Chapter 9
Paleopedology as a Tool for Reconstructing Paleoenvironments and Paleoecology

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Abstract Soils form as a product of physical, chemical, and biological activity at the outermost veneer of Earth's surface. Once buried and incorporated into the sedimentary record, these soils, now paleosols, preserve archives of ancient climates, ecosystems, and sedimentary systems. Paleopedology, the study of paleosols, includes qualitative interpretation of physical characteristics and quantitative analysis of geochemical and mineralogical assays. In this chapter, the paleosol macroscopic, micromorphological, mineralogical, and geochemical indicators of paleoecology are discussed with emphasis on basic analytical and interpretative techniques. These data can reveal a breadth of site-specific interpretations of vegetation, sedimentary processes, climatic variables, and durations of landscape stability. The well-known soil-forming factors are presented as a theoretical framework for understanding landscape-scale soil evolution through time. Vertical and lateral patterns of stacked paleosols that appear in the rock record are discussed in order to address practical approaches to identifying and describing paleosols in the field. This chapter emphasizes a robust multi-proxy approach to paleopedology that combines soil stratigraphy, morphology, mineralogy, biology, and chemistry to provide an in-depth understanding of paleoecology.

Keywords Paleo-Critical Zone • Soil • Paleosol • Paleocatena • Paleoenvironments • Human evolution

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Theoretical Background

Introduction

Soils are the product of organic matter accumulation and mineral weathering on the terrestrial surface, integrating the dynamic physical, biological, and chemical interactions that take place over time. Soils should not be confused with sediment, a material product of weathering and erosion occurring elsewhere and subject to transport by wind, water, or gravity. In contrast, soils form *in situ* and are the alteration products of weathered sediment, rock, and/or organic components. Dramatic changes in mineralogy and porosity accompany pedogenesis, which, along with bioturbation, act to disrupt primary depositional bedding. These processes impart new structural and chemical characteristics that are recognizable using field and lab techniques. Such patterns are preserved as soil profiles and can be buried and incorporated into the stratigraphic record (Fig. 9.1A). The resulting paleosols appear in many terrestrial depositional successions as intervals with variegated color, surface texture, and recessed erosional relief in outcrop.

State-Factor Approach

Paleopedology uses concepts and tools of Earth systems science to understand paleosols in the Quaternary and pre-Quaternary sedimentary record (Yaalon 1971; Retallack 2001; Sheldon and Tabor 2009; Tabor and Myers 2015).

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The conceptual framework of most paleosol-based paleoenvironmental studies relies upon the fundamental state-factor theory for modern soils (Jenny 1941):

$$S\,or\,s = f(cl, o, r, p, t, \ldots), \quad (1)$$

where S is the soil and s is any soil property, and cl is climate, o is organisms, r is relief, p is parent material, and t is time. The series of dots, or ellipsis, represent other possible unaccounted for factors influencing S. For modern soils, Eq. 1 shows that the five primary factors (cl, o, r, p, t) influence the direction and magnitude of pedogenesis, S, or in some cases seem to predict S (Jenny 1941). For example, modern climosequence relationships that relate S to cl are applied to paleosols using an environmental-factor approach (*sensu* Retallack 1994), where s is measured in the paleosol using a variety of bulk and isotope geochemical, magnetic mineral, and physical proxies (Maynard 1992; Maher and Thompson 1995; Maher 1998; Ludvigson et al. 1998; Birkeland 1999; Retallack 2001; Sheldon et al. 2002; Holliday 2004; Driese et al. 2005; Dworkin et al. 2005; Retallack 2005; Nordt et al. 2006; Sheldon 2006; Sheldon and Tabor 2009; Nordt and Driese 2010a, b; Passey et al. 2010; Retallack and Huang 2010; Gulbranson et al. 2011; Gallagher and Sheldon 2013; Ludvigson et al. 2013; Hyland et al. 2015). The results of such soil-sequence studies allow for the development of empirical, quantitative relationships between soil properties and the state-factor studied (e.g., climate). The reader is directed to Rasmussen et al. (2005) and Rasmussen and Tabor (2007) and references therein for an example of alternative, quantitative frameworks for pedogenesis.

A Brief History of Paleosols in Paleoenvironmental Studies

The application of paleopedology as a tool for paleoecological reconstruction originated in the mid- to late 19th century, when geologists used buried forest beds to correlate glacial stratigraphy in the Midwestern USA, and the early 20th century, when many studies in the USA were focused on identifying the environments of "early man", (i.e., Paleoindian culture) (Leighton 1937; Holliday 2004, 2006). The accumulated knowledge of the early to middle 20th century likely led to the formal designation of the field as paleopedology at International Union for Quaternary Science (INQUA) meetings, led by pioneering paleopedologists Ruhe and Yaalon, which eventually led to two highly influential volumes (Ruhe 1965; Yaalon 1971).

The study of paleosols in the past 50 years has seen advancements through both pre-Quaternary and Quaternary disciplines. Some of the first pre-Quaternary investigations into paleoecology using paleosols can be traced to Retallack's (1983) systematic study of Eocene-Oligocene paleosols in the Badlands of South Dakota, USA. Retallack's (1983) work and successive studies ushered in an era of paleopedological research in a wide range of geologic settings (e.g., Bown and Kraus 1987; Bestland et al. 1997; Terry 2001). The science of paleopedology gradually evolved to become more quantitative with an increasing number of numerical proxies for climate, vegetation, and edaphic soil properties (Retallack 1994; An and Porter 1997; Maher 1998; Sheldon et al. 2002; Driese and Ober 2005; Dworkin et al. 2005; Retallack 2005; Prochnow et al. 2006; Kraus and Hasiotis 2006; Nordt and Driese 2010a, b; Retallack and Huang 2010; Gulbranson et al. 2011; Gallagher and Sheldon 2013; Trendell et al. 2013a, b; Hyland et al. 2015).

Paleosols are important paleoecological archives because they are records of Critical Zones of the Earth's past, preserving the *in situ* substrate of ancient landscapes (Amundson et al. 2007; Leopold et al. 2011; Nordt et al. 2012, 2013; Marin-Spiotta et al. 2014). This approach to studying paleo-Critical Zones (paleo-CZs) continues to gain traction as a multi-disciplinary method for understanding the Earth's environments and processes. Recent work has focused on unraveling and quantifying the biogeochemical processes operating within a paleosol system (Nordt and Driese 2010b; Nordt et al. 2012, 2013), comparable with modern Critical Zone studies (Fig. 9.1A; Amundson et al. 2007; Brantley et al. 2007). In light of these advances, this chapter establishes a theoretical framework for the field of paleopedology through the lens of a paleo-CZ and is applicable throughout geologic time.

This chapter does not review isotope geochemical methods in paleopedology and paleoenvironmental reconstruction, which can be found elsewhere (Nordt 2001; Sheldon and Tabor 2009; Cerling 2014; Levin 2015; Tabor and Myers 2015; Zamanian et al. 2016; Berke 2017). Much of the information compiled in this chapter was adapted from journal articles and books published by pioneers in the field of Quaternary and pre-Quaternary paleopedology (e.g., Yaalon 1971; Brimhall and Dietrich 1987; Birkeland 1999; Retallack 2001; Anderson et al. 2002; Holliday 2004; Sheldon and Tabor 2009; Nordt et al. 2013; Schaetzl and Thompson 2015; Tabor and Myers 2015). The reader is encouraged to consult these publications, which contain conceptual and methodological frameworks for analyzing soils and paleosols.

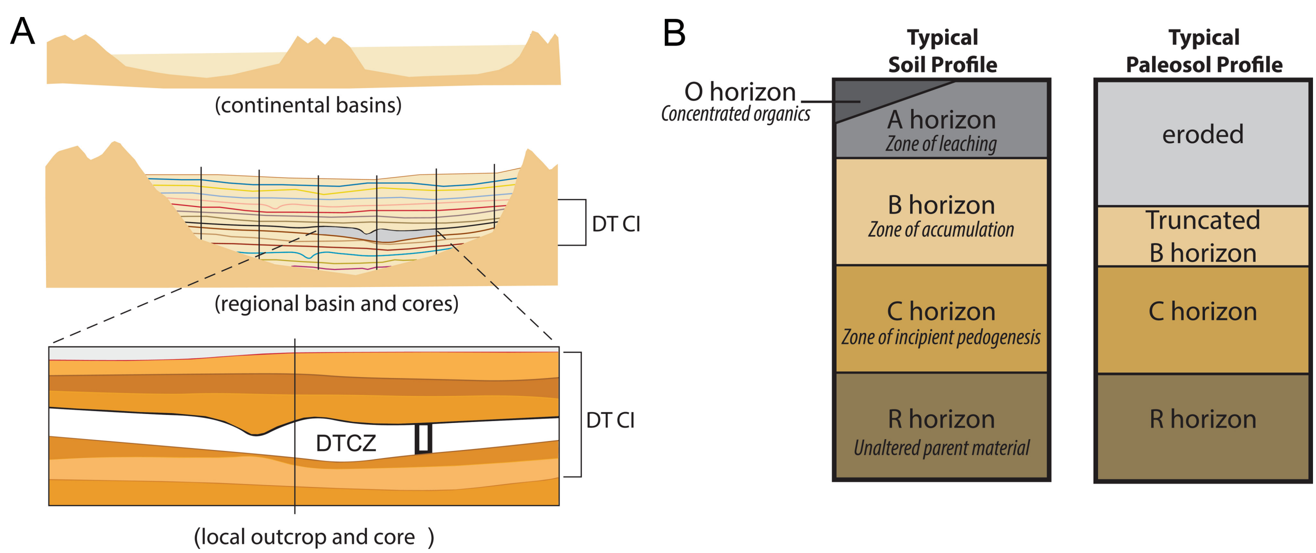

Fig. 9.1 A, Schematic cross-section of a paleo-Critical Zone reproduced from Nordt and Driese (2013). Top Panel: Alluvial deposits in a continental-scale basin. Middle Panel: Regional basin with paleo-Critical Zones illustrated by colored lines that could be studied in outcrop or using sediment cores indicated by vertical black lines. Lower Panel: Local outcrop of an alluvial succession with a paleo-Critical Zones and paleosol. DTCI = Deep time critical interval. The black box indicates the location of Fig. 9.1B. B, A typical soil profile and typical paleosol profile preserved after burial and diagenesis. Modified from Tabor and Myers (2015)

Approaches

Macroscale Physical Characterization

Field observation is the first step for identifying and describing paleosols. This step relies upon observing paleosol morphological features with unassisted vision or a hand lens with ~10× magnification. Much like modern soils, paleosols are identified in the field using color, texture, and weathering features within peds or along former voids (i.e., ped/void features). A number of features may suggest the presence of a paleosol: authigenic mineralization, rhizoliths, coloration, bioturbation, presence of soil aggregates (peds), redistribution of primary/authigenic minerals, shrink-swell features, disruption of primary sedimentary bedding, and deflation surfaces or biomantles as indicated by stone lines. After the paleosol(s) has been identified, a trench should be excavated to a depth that exposes sediment or rock unaffected by modern processes.

Field description and horizon delineation: The description of some paleosols differs from soils due to the inhibiting effects of compaction and lithification on the measurement of many properties. We advocate for describing paleosols in as much detail as the geologic setting allows, as these data will aid in more informed interpretations of paleolandscapes, paleoclimate, and paleovegetation. The reader is directed to the Illustrated Guide to Soil Taxonomy (Soil Survey Staff 2014a) and Keys to Soil Taxonomy (Soil Survey Staff 2006) for standard protocols for field description and horizon designation.

Paleosol horizons should be described using a standardized, published descriptive nomenclature, such as U.S. Department of Agriculture (USDA) soil nomenclature (Tables 9.1 and 9.2; Schoeneberger et al. 2012). A typical soil profile includes the O, A, B, C and R master horizons, where O is the zone of concentrated organics, A is the zone of leaching, B is the zone of accumulation, C is unaltered or unconsolidated parent material, and R is consolidated parent material. The A and O horizons are typically not preserved or recognized in paleosols due to erosion or oxidation of organic matter during the burial process (Fig. 9.1B). Less common master horizons are described in Table 9.1. Stratigraphic successions preserved in the rock record consist of aggrading sedimentary deposits, and therefore many paleosols do not have R horizons.

Subordinate indicators are added as lowercase suffixes to master horizons to identify the presence of pedogenic minerals, organic complexes, or other distinct features (Table 9.2). For example, common types of B horizons include Bk (accumulation of carbonate) and Bt (accumulation of translocated clay; Fig. 9.2B, H). The naming of soil horizons begins with field description but is frequently modified after laboratory and micromorphologic analysis (Birkeland 1999).

Detailed field descriptions of paleosol profiles include texture, peds, mineralizations, color, the presence of fossil roots, trace fossils and any other notable features (e.g., Schoeneberger et al. 2012). Paleosol texture is equivalent to grain size in sedimentology and is described in the field or measured in the laboratory. Peds should be identified and described with respect to size, shape, and grade (Fig. 9.3;

Table 9.1 Master soil horizons that may be observed in paleosols and their paleoecological implications. Modified from Guthrie and Witty (1982), Soil Classification Working Group (1998), and Schaetzl and Thompson (2015)

Master horizon	Characteristics	Paleoecological implications if present in paleosols
O	Layers dominated by organic material (litter and humus) in various stages of decomposition	*Forested environment with high biomass input*
A	Mineral horizons that formed at the surface or below an O horizon that are characterized by the accumulation of humified organic matter mixed with the mineral fraction	*Zone of bioturbation and highest organic content in the mineral soil. A horizons may best preserve signatures of past vegetation, particularly as root traces, phytoliths, biomarkers, and/or isotopic composition of bulk organics*
V	Mineral horizons that formed at the soil surface, or below a layer of rock fragments, e.g., desert pavement, or a physical or biological crust, and which are characterized by the predominance of vesicular pores	*Desert ecosystem, scrub vegetation likely*
E	Light-colored mineral horizons in which the main feature is loss of weatherable minerals. The result is often a concentration of uncoated quartz grains and other resistant minerals and poor soil structure	*Commonly found in forested environments*
B	Subsurface mineral horizons dominated by: (1) illuvial accumulation of clay, iron, aluminum, humus, etc.; (2) removal, addition, or transformation of primary carbonates or gypsum; (3) residual concentrations of sesquioxides; (4) loss of all or most sedimentary structure; (5) brittleness, and/or (6) gleying	*Paleoecological implications vary depending on the horizon suffix* (Table 9.2)
C	Mineral horizons that retain parent material properties and have thus been less affected by pedogenesis	*If C horizon is composed of allogenic parent material (e.g., alluvium, dust), then the organic matter may also be allogenic (detrital) and hold a basin-wide signature*
R	Consolidated hard, continuous bedrock	*N/A*

Table 9.3; Schoeneberger et al. 2012). Ped types are related to parent materials, intra-profile water cycling, biotic activity, and degree of pedogenic alteration, all of which have net paleoecological implications (Table 9.3). For example, columnar peds, which are vertically oriented with round tops, form as a result of clay flocculation in saline soils of arid and semi-arid environments (Fig. 9.3). Such soils tend to exclude most plant types due to the combined effect of low porosity, which inhibits root growth, and nutrient limitation (Qadir and Shubert 2002).

