative approach to soil occurrence in alluvial deposits and its application to the Old Red Sandstone of Britain. J. Geol. Soc. London 153, 907–913.

Wright, V.P., Platt, N.H., 1995. Seasonal wetland carbonate sequences and dynamic catenas; a re-appraisal of palustrine limestones. Sediment. Geol. 99, 65–71.

Wright, V.P., Robinson, D., 1988. Early Carboniferous floodplain deposits from South Wales: a case study of the controls on palaeosol development. J. Geol. Soc. London 145, 847–857.

Wright, V.P., Vanstone, S.D., 1991. Assessing the carbon dioxide content of ancient atmospheres using palaeo-calcretes: theoretical and empirical constraints. J. Geol. Soc. London 148, 945–947.

Wright, V.P., Vanstone, S.D., Robinson, D., 1991. Ferrolysis in Arundian alluvial palaeosols, evidence of a shift in the Early Carboniferous monsoonal system. J. Geol. Soc. London 148, 9–12.

Yapp, C.J., Poths, H., 1994. Productivity of pre-vascular continental biota inferred from the $Fe(CO_3)OH$ content of goethite. Nature 368, 49–51.

Ye, L., 1995. Paleosols in the Upper Guantae Formation (Miocene) of the Gudong oil field and their application to the correlation of fluvial deposits. Am. Assoc. Pet. Geol. Bull. 79, 981–988.

Zachos, J.C., Lohmann, K.C., Walker, J.C.G., Wise, S.W., 1993. Abrupt climate change and transient climates during the Paleogene: a marine perspective. J. Geol. 101, 191–213.

Mary Kraus has been a professor at the University of Colorado since 1983. She holds degree from Yale University (B.S., Biology), the University of Wyoming (M.S., Geology) and the University of Colorado (Ph.D., Geology). Her research and publications focus on (1) understanding the physical processes by which ancient river floodplains were constructed; (2) using floodplain paleosols to reconstruct ancient landscapes at different spatial and temporal scales; and (3) integrating the study of paleosols with the more traditional study of channel sandstone bodies to understand the factors that influence alluvial architecture.