Recognition of the pedogenic redistribution of primary minerals and precipitation of secondary minerals can offer support for interpretations of climate, redox conditions, or soil maturity. Clays are translocated downward with soil water from eluvial (leaching) to illuvial (accumulating) horizons during wetting events. Subsequent drying causes the suspended clays to adhere to void features, which include the surfaces of peds, slickensides, root traces, and tubules. Calcium carbonate precipitates along root traces as rhizoliths or as discrete nodules in the matrix of Bk horizons. Similarly, Fe–Mn masses precipitate along voids or as hard matrix nodules as a result of alternating redox conditions. Salts, such as barite or gypsum, typically accumulate as discrete masses in the matrix of soils with extreme water stress. Geochemical and mineralogical approaches to the characterization of pedogenic minerals are addressed below.

The color of the paleosol matrix, mineralizations, and other notable features are characterized using Munsell® soil color charts. Matrix color can be used to identify soil organic matter (SOM) input and iron speciation associated with redox states but should only be used as an initial tool for description before further laboratory analysis (Holliday 2004). Darker colors, such as the dark browns or dark grayish browns found in buried A horizons, may suggest higher SOM production. However, dark colors can also reflect parent materials, especially for Mg-smectite clays (Fig. 9.2D; Beverly et al. 2014, 2015) or mafic igneous rocks.

The oxidation state and mineralogy of Fe imparts distinct color characteristics on soils and paleosols (Schwertmann 1993; Birkeland 1999) and may directly reflect local environmental conditions such as drainage, fire, temperature, and duration of soil development (Table 9.4; Holliday 2004). Green, blue, or gray colors often suggest poorly-drained, anaerobic conditions (Vepraskas 1992, 2001). Prolonged saturation reduces the amount of available oxygen for soil respiration, which leads to microbially-mediated reduction of Fe and Mn species and an overall gray appearance (Vepraskas and Faulkner 2001). Re-precipitation of Fe and Mn in localized oxygen-rich zones creates a mottled appearance (Fig. 9.2C). These processes frequently occur in wetland or lake-margin paleosols proximal to paleo-lakes or on paleo-floodplains (Retallack 1994; Ufnar et al. 2004;

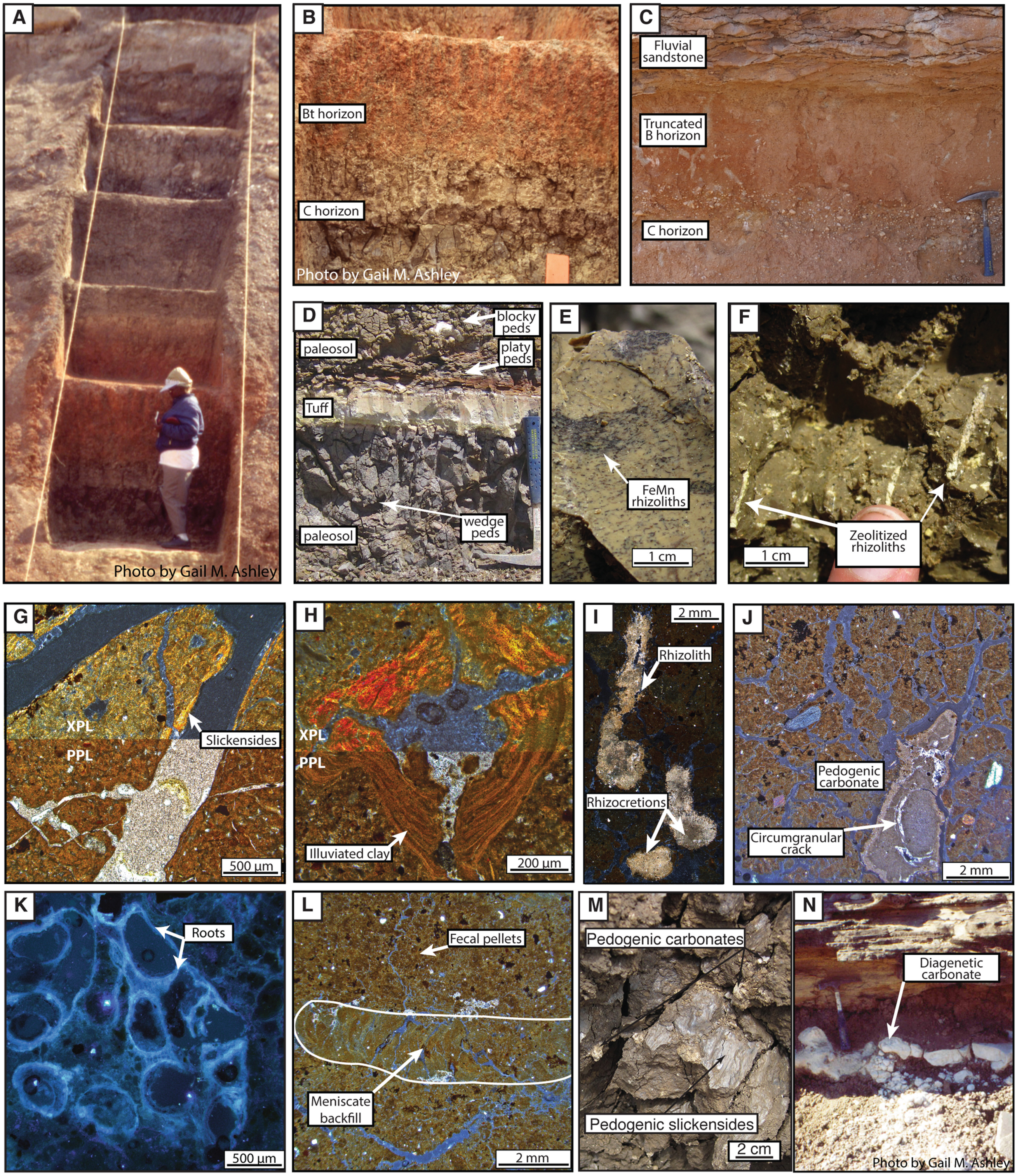

Fig. 9.2 Photographs of pedogenic features in the field and in micromorphology. A, Oxidized and volcaniclastic-rich paleo-Aridisol from Olduvai Gorge, Tanzania. B, Close-up of lowermost step in trench in Fig. 9.2A, which has excellent horizonation, clay illuviation, and ped structure (angular blocky and columnar). C, Fining upward, coarse-grained alluvial paleosol from the Chalbi Desert, Kenya. D, Stacked paleo-Vertisols from Olduvai Gorge separated by a tuff with examples of blocky, platy, and wedge peds. E, FeMn filled rhizoliths from Olduvai Gorge (Reproduced from Beverly et al. 2014). F, Zeolitized rhizoliths from Olduvai Gorge (Reproduced from Beverly et al. 2014). G, Pedogenic B-fabric (Reproduced from Beverly et al. 2015a). H, Illuviated clay in a pore (Reproduced from Beverly et al. 2015a). I, Carbonate rhizolith and rhizocretions in thin section from Karungu, Kenya. J, Pedogenic carbonate in thin section with circumgranular cracking (Reproduced from Beverly et al. 2015a). K, Preserved root structures identified in a paleosol using UVf (Reproduce from Beverly et al. 2015b). L, Fecal pellets and meniscate backfill (Reproduced from Beverly et al. 2015a). M, Pedogenic slickensides in a paleo-Vertisol from Karungu, Kenya, and pedogenic carbonate nodule. N, Non-pedogenic carbonate from Olduvai Gorge. Groundwater carbonate precipitates after primary deposition. Note rock hammer for scale

Table 9.2 Suffix symbols used with master soil horizons and their possible paleoecological implications when observed in paleosols. Modified from Guthrie and Witty (1982), Soil Classification Working Group (1998), Soil Survey Staff (2010), and Schaetzl and Thompson (2015)

Suffix	Used with master horizons	Characteristics/comments	Paleoecological implications
p	O, A	Plowed, tilled, or otherwise disturbed surface layer	*Humans likely played significant role in shaping local ecosystem*
c	B, C	Presence of concretions or nodules, usually of Fe, Al, Mn, or Ti	*Water table fluctuations*
h	B	Accumulation of amorphous and dispersable organic materials, low Munsell value and chroma	*Acidic soil environment; frequently observed in coniferous or boreal forests; Observed in cool-damp shrub environments; acid parent materials (e.g., quartz sand) may also influence development*
j	B, C	Accumulation of jarosite (iron hydroxyl sulfate), as either ped coatings or nodules	*Poorly-drained environment, frequently observed in tidal, swamp, or marsh settings*
k	B, C	Accumulation of visible pedogenic calcium carbonate (<50% by volume) in filaments, soft masses, nodules, pendants, and finely disseminated in matrix	*May occur in arid or semi-arid regions with seasonal wetness; Also known to occur in poorly-drained alkaline environments*
kk	B, C	Engulfment of horizon by pedogenic calcium carbonate ($\geq$50% by volume); coating particles and filling pores	*May occur in arid or semi-arid regions; Indicate protracted surface stability; Often confused with calcium carbonate layers that precipitated in or near the phreatic zone*
n	B, C	Accumulation of exchangeable sodium	*Frequently observed in arid or semi-arid regions*
q	B, C	Accumulation of secondary silica as concretions, opal, etc.	*May occur in arid or semi-arid regions with seasonal wetness; commonly associated with volcanic materials*
s	B	Illuvial accumulation of amorphous, dispersible sesquioxides (Fe- and Al-oxides) and organic matter	*Associated with humid forests*
t	B, C	Illuvial accumulation of clay with clay coatings, bridges, or lamellae	*Well-drained soil where water moves through profile to transport clay with periodic drying of soil when clay adheres to pores; Indicate protracted surface stability*
y	B	Accumulation of pedogenic gypsum (<50% by volume)	*Frequently observed in arid or playa-like environments; Shallow pedogenic gypsum more associated with arid environments; may also results from precipitation near/in the phreatic zone*
yy	B	Horizon dominated by pedogenic gypum where crystals disrupt pedogenic or sedimentary features ($\geq$50% by volume)	*Same as y suffix*
z	B	Accumulation of salts more soluble than gypsum such as NaCl	*May occur in arid or semi-arid regions*
w	B	Weakly expressed color or structural development; used when no other suffix is applicable	*Weak pedogenesis*
x	E, B	Horizon that is firm, brittle, and physically root restrictive	*Restrictive of water movement through the soil*
o	B, C	Residual accumulation of sesquioxides (Fe- and Al-oxides)	*May occur in humid tropical or subtropical environments; Typified by intense weathering and often associated with 1:1 clays (e.g., kaolinite)*
ss	A, B, C	Presence of slickensides	*May occur in seasonally humid environments. Seasonal water stress occurring in smectite-dominated soils; Strong contrast in water input through drought or flooding is frequently observed*
g	A, E, B, C	Strong gleying and Munsell Chroma of $\leq$2	*Fe reduction and loss due to saturation and anaerobic conditions*
r	C	Weathered or soft bedrock	N/A
m	B	Root-restrictive, pedogenically cemented horizon with >90% carbonate, Fe, silica, gypsum, or other salts	*May occur in arid or semi-arid regions*
u	Any	Presences of artifacts	*Evidence of hominins*
b	A, E, B	Buried horizon that exhibits past pedogenesis	*If the succession of measured paleosols includes the surface soil, then denoting buried soil with "b" is useful. However, many paleosol successions consist entirely of buried, fossilized, soils; thus, the "b" is not useful in this scenario and typically excluded*
^	Any	Horizons that are composed of human-transported material	*Evidence of hominins*

Table 9.3 Classification of soil peds or aggregates. Modified from Retallack (2001) and Brady and Weil (2008). Examples are shown in Fig. 9.3

Ped type	Description	Usual Horizon	Paleoecological Implications
Platy	Tabular and horizontal to land surface	Common in C, but may occur in any horizon	*Weathering of relict bedding, soil compaction, or impeded water flow*
Prismatic	Vertically oriented peds with flat, angular top	Usually found in B horizons	*Commonly occur in soils with good drainage and can be armored with translocated clays in humid environments. Can occur near the surface in arid and semi-arid environments*
Columnar	Vertically oriented peds with rounded top	Usually found in B horizons	*Commonly occur in soils affected by shrinking and swelling clays and high in sodium in arid and semi-arid environments*
Angular Blocky	Irregular and roughly cube-shaped with sharp edges	Usually found in B horizons, but may occur in A horizons	*Form as a result of rooting, burrowing, and wetting and drying of the soil*
Subangular Blocky	Irregular and roughly cube-shaped with rounded edges	Usually found in B horizons, but may occur in A horizons	*Form as a result of rooting, burrowing, and wetting and drying of the soil*
Granular	Spheroidal in shape with slightly interlocking edges, porous	Common in A horizons	*Common in grassland soils due to earthworms, high in organic matter, poorly preserved in the rock record due to erosion*
Crumb	Spheroidal in shape, very porous	Common in A horizons	*Common in grassland soils due to earthworms, high in organic matter, poorly preserved in the rock record due to erosion and oxidation of organic matter during lithification of paleosol*
Wedge	Triangular in shape	Common in Bss horizons	*Commonly occur in Vertisols affected by shrinking and swelling clays*

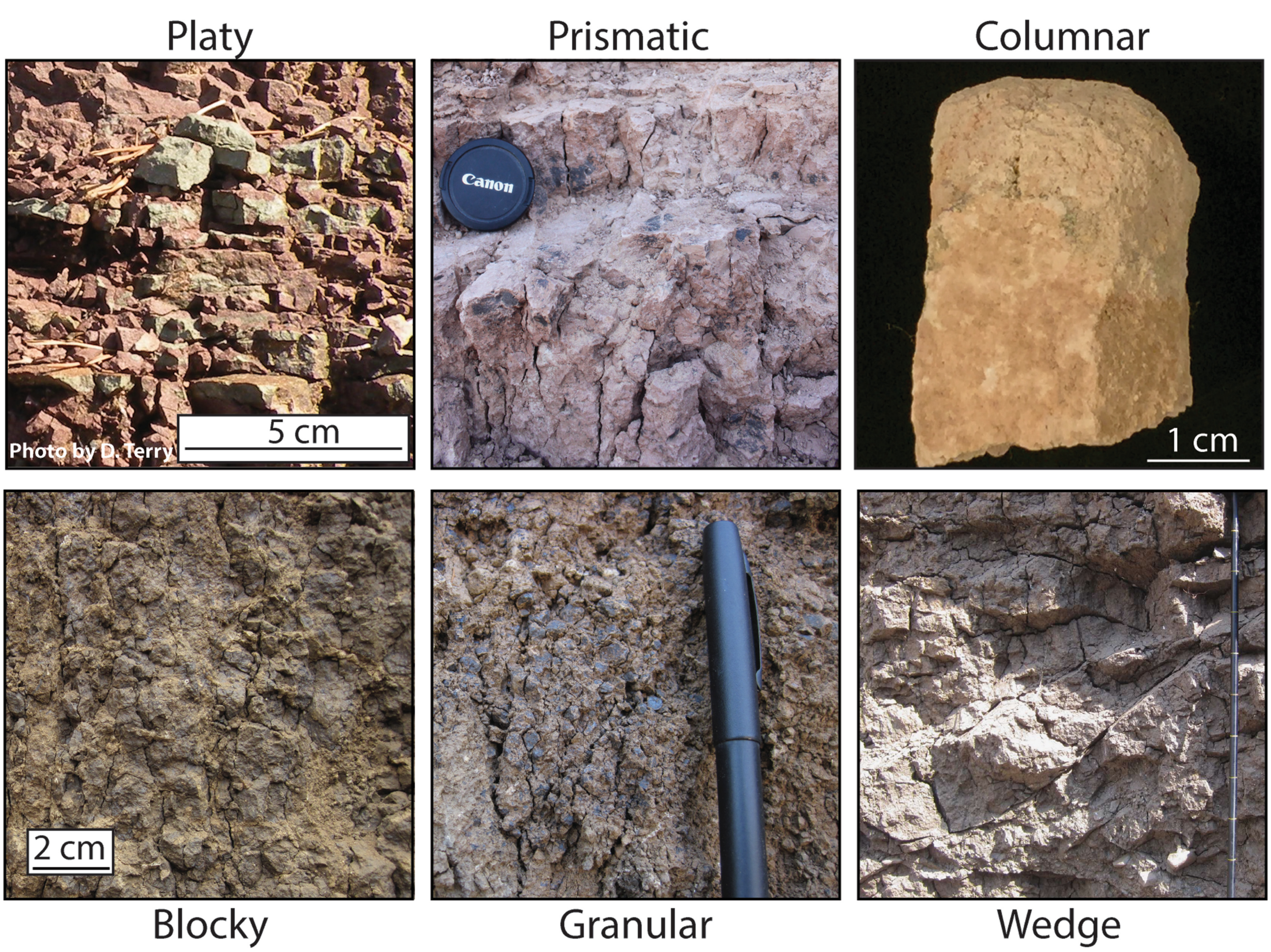

Fig. 9.3 Photographs of common peds (soil aggregates) identified in paleosols. See Table 9.3 for explanation of these ped types

Table 9.4 Paleoecological implications of soil color. Modified from Schwertmann (1993) and Holliday (2004)

Iron oxide	Munsell color	Soils	Paleoecological Implications
Goethite (FeOOH)	7.5YR-2.5Y	All soils with Fe released	*Wherever weathering takes place*
Hematite (Fe_2O_3)	7.5R-5YR	Aerobic soils of the tropics and subtropics with dry seasons	*High soil temperature, low water activity, rapid biomass turnover, high Fe release rate from rocks*
Lepidocrocite (FeOOH)	5YR-7.5YR (Value ≥ 6)	Aquic subgroups in temperate regions (Pseudogleys)	*Anaerobic/aerobic systems, noncalcareous*
Ferrihydrite ($Fe_5HO_8 \cdot 4H_2O$)	5YR-7.5YR (Value ≤ 6)	Gleys, Podzolic B horizons	*Rapid oxidation in humic environments*
Maghemite (Fe_2O_3)	2.5YR-5YR	Mainly tropical and subtropical soils	*Usually a product of fires*

Tabor et al. 2007; Rosenau et al. 2013; Ashley et al. 2013, 2014; Beverly et al. 2014; Driese and Ashley 2016).

Red or reddish-brown horizons usually indicate well-drained conditions or higher temperature soil-forming environments (Fig. 9.2A–C; Ashley and Driese 2000; Driese et al. 2016). Well-drained conditions promote Fe oxidation in O_2-rich soils, which leads to the development of goethite and hematite that impart a reddish-brown to red color in the soil (Fig. 9.2A–C; Schwertmann 1993; Birkeland 1999; Holliday 2004). Increasing temperatures can facilitate chemical reactions, and this is thought to influence the reddening of the soil. Fire-altered soil often shows reddening (Schwertmann 1993; Mentzer 2014) like those affected by heating from lavas. However, careful observation of pedogenic features and geochemical trends can disentangle the competing roles of pedogenesis versus thermal alteration in causing soil redness. For example, Sheldon (2003) showed evidence that rubified paleosols between Columbia River basalt flows appear red primarily due to oxidation and clay production during pedogenesis rather than thermal alteration. Paleosols can also inherit their color from parent materials, so care must be taken when interpreting the paleoclimatic or paleoenvironmental significance of red paleosols (Gile 1979; Sheldon 2005). Finally, Maxbauer et al. (2016b) and Clyde et al. (2013) recently showed that oxidation has altered paleosol color to depths of 20 to 30 m below the modern outcrop surface in core from the Bighorn Basin Coring Project. In the unweathered portion of the core, the colors are much more muted in comparison to the weathered paleosols in outcrop and in the upper 25 m of the core.

Sampling of paleosol horizons: After paleosol horizons are identified and their thicknesses and properties have been described, bulk soil, sediment, or paleosol samples should be collected by horizon. It is important to collect "unweathered" rock sample and avoid modern fractures and weathering (see Maxbauer et al. 2016b). Soil micromorphology (see below) can be used to identify pedogenic processes, evidence of overprinting, and mineralogical composition. Undisturbed, oriented thin-section samples should be collected from all representative horizons, air-dried, epoxy impregnated, and then prepared for in-house or commercial fabrication. In unlithified paleosols, epoxy impregnation may be necessary in the field for removal of oriented samples because of their unconsolidated state.

Bulk samples can also be collected using an arbitrary sampling interval (e.g., every 10 cm), depending on the research question(s) that are being addressed. The bulk samples can be analyzed for their particle size (depending on degree of lithification), bulk and isotopic geochemistry, and bulk and clay mineralogy. Samples of unique features should also be collected if time and space permit, e.g., root traces, mineralizations, diagnostic ped/void features (see below), organic mats, and trace fossils. Soil trace fossils can be used as a complementary data set to paleosol analysis. For further discussion of ichnology see Hasiotis (2007), Hasiotis et al. (2007a, b), and Hembree et al. (2014).

Bulk samples for geochemical analysis should be collected from uppermost B horizons (ideally 25–100 cm depth), as required for some paleoclimate reconstructions (Sheldon et al. 2002; Nordt and Driese 2010a, b; Stinchcomb et al. 2016). Bulk density of paleosols is required for constitutive characterization (e.g., mass-balance geochemistry; see section on Constitutive Mass-Balance). The bulk density of soil and sediment samples can be calculated using the clod method on peds or core method if unconsolidated (Blake and Hartge 1986). If this is not possible, bulk density can be estimated using pedotransfer functions based on the organic content and texture (Rawls 1983) or SiO_2 concentration in Vertisols (Nordt and Driese 2010b).

Micromorphological Characterization

Soil micromorphological features can be studied using a polarized light microscope and described using a variety of terms developed for micromorphological observation of soils. Brewer (1976) is the seminal work regarding the theoretical framework and application of soil micromorphology for interpreting soil development. Later, simpler terminology was developed by Bullock et al. (1985), Fitzpatrick (1993), and

Stoops (2003). Micromorphological terminology can be extensively subdivided and modified depending on the data needed to answer specific research questions. An analysis of the microstructure often begins with a description of the mineral soil matrix by describing the coarse to fine ratio (c/f) and oriented birefringent fabric (b-fabric). A description of weathering of primary minerals can be useful for determining the degree of pedogenic development (Bullock et al. 1985). B-fabric within the soil matrix is often described and can be indicative of the paleo-hydrology of the system. For further discussion of b-fabrics see Stoops (2003) and Stoops et al. (2010). Pedogenic slickensides can be identified from illuviated clay by a lack of laminations and an identical grain size between the matrix and oriented clay (Fig. 9.2G). Oriented b-fabric may indicate that repetitive wetting and drying caused clays to shrink and swell, leading to inclined orientation due to the confining pressure of overlying soil (Blokhuis et al. 1990). It could also indicate bioturbation and reorganization of the soil matrix by other means creating grano-, channel, or poro-striated b-fabric (Stoops et al. 2010).

Pedofeatures, which are >20 μm and are distinguishable from the soil matrix, should also be described in detail (Stoops 2003). In thin section, burrows and root voids can be distinguished by analyzing the textures around the void (Stoops 2003; Stoops et al. 2010). Coatings (lining voids), hypocoatings (impregnated into matrix around pore), quasicoatings (impregnated into matrix surrounding pore but at some distance from feature), and infillings are also important to describe. A variety of materials such as Fe–Mn (Fig. 9.2E), illuviated clay or silt (Fig. 9.2H), or carbonate can fill these voids. Both silt and clay can be translocated down profile via the vertical drainage of water (Stoops et al. 2010) and indicate that the water table was low enough for drainage, but that there was enough water to translocate clay or silt. The illuviation of silt in thin section is sometimes used to infer the physical translocation of freshly deposited silt after a flood or dust accumulation (Kubiena 1970; Fedoroff and Goldberg 1982; Kemp 1999).

The accumulation of clay coatings observed in thin section has been related to the residence time (duration of weathering in the system) of soils using a chronosequence of deeply weathered, sub-tropical soils from Mississippi (Ufnar 2007). It is possible that the Ufnar (2007) chronofunction may overestimate the length of pedogenesis in certain climates (i.e., monsoonal), but this technique can be a useful approximation (Beverly et al. 2015a). Although clay illuviation has been inferred to represent warm-wet climate conditions in some studies (e.g., Bronger and Heinkele 1989), later work showed that Bt horizons can form in semi-arid grassland landscapes when enough time has elapsed (Han et al. 1998). A corollary of the Han et al. (1998) findings is that clay coatings could result from windblown dust deposits weathered and translocated down from silty A horizons.

Fe–Mn concentrations and depletions can be used to infer drainage (Fitzpatrick 1984; Bullock et al. 1985) and more specifically to infer the relative influence of epi- versus endo-saturation in the paleosol. Fe oxides coating peds with reduced interiors are indicative of endosaturation, where water is wicked upward through micropores within aggregate interiors. Oxidizing conditions remain in macropores, where the spacing between grains is too great for water to adhere. Reduction present along the exterior of peds that contain oxidized interiors suggests periodic episaturation (Fitzpatrick 1984; Bullock et al. 1985). Water saturation from surficial ponding will fill macropores and inhibit the diffusion of atmospheric oxygen into pore waters. Consumption of oxygen by microbes will result in the reduction of Fe on aggregate exteriors, which will penetrate into aggregate interiors with time. If the soil becomes drained of ponded water before the anoxic pore waters completely infiltrate aggregate interiors, Fe within aggregates will remain oxidized (Fitzpatrick 1984; Bullock et al. 1985).

Nodules are three-dimensional features not associated with a void or pore feature and are important as they can reveal information about calcification or redox conditions (Stoops 2003). The most commonly identified nodules are pedogenic carbonate, but Fe–Mn nodules are also important (Stiles et al. 2001). The accumulation of $CaCO_3$ in soils frequently indicates an arid or semi-arid soil-forming environment (Gile et al. 1966; Machete 1985). Cathodoluminescence (CL) is also a useful tool for determining the redoximorphic environment of carbonate identified in a paleosol, assessing diagenetic alteration (Barnaby and Rimstidt 1989; Machel et al. 1991), or identifying parent material contamination by marine carbonates (Michel et al. 2013). Carbonate forming in a shallow reducing environment with higher manganese to iron ratios will luminesce; whereas, carbonate forming in an oxidizing environment will have no luminescence (Barnaby and Rimstidt 1989; Machel et al. 1991). The importance of pedogenic carbonate is further discussed below.

An ultraviolet fluorescence (UVf) microscope attachment causes organic matter to fluoresce, facilitating the identification of organics that would be otherwise unobservable. UVf can reveal features such as root cell structure not visible in cross-polarized or plane light (Fig. 9.2K; Beverly et al. 2014, 2015b). Thin sections are often point counted to quantify mineralogy, organic content, or other morphological features that elucidate pedogenic processes (e.g., Stinchcomb et al. 2014).

Identifying features created by soil fauna often yields critical paleoecological information. Faunal features include: fecal pellets, channels or burrows, chambers, coatings, and infillings (Stoops 2003). Faunal features are not always readily identifiable but can yield significant information about paleohydrology. Termites, for example, do not live below the water table, and thus the depth of their burrowing can indicate the height of the water table associated with their colonization surface. Soil-dwelling crayfish can be of similar use in identifying the water table because they require standing water at the base of their burrows (Hasiotis and Honey 2000). Earthworm fecal pellets and burrows are often identified in paleosols and provide macropores favorable to root growth and microbial activity and stabilize organic carbon in the soil (Fig. 9.2L; Stoops et al. 2010). For further discussion of faunal features in micromorphology see Ch. 18 of Stoops et al. (2010).

Characterizing Pedogenic Minerals

Bulk mineralogy of soils and paleosols can be characterized using X-ray diffraction (XRD) (Poppe et al. 2001). Common pedogenic minerals and their paleoecological implications are shown in Table 9.5 (Bullock et al. 1985). Harris and White (2008) provide a summary of how to use XRD to identify common minerals in soils.

Clay Mineralogical Characterization: Clay mineralogical analysis is a powerful tool for recognizing weathering products versus inherited materials, which ultimately aids in understanding the weathering pathway(s) that occurred in paleosols (e.g., Tabor and Montañez 2002; Rosenau et al. 2013). The composition of the clay-sized fraction (<2 μm) is analyzed using XRD and should be prepared following standard methods (Moore and Reynolds 1997). When preparing samples for clay mineralogy, multiple aliquots may be needed to help diagnose 2:1 clay minerals using the following treatments: (1) air-dry, (2) K saturation, (3) Mg saturation by ethylene glycol solvation, and (4) subsequent 550°C heat treatment of the K-saturated sample. Tabor and Myers (2015) address the characterization and interpretation of clay minerals for paleoclimatic purposes.

Pedogenic Carbonate: Pedogenic calcium carbonates ($CaCO_3$ – limestone; $(Ca,Mg)(CO_3)_2$) – dolomite) are an important paleoecological indicator as they typically form in dry environments where precipitation is exceeded by evapotranspiration (Fig. 9.2K), though carbonate precipitation can occur in some poorly-drained alluvial back-swamp soils (Aslan and Autin 1998; Mintz et al. 2011). Carbonates initially precipitate in the soil matrix as soft masses with diffuse boundaries and micritic crystalline texture (Nordt et al. 2004). Zones of carbonate can develop septarian shrinkage cracks (Fig. 9.2J), multi-generational carbonate overgrowths, and coatings of other minerals, such as Fe and Mn oxides (Bullock et al. 1985; Stoops 2003). Such alteration converts the soft masses to hard nodules, which has been interpreted to occur outside the zone of active carbonate precipitation (Nordt et al. 2004). Carbonates also precipitate locally around roots, where CO_2 is high due to respiration, forming filaments and nodules that converge to form rhizoliths. If vertic soil processes caused by shrinking and swelling of clays during the wet and dry seasons are present, these rhizoliths can be broken up to form rhizocretions (Fig. 9.2I).

The degree to which carbonates precipitate in the soil matrix is primarily a function of time (provided that the other soil forming factors are held constant) and can be described

Table 9.5 Common pedogenic minerals and their paleocological implications. Modified from Bullock et al. (1985)

Mineral	Composition	Potential paleoecological implications
Pyrite	FeS_2	*Potentially acid sulphate soils (i.e., waterlogged conditions)*
Hematite	FeS_2O_3	*Potentially tropical or subtropical soils*
Lepidocrocite	$FeO(OH)$	*Gleyed soils (i.e., waterlogged conditions)*
Goethite	$FeO(OH)$	*Temperate and tropical soils*
Gibbsite	$Al(OH)_3$	*Tropical or subtropical soils, Andisols*
Halite	$NaCl$	*Saline soils*
Calcite	$CaCO_3$	*Arid to semi-arid soils or from calcareous parent material*
Siderite	$FeCO_3$	*Marshy soils*
Natron	$Na_2O_3*10H_2O$	*Saline soils*
Trona	$Na_3H(CO_3)_2*2H_2O$	*Saline soils*
Thenardite	Na_2SO_4	*Saline soils*
Barite	$BaSO_4$	*Soils with saline groundwater*
Celestite	$SrSO_4$	*Gypsiferous soils*
Anhydrite	$CaSO_4$	*Arid soils*
Gypsum	$CaSO_4*2H_2O$	*Semi-arid soils; acid sulphate soils*
Jarosite	$KFe_3(OH)_6(SO_4)_2$	*Acid sulphate soils (i.e., waterlogged conditions)*
Mirabilite	$Na_2SO_4*10H_2O$	*Saline soils*
Vivianite	$Fe_3(PO_4)_2*8H_2O$	*Marshy soils*
Bassanite	$CaSO_4*1/2H_2O$	*Gypsiferous soils*

Table 9.6 Stages of calcium formation in calcic soils and pedogenic calcretes. Modified from Gile et al. (1966) and Machete (1985)

Type	Stage	Gravel content	Diagnostic morphology	$CaCO_3$ distribution	Max. $CaCO_3$ content
Calcic soil	I	High	Thin, discontinuous coatings on pebbles, usually on undersides	Coatings sparse to common	tr – 2
		Low	A few filaments in soil or faint coatings on ped faces	Filaments sparse to common	tr – 4
	II	High	Continuous, thin to thick coatings on tops and undersides of pebbles	Coatings common, some carbonate in matrix, but matrix still loose	2–10
		Low	Nodules, soft, 0.5 to 4 cm in diameter	Nodules common, matrix generally noncalcareous to slightly calcareous	4–20
	III	High	Massive accumulations between clasts, becomes cemented in advanced form	Essentially continuous dispersion in matrix	10–25
		Low	Many coalesced nodules, matrix is firmly to moderately cemented		20–60
Pedogenic calcrete (indurated calcic soil)	IV	Any	Thin (<0.2 cm) to moderately thick (1 cm) laminae in upper part of K horizon. Thin laminae may drape over fractured surfaces	Cemented platy to weak tabular structure and indurated laminae. K horizon is 0.5 – 1 m thick	>25 in high gravel content; >60 in low gravel content
	V	Any	Thick laminae (>1 cm) and thin to thick pisolites. Vertical faces and fractures are coated with laminated carbonate (case-hardened surfaces)	Indurated dense, strong platy to tabular structure. K horizon is 1–2 m thick	>50 in high gravel content; >75 in low gravel content
	VI	Any	Multiple generations of laminae, breccia, and pisolites; cemented. Many case-hardened surfaces	Indurated and dense, thick strong tabular structure. K horizon is commonly >2 m thick	>75 in all gravel contents

High: >50% gravel content; Low: <20% gravel content; Tr – trace

in stages of accumulation (Table 9.6; Gile et al. 1966; Machete 1985). Soil carbonates can also be inherited from the parent material, particularly when materials are sourced from marine limestone (Michel et al. 2013). When describing carbonates in the field, an effervescence test is useful for describing the degree of effervescence in the paleosol matrix and/or in secondary carbonate features, which may indicate the relative concentration of carbonate in the paleosol horizon (Schoeneberger et al. 2012).

Stable isotopes of carbon and oxygen from pedogenic carbonates have been used extensively to interpret paleoecology (Cerling and Hay 1986; Cerling et al. 1988), where long sedimentary records have been used to understand the changing climate and environment of the past several million years. Carbon isotopes in pedogenic carbonates are of particular importance because they can be used to interpret open and closed environments by using the enrichment or depletion of ^{13}C relative to ^{12}C, which can result from changes in biomass utilizing disparate photosynthetic pathways (C_3 vs C_4) or from water stress in C_3 plants (Cerling and Quade 1993). Pedogenic carbonates should be collected from at least 50 cm depth to ensure that the soil CO_2 reflects the ^{13}C composition of respired organic C in soil CO_2 rather than infiltrated atmospheric CO_2 (Cerling and Quade 1993). A discussion of biases associated with pedogenic carbonate can be found in **Biases and Shortcomings**.

Pedogenic Siderite: Siderite ($Fe(II)CO_3$) crystallizes in the reducing phreatic waters of poorly drained soils, commonly around the roots of hydrophilic vegetation (Ludvigson et al. 1998, 2013). Spherulitic masses of siderite are detectable in thin-section as typically mm-scale crystals with radial extinction. In addition to its use as a drainage and redox indicator, siderite has been used to reconstruct the $\delta^{18}O$ composition of ancient meteoric water (Ludvigson et al. 2013, and references therein).

Geochemical Characterization

In contrast to mineralogical characterization, which reveals the organization of elements in soils, geochemical analysis measures the concentrations of elements in a sample. Paleosols should be geochemically analyzed to determine parent material uniformity, the presence and magnitude of pedogenic processes (Birkeland 1999), and to estimate paleoenvironments or paleoclimate variables (Sheldon and Tabor

2009; Tabor and Myers 2015). Bulk geochemical samples should be prepared in a similar fashion to modern soils. Samples are air-dried, broken up with a mortar and pestle, and macroscopic minerals, such as mineral nodules or salts, are discarded so that only the soil matrix is analyzed (Soil Survey Staff 2014b). Treatment with HCl is not necessary and will bias quantitative geochemical analysis if Ca^{2+} in the matrix is removed with carbonate (Myers et al. 2014; Tabor and Myers 2015). Samples are then powdered (typically with >85% of sample finer than 100 μm) and analyzed using inductively-coupled plasma (ICP) – mass spectrometry (MS), – optical emission spectrometry (OES), – atomic emission spectrometry (AES), or X-ray fluorescence (XRF). The results are usually reported in elemental or oxide weight percent. Several applications of geochemical characterization are described here, but for a more thorough review, see Sheldon and Tabor (2009) and Tabor and Myers (2015).

Molecular Weathering Ratios: Molecular weathering ratios can be used as a qualitative approximation to understand weathering in the paleosol and to evaluate relative changes down profile (Retallack 2001), but more quantitative methods such as constitutive mass-balance are now preferred by paleopedologists (see below). Oxides are by convention converted from weight percent to molar percent (dividing by molecular weight of each oxide) before calculating molecular weathering ratios. A discussion of the inferences that can be drawn from molecular weathering ratios can be found in Retallack (2001). The chemical index of alteration (CIA) and chemical index of alteration minus potassium (CIA-K) can be used as a proxy for weathering intensity in a paleosol (Table 9.7; Nesbitt and Young 1982; Maynard 1992; Sheldon et al. 2002). These weathering indices measure clay formation and base loss associated with feldspar weathering and base retention and carbonate precipitation in arid environments (Maynard 1992; Sheldon et al. 2002; Lukens et al. 2018). CIA-K is the preferred weathering ratio for pre-Quaternary paleosols as it reduces digenetic effects of potassium metasomatism (Maynard 1992; Sheldon et al. 2002).

Constitutive Mass-Balance: Mass-balance is a more powerful tool than molecular weathering ratios because it quantitatively compares geochemical and volumetric changes in soil horizons relative to their parent material (Brimhall and Dietrich 1987; Brimhall et al. 1991a, b). Due to variability in parent material compositions, the first step in the mass-balance approach is to evaluate parent material

Table 9.7 Common MAP proxies

Proxy	Soil Orders	Equation for MAP (mm yr^{-1})	Error (mm yr^{-1})	Range (mm yr^{-1})	Reference
Depth to Bk	Any soil with a Bk horizon	$MAP = 137.24 + 6.45D - 0.013D^2$ D = depth to Bk (cm)	±147	≤ 1000	Retallack (2005)
Depth to Bk	Only Vertisol microlows	$MAP = 2.81 \times 10^{-4}D^2 - 0.4025D + 217.35$ (1%) $MAP = 1.41 \times 10^{-4}D^2 - 0.1255D + 109.58$ (2%) $MAP = 6.00 \times 10^{-5}D^2 - 0.0328D + 40.16$ (5%) D = depth to Bk (cm)	±420 ±230 ±230	700–1400	Nordt et al. (2006)
Depth to Bk	Only Vertisol microhighs	$MAP = 4.42 \times 10^{-4}D^2 - 0.9437D + 51252$ (1%) $MAP = 9.48 \times 10^{-4}D^2 - 2.0687D + 1141.12$ (2%) $MAP = 1.52 \times 10^{-3}D^2 - 3.35D + 1853.2$ (5%) D = depth to Bk (cm)	±480 ±490 ±370	700–1400	Nordt et al. (2006)
Depth to By	Any soil with gypsic horizon	$MAP = 87.593e^{0.0209D}$ D = depth to By (cm)	±129	≤ 1000	Retallack and Huang (2010)
CIA-K	Any soil without gleying or with ≤ 5 wt% carbonate as well as Oxisols, Entisols, Andisols, and Vertisols	$MAP = 14.3x - 37.6$ $x = Al_2O_3/(Al_2O_3 + CaO + Na_2O) \times 100$	±172	200–1600	Maynard (1992), Sheldon et al. (2002a), Tabor and Myers (2015)
Bases/ Alumina	Same as CIA-K	$MAP = -259.3 \ln(x) + 759$ $x = (CaO + MgO + Na_2O + K_2O)/Al_2O_3$		200–1600	Sheldon et al. (2002)
CALMAG	Only Vertisols	$MAP = 22.69x - 435.8$ $x = Al_2O_3/(Al_2O_3 + CaO + MgO)$	±108	200–1400	Nordt and Driese (2010a, b)
Goethite/ Hematite	All except Vertisols, Andisols, Entisols, and Gelisols	$MAP = 1003.5[G/H] + 49.3$ G/H = goethite-hematite ratio	±157	100–3300	Hyland et al. (2015)
Fe_{TOT}	Only Vertisols	$MAP = 31.52Fe_{TOT} + 654.38$ Fe_{TOT} = total Fe content of FeMn nodules (wt%)		850–1300	Stiles et al. (2001)

composition using immobile and chemically inert major, trace, and rare earth elements (e.g., Ti, Zr, Al, Nb, La, Ce). The most commonly used immobile constituents are Ti, Zr, and Al, as the deviation from the parent material tends to be relatively small. The concentration of other elements (e.g., Na, Mg, K, Ca, Si, Fe(II), Mn(II), C, P) is weighed against the immobile element to elucidate additions and/or losses relative to the parent material. For these methods, selection of an appropriate, homogeneous, and unweathered parent material is best, although some studies have found that using the lowermost, least-weathered horizon for each paleosol provides acceptable results (e.g., Driese et al. 2016). If the parent material is approximated, it is best to discuss the mass-balance results in a conservative manner (e.g., minimum apparent loss or gain of an element).

Vertical changes in grain size in soil profiles are common and induce heterogeneity in starting mineralogy, which can influence the results of mass balance calculations. Likewise, additions of sedimentary material during pedogenesis can produce variability in weathering trends toward the top of profiles. To account for this issue, parent material uniformity can be examined using geochemical reconstructions applied to buried soils and paleosols following Maynard (1992). The ratio of TiO_2/Zr for each candidate parent material and the ratio TiO_2/Zr of the average composition of the paleosol can be used to determine percent deviation from the parent material (Maynard 1992):

$$\text{Percent Deviation} = \left(\frac{\text{Ti}_{\text{soil}}}{\text{Zr}_{\text{soil}}} - \frac{\text{Ti}_{\text{parent}}}{\text{Zr}_{\text{parent}}}\right) \Big/ \left(\frac{\text{Ti}_{\text{parent}}}{\text{Zr}_{\text{parent}}}\right) \times 100 \tag{2}$$

Cross plots of immobile elements are also useful for determining parent material such as Zr/(Zr/TiO_2) vs. TiO_2/(TiO_2/Zr) (e.g. Ashley and Driese 2000; Driese et al. 2000; Beverly et al. 2014). The selection of Ti and Zr as immobile index elements has also been discussed by Stiles et al. (2003a, b), who found that Zr can be introduced as wind-blown silt in semi-arid environments.

The following calculations for constitutive mass-balance analysis are derived from Brimhall and Dietrich (1987), Brimhall et al. (1991a, 1991b), Chadwick et al. (1991) and Anderson et al. (2002). Closed system weathering conserves mass and volume, such that:

$$\mathrm{V_p \rho_p C_{i,p} = V_w \rho_w C_{i,w}} \tag{3}$$

where the volume (V), bulk density (ρ) and concentration (C) of an immobile element (i) in the parent material (p) are equal to those of a weathered horizon (w).

In most soils, weathering is an open system process, wherein constituents are added or removed from the system. In such cases, volumetric change is tracked as strain (ε):

$$\varepsilon_{i,w} = \frac{\rho_p C_{i,p}}{\rho_w C_{i,w}} - 1 \tag{4}$$

Note that if losses of an element in a weathered horizon are offset by changes in volume, strain is zero. Elemental additions and losses are calculated using the open-system mass-transport function (τ), which incorporates volumetric changes:

$$\tau_{j,w} = \frac{\rho_w C_{j,w}}{\rho_p C_{j,p}}\left(\varepsilon_{i,w} + 1\right) - 1 \tag{5}$$

where $C_{j,w}$ is the concentration of an element of interest (j) in weathered material and $C_{j,p}$ is the concentration of an element of interest (j) in the parent material. When $\tau_{j,w} = -1$, 100% of the element j has been removed in a weathered horizon compared the parent material. Likewise, when $\tau_{j,w} = 1$, 100% of element j has been added to a weathered horizon compared the parent material starting composition. Definitions of all variables in the above equations are summarized in Table 9.8.

Interpretation of mass-balance calculations should be formed in terms of soil-forming process, which may vary depending on the environmental setting or parent material type. Translocation of carbonate is identified by simultaneous changes in Ca, Mg, and Sr. Clay or zeolite accumulation is indicated by increases in Na, Al, Si, and Mg (Ashley and Driese 2000; Beverly et al. 2014). In contrast, Mg fluxes have also been found to closely track clay-rich, Mg-smectite parent material rather than carbonate (Beverly et al. 2014). Losses of the redox-sensitive elements (e.g., V, Fe, Cu, and Mn) may indicate prolonged waterlogging followed by drainage. Cu is also important in the formation organic ligands and can be indicative of changes in organic matter content (Brantley et al. 2007). Likewise, phosphorus

Table 9.8 Glossary of variables modified from Brimhall and Dietrich (1987)

$C_{i,w}$	= concentration of immobile element (i) in weathered material
$C_{i,p}$	= concentration of immobile element (i) in parent material
ρ_p	= bulk density of parent material
ρ_w	= bulk density of weathered material
$\varepsilon_{i,w}$	= strain or volumetric change of immobile element (i) in weathered material
$\tau_{j,w}$	= translocation of element of interest (j) in weathered material
$C_{j,w}$	= concentration of element of interest (j) in weathered material
$C_{j,p}$	= concentration of element of interest (j) in parent material

enrichment at the soil surface can indicate biocycling (Brantley et al. 2007). Human remains and debris can also be enriched in P relative to surrounding material, but additional evidence would be needed to support this interpretation (Eidt 1985; Holliday 2004; Holliday and Gartner 2007).

Pedotransfer Functions: Pedotransfer functions relate bulk geochemistry of soils to properties that are not directly measureable in paleosols. For example, modern soils are studied using a suite of standard characterization properties, which are measured on unconsolidated material to assess physical and chemical properties operating at the colloidal scale (Soil Survey Staff 2014b). Lithification and compaction of paleosols precludes such measurements, but the relationship between characterization properties and geochemical indices allows for detailed pedologic study of paleosols (Nordt and Driese 2010b). Soil characterization properties can yield quantitative paleoecological information on paleosol fertility and hydrology unavailable by any other means (Table 9.9; Nordt et al. 2011). A detailed explanation of modern soil properties can be found in Soil Survey Staff (2014b). Currently, most pedotransfer functions are restricted to paleo-Vertisols (Nordt and Driese 2010b), which are commonly found in the geologic record (Driese and Forman 1992; Driese et al. 1993; Prochnow et al. 2006; Cleveland et al. 2007; Rosenau et al. 2013; Torres and Gaines 2013; Driese et al. 2016). Efforts to develop more broadly applicable pedotransfer functions for soil characterization variables are ongoing and currently limited to soil pH (Lukens et al. 2018). If a paleosol is unconsolidated, soil properties could theoretically be directly measured using established methods (Nordt et al. 2013; Soil Survey Staff 2014b). However, it is not clear to what extent characterization properties are affected in buried paleosols that have not experienced macroscopic evidence of diagenesis.

Table 9.9 Pedotransfer functions for paleo-Vertisols modified from Nordt and Driese (2010b)

Soil Property	Soil Horizon	Equation	Error	Paleoecological importance
Total clay (%)	Noncalcareous	$y = 3.965x + 3.681$ $x = Al_2O_3$ (%)	±4	*Porosity and water holding capacity*
	All ≤5% CaO	$y = 3.118x + 15.22$ $x = Al_2O_3$ (%)	±5	
Fine clay (%)	All ≤10% CaO	$y = 9.974 \ln(x) + 23.68$ $x = Al_2O_3/(CaO + MgO)$*	±7	*Translocation of pedogenic clay*
COLE Coefficient of linear extensibility ($cm\ cm^{-1}$)	Noncalcareous	$y = 1.345x + 0.077$ $x = CaO + MgO$*	±0.019	*Shrink-swell behavior*
	All ≤10% CaO	$y = 0.00238x - 0.02965$ $x = SiO_2$ (%)	±0.025	
Bulk density ($g\ cm^{-3}$)	Noncalcareous	$y = 0.011x + 0.457$ $x = SiO_2$ (%)	±0.10	*Porosity and aeration of soil*
CEC_7 ($cmol_c\ kg^{-1}$) Cation exchange capacity	Noncalcareous	$y = 585.8x + 14.66$ $x = CaO + MgO$*	±4	*Soil fertility (plant available nutrients)*
	All ≤10% CaO	$y = -8.203x + 44$ $x = (CaO + MgO)/Al_2O_3$*	±8	
pH (H_2O)	All	$y = 0.87 \ln(x) + 9.648$ $x = CaO + MgO$*	±0.6	*Soil fertility (acidity)*
BS Base saturation (%)	All	$y = 14.69 \ln(x) + 134.0$ $x = CaO + MgO$*	±8	*Soil fertility (alkalinity)*
$CaCO_3$ (%)	All	$y = 1.436x + 0.025$ $x = CaO$ (%)	±4	*Buffers soil pH*
Fe_d (%) Crystalline Iron Oxide	All	$y = 0.085x^2 - 0.352x + 0.658$ $x = Fe_2O_3$ (%)	±0.4	*Oxidizing conditions*
Fe_o (%) Amorphous Iron Oxide	All	$y = 0.018e^{570.3x}$ $x = Ni$ (%)	± 0.05	*Poorly drained conditions*
ESP (%) Exchangeable Sodium Percentage	All ≤10% CaO	$y = 110.4x - 2.935$ $x = Na_2O/Al_2O_3$*	±4	*Soil fertility (sodicity)*
EC ($dS\ m^{-1}$) Electrical conductivity	All ≤10% CaO	$y = 290.7x^2 - 24.33x + 1.145$ $x = Na_2O/Al_2O_3$*	±3	*Soil fertility (salinity)*

% calculated using weight percent; *converted to molar before calculation

Mean Annual Precipitation and Temperature Proxies

Mean annual precipitation (MAP) can be estimated from paleosols by either physical or geochemical techniques (Table 9.7). The depth to calcic horizon (Bk) and depth to gypsic horizon (By) in soil profiles increases as a function of increasing MAP (Retallack 2005; Nordt et al. 2006; Retallack and Huang 2010). Nordt et al. (2006) used a "family of curves" approach to derive significant relationships between depth to nodular Bk horizons based on the concentration of nodules (1, 2, and 5%) for each of the microhigh and microlow portion of Vertisol profiles. The limitations for applying these depth-to-mineral functions are: (1) the tops of paleosols are commonly eroded (Myers et al. 2014), which results in MAP underestimates, (2) carbonate and gypsum must be nodular in morphology, (3) carbonate and gypsum must be analyzed in thin section to demonstrate pedogenic origin, and (4) burial compaction must be accounted for (Sheldon and Retallack 2001).

The bulk geochemistry of paleosol B horizons is a more reliable method of reconstructing MAP because it does not require the original thickness of the paleosol to be preserved. Typically, the uppermost B horizon is used as the "control interval" (25–100 cm depth in profile; Nordt and Driese 2010a). Bulk geochemistry is related to MAP through two processes: (1) hydrolysis of silicate minerals and progressive leaching of base cations at high MAP, and (2) precipitation of carbonates and retention of base cations at low MAP. Sheldon et al. (2002) found the weathering ratio CIA-K (Sect. 4.5.1) to be the most useful predictor of MAP across a range of soil types, which increases as CaO and Na_2O are preferentially leached relative to Al_2O_3. The acid neutralization capacity of secondary carbonates precipitated at low MAP (*sensu* Chadwick and Chorover 2001) results in retention of structural CaO and Na_2O and correspondingly low CIA-K values. K_2O is excluded from CIA-K because of potential diagenetic addition of K^+ during illitization (Fedo et al. 1995; Sheldon et al. 2002; Lukens et al. 2018).

More recently, attempts have been made to decrease standard error in MAP functions by developing proxies specific to more narrow ranges of soil environments (i.e., *cl*, *o*, *r*, *p*, *t;* Sect. 3.1). CALMAG (Nordt and Driese 2010a) was developed specifically for clay-rich Vertisols, which are soils dominated by vertic features such as gilgai topography, pedogenic slickensides, and wedge peds that develop in soils that experience seasonal water deficit. Vertisols tend to be developed on pre-weathered parent material where little hydrolysis occurs, and therefore, tracking the flux of CaO and MgO (relative to Al_2O_3) as carbonates or exchangeable cations within the system results in improved MAP prediction (Nordt and Driese 2010a).

It should be noted that climofunctions using weathering indices that relate base cations to Al_2O_3 are only useful for estimating MAP up to a maximum of ~1600 mm (Sheldon et al. 2002; Nordt and Driese 2010a). Above 1600 mm, kaolinite is the dominant clay mineral and any weatherable bases are either completely depleted – yielding an effective ratio of alumina:alumina that does not change with increasing MAP – or bases have yet to be leached and the soil is not in equilibrium with climatic conditions. A few novel methods to estimate MAP utilize Fe in Fe–Mn nodules (Stiles et al. 2001) or the relative abundance of goethite to hematite (Hyland et al. 2015). The weight percent of Fe in Fe–Mn nodules found in Vertisols increases directly with MAP (Stiles et al. 2001) and can be used in conjunction with other methods. Most recently the goethite to hematite ratio (G:H) in the B horizon of a global dataset of soils is shown to correlate with MAP (Hyland et al. 2015). The ratio of G:H can be measured using either X-ray diffraction (XRD) or isothermal remnant magnetization (IRM) acquisition curves, but IRM is preferred due to the higher sensitivity in low concentrations of magnetic minerals in soils and paleosols (Hyland et al. 2015). G:H is the first MAP proxy to extend beyond the barrier of 1600 mm, with a predictive range of 100–3300 mm (Hyland et al. 2015). However, Maxbauer et al. (2016a) note that goethite is difficult to measure using IRM methods and easily recrystallizes during diagenesis, which would preclude measurements of original G:H ratios for most paleosols.

Mean annual temperature (MAT) proxies for paleosols have also improved over the last decade. Early attempts relied on bulk geochemical proxies such as the salinization index (Sheldon et al. 2002). An alternative approach to predicting paleo-MAT uses the oxygen stable isotopic composition of carbonates ($\delta^{18}O$), which is dependent on both water ($\delta^{18}O$) composition and the temperature at which carbonates precipitated (i.e., soil temperature) (Cerling 1984). Dworkin et al. (2005) found a possible way to side-step original water composition by combining two equations that relate (1) the fractionation of oxygen from meteoric water into carbonate, and (2) the spatial relationship between MAT and the $\delta^{18}O$ of meteoric water. However, it is possible that past time intervals had systematic differences in the spatial distribution of meteoric water $\delta^{18}O$ and variation in mean annual temperature, which would preclude the application of this proxy (e.g., Lukens et al. 2017a).

Recent advances in the analytical resolution of isotopes of carbon and oxygen have allowed for the measurement of clumped isotopes (Δ_{47}). In this technique, the abundance of a doubly substituted CO_2 molecule of mass 47 ($^{13}C^{18}O^{16}O$) produced when carbonate is digested during measurement is proportional to soil temperature, without assumptions based on original $\delta^{18}O$ composition of soil waters (Ghosh et al. 2006; Passey et al. 2010; Quade et al. 2013). These carbonates are probably recording the warm, dry season temperature when carbonate is more likely to form and therefore do not represent a mean annual temperature (Passey et al. 2010; Gallagher and Sheldon 2016). The Δ_{47} temperature estimates provide a unique view into critical paleoclimatic variables recovered from paleosols (e.g., Snell et al. 2013) when burial temperatures do not exceed 100°C (Henkes et al. 2014).

Classifying Paleosols

Several perspectives exist on the taxonomic classification of paleosols (Mack et al. 1993; Retallack et al. 1993; Nettleton et al. 2000), which stem from the fact that some distinctions in modern soil taxonomy rely on soil and climate properties (e.g., moisture and temperature regimes) that are immeasurable in paleosols. Each system has strengths and weaknesses (i.e., Retallack et al. 1993, comment and reply), and to date no taxonomy is uniformly accepted among paleopedologists. The two most commonly applied paleosol classification systems are Mack et al. (1993) and Retallack et al. (1993). The Mack et al. (1993) system classifies paleosols without the use of modern soil properties such as bulk density or base saturation, which are not measurable in most paleosols, and creates new names for some paleosol orders. The Retallack et al. (1993) classification system simplifies USDA Soil Taxonomy to allow for paleosol classification by omitting properties that cannot be directly measured in paleosols but retains the naming conventions of Soil Taxonomy. The authors prefer to adapt Soil Taxonomy where possible for comparisons with modern soils and to encourage collaboration with modern soil scientists. The Illustrated Guide to Soil Taxonomy (Soil Survey Staff 2014a, b) is a useful resource for those unfamiliar with the complexities of Soil Taxonomy and is practical for most paleopedological classification purposes.

Rates of Pedogenesis: Paleosols and paleosol successions can also be categorized based on the varying degrees of pedogenesis and sedimentation. Paleosols in alluvial environments are broadly split into three different categories: compound, composite, and cumulative (Fig. 9.4). Compound soils are weakly developed with vertically stacked profiles that occur when sedimentation rate is not steady and there is little erosion. In composite or welded paleosols the rate of pedogenesis exceeds that of sedimentation, and with little erosion creates vertically stacked profiles that partially overlap. Cumulative paleosols have slow but overall constant sedimentation rates where pedogenesis can modify the sediment as it is deposited (Kraus 1999). Cumulative soils tend to preserve remains, such as artifacts or fossils, in discrete depth ranges (Holliday 2004, and references therein). With erosion, these three categories can be further split into compound truncated set, composite truncated set, and cumulative truncated set (Kraus 1999). A careful study of root traces can also help classify rates of pedogenesis. Root mats can help identify surfaces within the paleosol but can also form at a lithologic boundary or fragipan (x horizon suffix; Table 9.3) preventing growth. Clay coatings, redoximorphic features, carbonate, Fe–Mn, and zeolites can all preserve roots features (Fig. 9.2E–F; Ashley and Driese 2000; Stoops et al. 2010; Beverly et al. 2014, 2015a). For further discussion of rhizoliths and their paleoecological implications see Kraus and Hasiotis (2006).

Pedofacies and Pedotypes: Paleosols in stratigraphic successions often contain similar attributes, which relate to their depositional environment. Such clustering can reveal vertical and lateral trends that are the result of geologic and climatic phenomena. Changes in paleosol maturity and drainage are typically assessed using qualitative indices (Table 9.10). Pedofacies use a combination of sedimentological and pedological features to categorize paleosols (Bown and Kraus 1987; Kraus 1987). The pedofacies approach was developed in alluvial settings and relies on the inverse relationship between deposition and soil development – grain size and flooding frequency decrease with distance from sediment source, which is accompanied by increased soil maturity and thicker individual soil profiles.

Retallack (1994) developed a more universal model of paleosol categorization, in which paleosols with similar features are grouped as pedotypes. A pedotype is a representative paleosol and reference to the type profile for each pedotype allows for a non-genetic classification of each paleosol and is indicative of one set of soil forming (*cl, o, r, p, t*) conditions (Retallack 1994). Analysis of pedofacies and pedotypes in vertical stratigraphic successions can reveal long-term trends in climate or basin subsidence (Atchley et al. 2004; Cleveland et al. 2007; Aziz et al. 2008; Atchley et al. 2013). By modifying concepts originally developed in marine sequences, Atchley et al. (2004, 2013) used combinations of grain size and paleosol thickness, maturity, and drainage to delineate fluvial aggradational cycles (FACs) where fining upward successions are overlain by paleosols (e.g., Figure 9.2C). See Atchley et al. (2013) for a review of paleosol stacking pattern analysis.

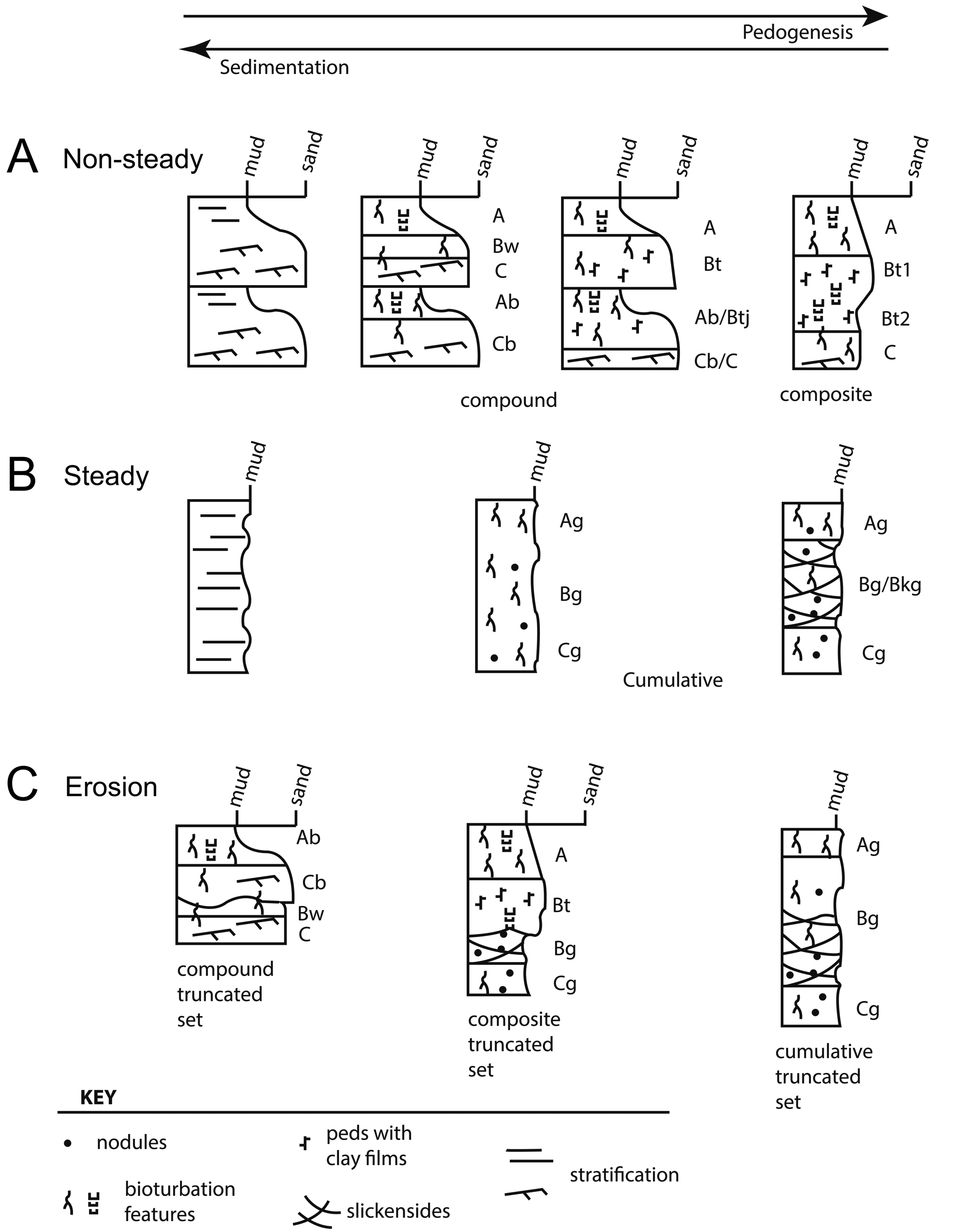

Fig. 9.4 Reproduction of Fig. 9.2 from Kraus (1999) illustrating the pedofacies method of categorizing paleosols. Abbreviations: Ag – gleyed A horizon; Bg – gleyed B horizon; Bw – B horizon showing color or structure development but little if any illuvial accumulation; Bt – B horizon showing accumulation of clays; Btj – incipient development of a Bt horizon; Cg – gleyed C horizon. Reprinted with permission from Elsevier

Table 9.10 Paleosol maturity and drainage indices. Modified from Atchley et al. (2013)

	Rank	Features
Maturity		
1	Very weakly developed	Weak mineral surface horizon (A); no subsurface horizon; massive to faint bedding; root traces
2	Weakly developed	Mineral surface horizon (A); incipient subsurface Bk horizons with less than 5% carbonate nodules, not qualifying as a Calcic diagnostic horizon
3	Moderately developed	Mineral surface horizon (A); subsurface Bk horizon with more than 5% carbonate nodules qualifying as a Calcic diagnostic horizon
4	Strongly developed	Mineral surface horizon (A) and E horizon; red or purple subsurface Bt horizon with sufficient clay enrichment to qualify as an Argillic diagnostic horizon
Drainage		
1	Poorly drained	Gray of blue green surface and subsurface horizons
2	Moderately drained	Yellow or brown subsurface horizons
3	Well drained	Red or purple subsurface horizons

Strengths of Approaches

Because paleosols are by their nature an unconformable surface, they have been ignored in the past in favor of sediment cores from lakes or oceans to understand the effect of long-term climate changes on paleoecology. However, paleosols provide regional, site-based, environmental reconstructions and are thus very useful for understanding paleoecological variability in the immediate site area. Collecting paleosol samples for analysis using the variety of methods previously mentioned is fairly simple. When used properly to reconstruct climate or vegetative cover, paleosols act as *in situ* archives that record environmental conditions on moderate time scales.

With an increase in the detail to which paleoecological data can be extracted from individual specimens and bonebeds as a whole, it is essential to understand the context of fossil accumulations. Where paleosols contain fossil remains, they can be used to infer the method in which fossils were incorporated into the stratigraphic record. The relative degree of paleosol development at or near an archaeological or paleontological site can also provide information on the state of artifact and bone preservation (Holliday 2004). Understanding the depositional and post-depositional context of fossil and archeological sites is a key component to unraveling the taphonomy of the remains to be studied, the evolution of landscapes between depositional events, and can inform researchers on potential biases of preservation for a given locality.

Measuring and analyzing multiple paleosol profiles along a particular buried or exhumed paleo-CZ or paleocatena is a popular approach for reconstructing the paleoecosystem associated with an archaeological site. A catena study assesses the variability of soils across a landscape caused by changes in topography due to changes in hydrology (Morrison 1967; Birkeland 1999; Holliday 2004). Paleolandscapes can be studied at different scales depending on the research question being addressed. At the local scale, the relationship between depositional processes, weathering styles, drainage, and artifact accumulation can be examined. On the basin-scale, amalgamated strata representing a multitude of paleoenvironments can be used to understand long-term changes in climate and subsidence (Kraus 1999; Atchley et al. 2004; Cleveland et al. 2007, 2008; Aziz et al. 2008; Nordt et al. 2012; Atchley et al. 2013; Nordt and Driese 2013). An example of this method at Olduvai Gorge in Tanzania, where correlative surfaces buried by volcanic ashes have been used to reconstruct aspects of paleo-CZs from paleocatena successions (see description in **Applications**; Ashley and Driese 2000; Sikes and Ashley 2007; Ashley et al. 2014; Beverly et al. 2014, 2015a; Driese and Ashley 2016).

Biases and Shortcomings

In the spirit of Vance Holliday's (2004) insightful review of paleosols as paleoenvironmental indicators, we also quote Valentine and Dalrymple (1976, p. 218), "Although soil science is under great pressure to furnish environmental evidence, it is debatable whether we understand the interaction of the soil-forming processes with the site and environmental factors well enough yet to make confident extrapolations." This point alludes to the fact that many researchers make interpretive leaps to paleoenvironmental reconstructions without detailed understanding of the soil-forming and post-burial (diagenetic) processes that may have altered the paleosol system. This point should not deter future paleosol work. Rather, with a detailed analysis and an understanding of the biases discussed here, paleosol research using multiple proxies can yield a more nuanced interpretation of the paleoenvironment.

Disequilibrium with Climate or Environment

Soils are the product of local conditions, which may be considered an advantage in some paleoecological studies using paleosols; however, soils may not reflect the regional conditions due to the local hydrology, which may mimic changes in climate (e.g., Driese and Ashley 2016). It has been shown that multiple pedogenic pathways can yield the same soil property (Holliday 2004). For example, lateral changes in the development of a subsoil argillic (Bt) horizon from clay illuviation may be due to drainage or the duration of soil formation. This phenomenon, known as equifinality, suggests that in some cases a single paleoecological interpretation may not be possible when relying solely on the paleosol record. Thus, it may be difficult to extrapolate environmental and climate reconstructions beyond the site or pedon (Holliday 2004), but if correlation using precise dating methods or correlative tephras is possible, this issue can be avoided by studying the site as a paleocatena or paleo-CZ. By looking at the variations in a paleocatena across the whole landscape rather than focusing solely on the archaeological or fossil site, diachronous soil development can be identified and help researchers better interpret the paleoecology from paleosols. With this detailed method, the changes in local hydrology will be understood within the regional context, and paleoenvironmental and paleoclimatic interpretations can be made.

Pedogenesis on any substrate is a time-transgressive process and may not reach steady state with respect to climate and other environmental factors for upwards of 10^3–10^4 years (Yaalon 1971; Huggett 1998, and references therein). Therefore, less than centennial-scale changes may not be detectable in the soil or paleosol record and may reflect a sum of the environmental changes over the course of formation (Holliday 2004; Richter and Yaalon 2012). The relatively long durations of time required for soils to equilibrate with environmental conditions stands in stark contrast to the time required for the accumulation of artifacts or fossils, which can be instantaneous. Thus, a disparity of time resolution exists between the presence or burial of fossil remains and our ability to reconstruct the environment from which they are derived. This may result in disagreement between results of pedogenic and paleoecological or archeological proxies for environments. Young soils can have morphological and geochemical properties present in other (often drier) climatic settings. For example, excess base cations in immature soils will yield artificially low MAP and MAT estimates using climate proxies, resulting in "mock-aridity" or "mock-frigidity." This should be kept in mind when interpreting climate change in stratigraphic sequences using paleosols of varying maturity (Stinchcomb et al. 2016).

Misuse of Proxies

Many paleosol proxies for climate have inherent limitations and are therefore subject to misapplication. For example, the depth to Bk and By proxies (Retallack 2005; Retallack and Huang 2010) predicts MAP based on the depth from the surface of a soil to the zone of nodular carbonate precipitation. As mentioned previously, many paleosols in the rock record are truncated by overlying fluvial deposits such that A horizons are a rarity in alluvial paleosols (Fig. 9.2C; Myers et al. 2014). This limits the depth to carbonate proxy to relatively few paleosols and must be applied with justification that A horizons were present or that the MAP value is an under estimation (Nordt et al. 2006; Prochnow et al. 2006). Early applications of the depth to carbonate proxy were performed without assessment of the origin of the carbonate, presence of surface horizons, or taxonomic interpretation of paleosols (e.g., Retallack 1997; Sheldon and Retallack 2004). It is now recognized that depth to carbonate in Vertisols follows a unique function (Nordt et al. 2006; Prochnow et al. 2006) that deviates from the more universal equation of Retallack (2005), which should caution researchers in application of such proxies in the absence of detailed paleosol description. Bulk geochemical proxies for MAP and MAT can be easily misused as well. Most paleosol proxies for climate are limited to specific soil orders or depositional conditions (Table 9.9). Misclassification of paleosols by horizons or soil order can therefore result in misapplication of climofunctions.

Perhaps a more widespread abuse of paleosols in climate studies is that very few researchers verify that the discrete geochemical values of paleosol samples fall within the range of values used to generate climate proxies. Because the relationship between a geochemical index and climate (e.g., CIA-K and MAP) is derived through a training data set, any applications of the regression equation must be confined to data that are similar to the training data. For example, unusually high concentrations of MgO (up to 17%) are present in paleosols at Olduvai Gorge because the parent material is an authigenic lacustrine Mg-rich smectite (Beverly et al. 2014). This is significantly more MgO than the dataset on which the CALMAG precipitation proxy was developed, which is only applicable in paleosols with <3% MgO (Nordt and Driese 2010a, b). This does not mean that these proxies are ineffective at predicting precipitation in diverse settings, but that care must be taken to correctly apply the proxies.

Stable isotopes of pedogenic carbonates are also commonly used as paleoecological indicators, but an understanding of soils and soil formation is necessary to apply these proxies. Researchers should interpret the presence of carbonate nodules or pedogenic carbonates in general with caution as several studies have shown that these features can be influenced by soil hydraulic properties (porosity, permeability), texture, clay content, and slope position (Holliday 2004). Also, the identification of carbonate in a paleosol does not mean that it is pedogenic in origin. For example, carbonate concretions are commonly found at Olduvai Gorge up to 20 or 30 cm in diameter and are associated with post-depositional groundwater movement through some paleosols (Fig. 9.2N), as indicated by their large size, irregular morphology, and sparry texture.

Reworking and Inheritance

Paleosols are components of sedimentary systems and are therefore subject to processes of erosion and deposition. The parent material of a soil reflects regional geology, which can include soils actively forming under variegated environmental (e.g., *cl, o, r, p, t)* conditions, bedrock of various compositions and windblown sediment. Any material removed from these reservoirs can be deposited on or incorporated into a soil. Geochemical proxies for climate (e.g., CIA-K) and vegetation (e.g., phytoliths, $\delta^{13}C$ of organic matter) can be altered or biased if any constituent is included from reworked sediment.

Pedotransfer functions for climate that rely on paleosol bulk geochemistry will provide underestimates if sediment is added through any means of deposition. Geochemical climofunctions should not be applied to cumulative paleosols (Sheldon et al. 2002; Sheldon and Tabor 2009), which can form in eolian, fluvial, or mixed-source environments. Nordt and Driese (2010a) note that the smectite on which most Vertisols form is inherited from the surrounding catchment as alluvial clays rather than being authigenically manufactured in the soil profile. Other reworked pedogenic minerals, such as kaolinite or hematite, can make relatively immature paleosols appear to be mature, or can influence the color of a paleosol. Cross-referencing mineralogy with field and micromorphologic observations can mitigate misapplication or misinterpretation of paleoclimate in such settings.

Reworked organic matter and pedogenic carbonates are also a potential problem in alluvial deposits. Because paleosols typically have very low organic C concentrations (Wynn 2007), the relative abundance of *in situ* versus inherited organic matter can be difficult to discern. Rounded pedogenic carbonates found with coarser grains at the base of alluvial deposits are clear evidence for reworking (Lukens et al. 2017b). Comparison between SOM $\delta^{13}C$, pedogenic carbonate $\delta^{13}C$, and phytoliths has been a useful tool to reveal biases in the pedogenic carbonate vegetation signal (Cotton et al. 2012; Chen et al. 2015; Garrett et al. 2015). The difference between the $\delta^{13}C$ of co-occurring organic matter and pedogenic carbonate is 14–17‰ in modern systems (Cerling and Quade 1993) but can potentially shift beyond that range given changes in atmospheric pCO_2 in the past (e.g., Mora et al. 1996) or isotopic fractionation of SOM during decomposition (Wynn 2007).

Diagenesis

Any alteration of paleosols after burial is referred to as diagenesis, and since the study of paleosols inherently involves analyzing soils that have left the *cl*, *o*, *r*, *p*, *t* conditions in which they were formed, one must carefully differentiate physical and geochemical features formed during pedogenesis versus those formed through diagenesis. Lithification is perhaps the most readily recognizable product of diagenesis and involves compaction and cementation. Compaction reduces pore volume, which may sufficiently inhibit fluid flow in clay-rich paleosols, but is unlikely to prevent groundwater transmission through paleosols with sandier textures. Compaction can also lead to deformation of mineral grains and reduction on primary porosity (Pettijohn et al. 1987).

Post burial cementation – not to be confused with pedogenic calcretes or fragipans – occurs in groundwater and plugs void space with a range of cementing agents. Calcite, silica, iron oxides, and clays are the most common cements in sedimentary rocks and are recognizable due to their occurrence within primary voids (e.g., root traces), along features of textural contrast (e.g., burrows infilled with sediment of an overlying deposit), or as overgrowths on mineral grains and pedofeatures (Stoops 2003). Calcareous paleosols typically show a mix of carbonate morphologies due to both pedogenic and diagenetic remobilization of primary pedogenic carbonates. The current rule of thumb in sampling pedogenic carbonate is to target micrite and avoid any material with sparry or megacrystalline texture (Sheldon and Tabor 2009; Tabor and Meyers 2015), though soils formed on coarse-grained sediments can contain pedogenic carbonates with sparry or microspar texture (Weider and Yaalon 1982). Any researcher attempting to study pedogenic carbonate for paleovegetation or paleoclimate reconstruction should first analyze carbonates in thin section and micro-drill targeted zones to avoid diagenetic overgrowths. Cathodoluminescence is a useful technique to identify histories of carbonate cementation (Barnaby and Rimstidt 1989; Machel et al. 1991; Michel et al. 2013) and should be considered before isotopic sampling is performed.

Diagenetic mineralization can alter the color of a paleosol to reflect the color of cementing agents or as a result of the dissolution of unstable minerals. Burial reddening occurs due to hematite cementation in oxidizing groundwater (Retallack 2001), whereas calcite cements can impart a white to pale color. Decomposition of organic matter in paleo-surface horizons will remove the dark colors evident in modern O and A horizons, leaving more uniform coloration to preserved paleosol profiles. If organic matter is decomposed by anaerobes after the paleosol is buried below the water table, blue-green drab halos will form around individual root traces or laterally along paleo-root mats (Retallack 2001). These colors can be differentiated from gleying due to saturation by identification of pedogenic features indicative of well-drained conditions, such as ped structure, translocation of clay or Fe, and precipitation of pedogenic calcite at depth.

Paleosols flooded by lacustrine or marine waters can exhibit a drab, reduced zone near their upper boundary (e.g., Driese and Ober 2005). This process is called pseudogley because it imparts gleyed colors typically observed in saturated soils. Pedogenic features indicative of episaturation should be present if gleying were due to periodic standing water on the soil surface, such as lateral-branching root traces, Fe-redox halos around roots and voids, pedogenic siderite, pyrite or jarosite, or accumulations of organic matter. Pseudogley has also been attributed to flushing of alkaline groundwater through buried paleosols (Pimentel et al. 1996).

Potassium is added during diagenetic conversion of smectite to illite (Fedo et al. 1995), which is the primary reason for excluding K_2O from pedotransfer functions and climate proxies (Harnois 1988; Maynard 1992). However, the physical pedogenic features indicative of smectite-rich Vertisols – including slickensides, sepic-plasmic clay fabrics and wedge peds – will remain after illitization has taken place (e.g., Driese and Foreman 1992). Current practice in the paleopedology community is to avoid using bulk geochemical data for climofunctions or pedotransfer functions if paleosols have experienced illitization or significant diagenetic mineralization (Sheldon and Tabor 2009; Driese et al. 2007; Medaris et al. 2017).

Pedogenic Mixing of Artifacts or Fossils

Pedogenic mixing processes can obscure discrete artifact layers and widen the depth range (Ferring 1992; Waters 1992; Cremeens et al. 1998). This can occur with aggressive bioturbation by soil fauna, through pedogenic processes such as cryoturbation from freeze-thaw cycles or due to vertic processes causing movement of artifacts or bones. Vertisol movement can also break and polish artifacts (Quade et al. 2004). However, even in Vertisols, which have considerable internal mobility of mass due to shrink-swell processes, artifacts can be placed in relative stratigraphic order with careful study (Waters et al. 2011).

Applications

The paleopedological techniques discussed in this chapter have been applied throughout geologic time, including diverse settings such as Neoarchean paleoweathering surfaces (Driese et al. 2011), late Ordovician and Silurian Appalachian Basin paleosols (Driese and Foreman 1992; Driese et al. 1992), Triassic paleosols at Petrified Forest, Arizona (Cleveland et al. 2007, 2008; Trendell et al. 2013a, b), paleosols spanning the Paleocene-Eocene Thermal Maximum in the Bighorn Basin, Wyoming (Kraus 1997; Kraus and Brown 1988; Kraus and Hasiotis 2006; Aziz et al. 2008; Snell et al. 2013) and Pleistocene-Holocene paleosols in eastern North America (Driese et al. 2005). Here we present examples from East Africa where paleosols have been used to understand the paleoecology at paleontological and archaeological sites throughout the Miocene and Quaternary.

There is a long history of paleosol research in East Africa that began with stable isotopes of C and O from pedogenic carbonates in the Turkana Basin (Cerling et al. 1988) and Olduvai Gorge (Cerling and Hay 1986). These extensive, well-dated paleosol records showed that climate became warmer and drier over the past 2 Ma (Cerling and Hay 1986; Cerling et al. 1988; Levin et al. 2011). This research ushered in a wave of studies focusing on stable isotopes of pedogenic carbonates to understand the paleo-vegetation and paleo-rainfall history at archaeological sites (Cerling et al. 1988; Wynn and Feibel 1995; Sikes et al. 1999; Wynn 2000; Levin et al. 2004; Quade et al. 2004; Wynn 2004; Lepre et al. 2007; Quinn et al. 2007; Campisano and Feibel 2008; White et al. 2009; WoldeGabriel et al. 2009; Cerling et al. 2010; Passey et al. 2010; Cerling et al. 2011; Levin et al. 2011; Quinn et al. 2013; Garrett et al. 2015; Lüdecke et al. 2016). For summaries of the use of stable isotopes in paleoenvironmental reconstructions at hominin sites see reviews by Cerling (2014) and Levin (2015).

This focus on stable isotopes of pedogenic carbonate has yielded long-term records during the Pleistocene for comparison with marine records but has also ignored the wealth of paleoclimatic and paleoenvironmental information that can be gained from the breadth of methods discussed in this chapter. Much remains to be discovered regarding the historical and process-based details of how hominins interacted with their surrounding ecosystem (National Research Council 2010). In this section, we focus paleosol applications at Olduvai Gorge to demonstrate the process and utility of paleosol research (Fig. 9.5). Other similar studies have recently been completed at significant fossil sites in western Kenya at Rusinga Island (Michel et al. 2014) and Karungu (Beverly et al. 2015a; Driese et al. 2016; Lukens et al. 2017b).

Olduvai Gorge is a well-dated archeological site that has been studied for over 50 years, but only recently have the paleosols been recognized for their paleoenvironmental potential (Ashley and Driese 2000; Hover and Ashley 2003; Sikes and Ashley 2007; Magill et al. 2013a, b; Ashley et al. 2014; Beverly et al. 2014; Driese and Ashley 2016). Using well-dated and laterally correlative tephras, Olduvai Gorge can be reconstructed as a series of paleocatenas or paleo-CZs at multiple scales of detail (Fig. 9.6; Ashley et al. 2014; Beverly et al. 2014).

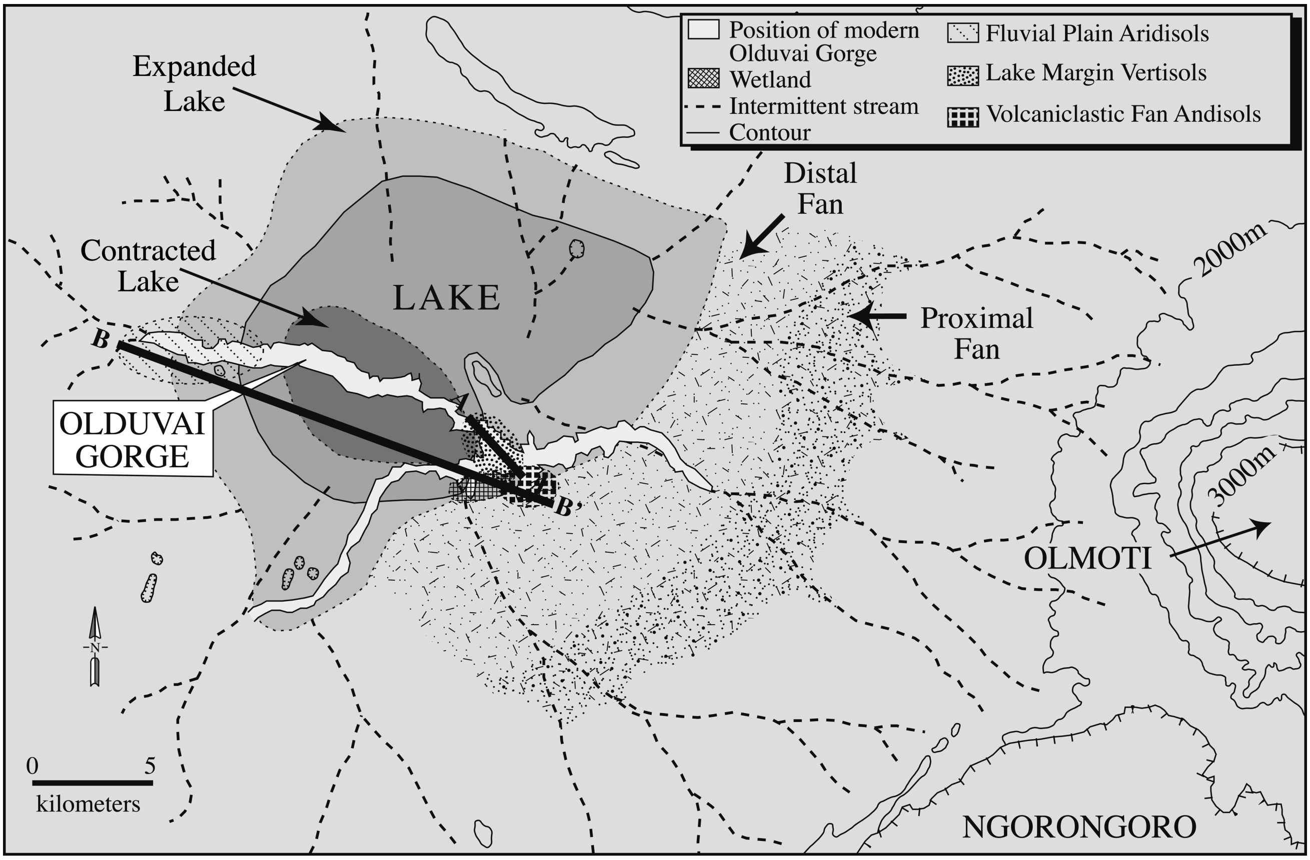

Fig. 9.5 Paleogeographic map of Olduvai Gorge modified from Ashley et al. (2010) and Beverly et al. (2014) showing paleo-Lake Olduvai and the locations of soil-forming environments. Cross-sections A and B are shown in Fig. 9.6

Near the margin of paleo-Lake Olduvai a series of time-equivalent, stacked paleosols were identified and correlated using field relationships and tephrostratigraphy. Four trenches were described by horizon and classified as paleo-Vertisols and then sampled every 10 cm for bulk geochemistry. Oriented samples for micromorphology were also collected. The combination of micromorphology and bulk geochemistry proved to be a powerful tool in describing a paleocatena where minor differences in paleotopography due to small faults and distance from a fault-controlled spring affected the amount and type of pedogenesis. Close to the Zinj fault and a related spring on the downthrown fault block, stacked paleosols developed (Figs. 9.3D and 9.6A), but further away cumulative soils developed (Fig. 9.6A). Using micromorphology, the degree of pedogenesis could be assessed, and with UVf, the amount of organic matter could be visually estimated. Closer to the spring, weakly developed b-fabrics (monostriated), Fe–Mn filled rhizoliths, and high abundances of organic matter were observed. Pedogenesis was weaker due to a higher water table that did not allow for movement of water through the soil (Fig. 9.6A). Further from the spring, organic matter abundance decreased, but b-fabrics were better developed (parallel striated, cross-striated, and granostriated), volcaniclastic grains were weathered and pitted, and Fe–Mn and zeolites formed hypocoatings along ped and pore boundaries. When combined with the phytolith and pollen record from these paleosols (Barboni et al. 2010), a reconstruction of the paleocatena is complete with a groundwater forest closest to the spring and a more open environment with palms and grasses further from the spring (Fig. 9.6A).

Because the carbonates in these paleo-Vertisols were interpreted to be diagenetic precipitates based on their large, irregular size and sparry crystalline texture, bulk geochemical methods were utilized to reconstruct paleoclimatic conditions. An example of mass-balance from the bulk geochemistry is presented here using Na, Al, Si, and K, which are typically associated with clay translocation, but because this paleo-Vertisol formed on Mg-rich smectites, Mg is also included (Fig. 9.7A–B). Results of mass-balance calculations indicate that greater translocation occurs up-section, which is best illustrated with a box and whisker plot using Na as an example (Fig. 9.7C). The increased translocation overlaps with a precession cycle that has been previously identified in uppermost Bed I using stable isotopes, fauna, flora and lithostratigraphy (Cerling and Hay

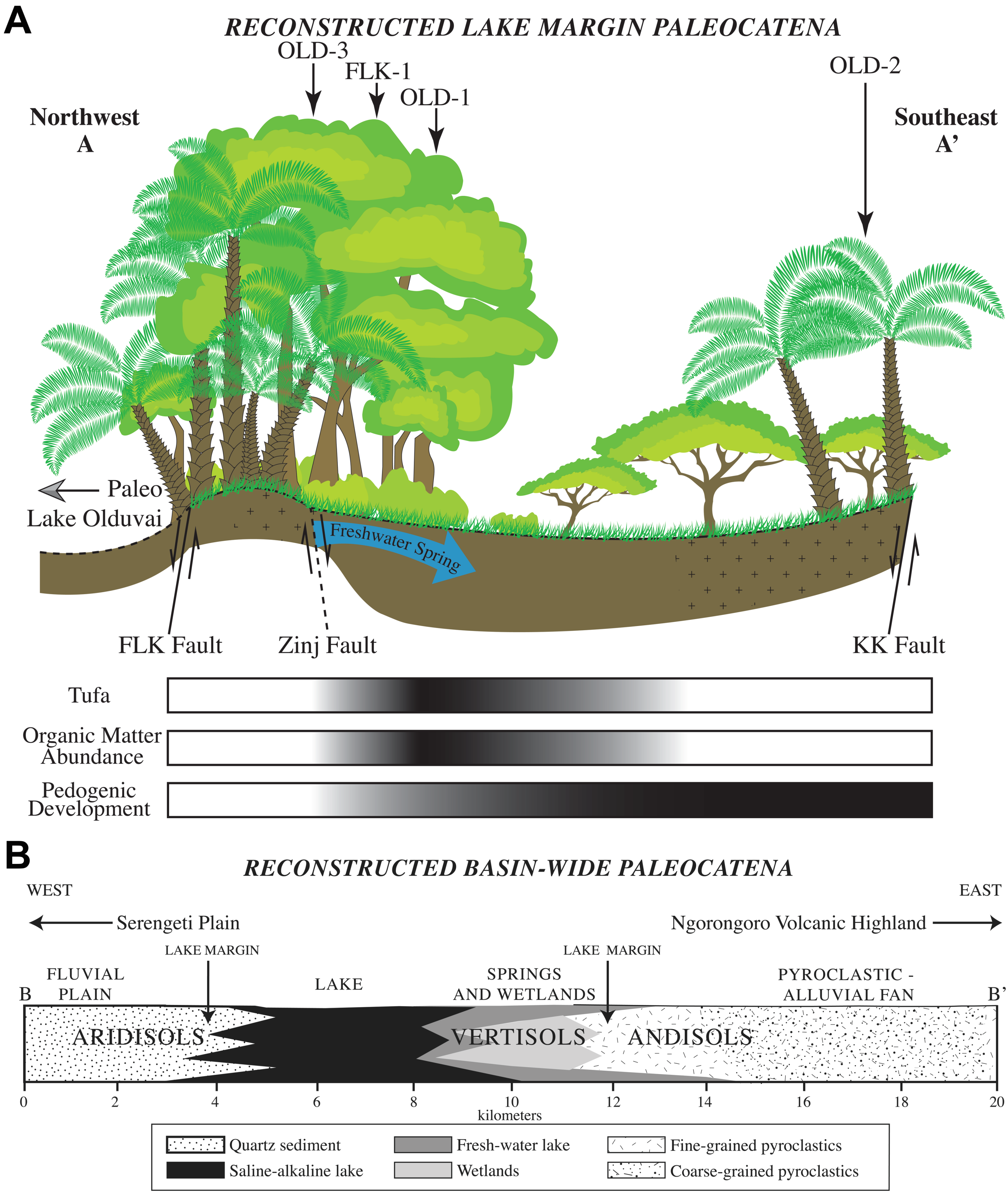

Fig. 9.6 Reconstruction of a paleocatena or paleo-Critical Zone from Olduvai Gorge at two different scales reproduced from Beverly et al. (2014). The locations of cross-sections A and B are shown in Fig. 9.4. A. The local scale reconstructs a paleocatena along the margin of paleo-Lake Olduvai using field description of paleosols, micromorphology, and mass-balance of bulk geochemistry where very minor differences in topography and distance from the spring affected pedogenesis. B. The basin scale reconstruction is a summary of a decade of paleosol research and combines the use of bulk geochemistry, stable isotopes, mineralogy, and detailed field descriptions

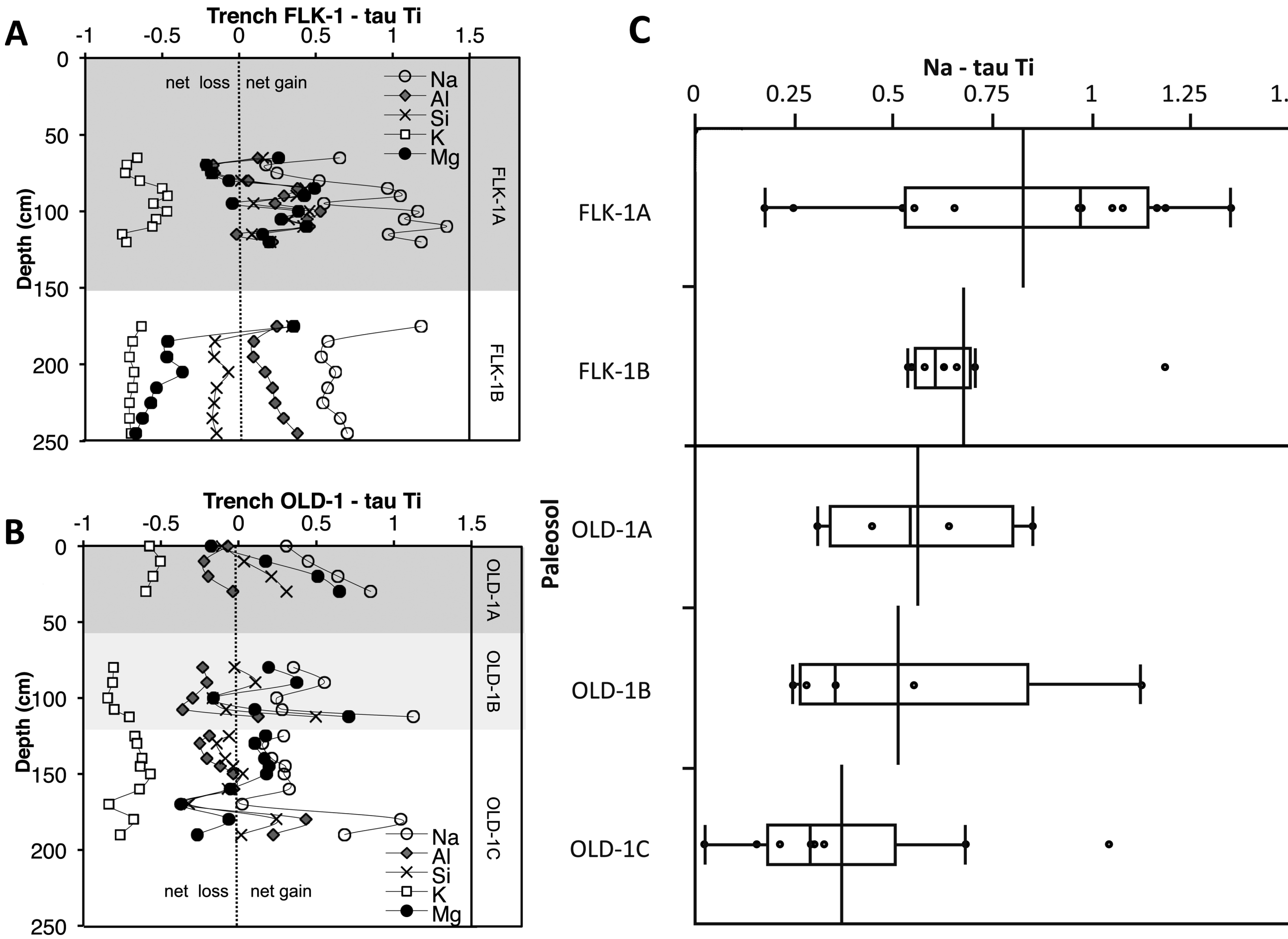

Fig. 9.7 Application of mass-balance to Olduvai paleosols. Reproduced from Beverly et al. (2014). A and B, Time-equivalent stacked paleosols near the lake margin of paleo-Lake Olduvai that can be correlated using field relationships and tephrostratigraphy. This example shows elements grouped by association with clay or zeolite accumulation. Ti is held constant as the immobile element (Tau Ti) and positive values indicate a net gain and negative values indicate a net loss relative to the parent material. Multiply by 100 to convert to percent gains or losses. C, The box and whisker plot of Tau – Na using Ti as the immobile element for each paleosol indicates that the amount of translocation increases up section

Table 9.11 Summary of Olduvai paleosol characteristics modified from Ashley et al. (2014)

Paleosol Characteristics	Fluvial Plain Paleo-Aridisol	Lake Clay Paleo-Vertisol	Pyroclastic Fan Paleo-Andisol
Parent material	Quartzo-feldspathic sediments	Dominantly Mg-smectite lake clay with some tephra influence	Volcaniclastic sediments and tephra
Texture	Sand and silt	Clay dominant	Very coarse sand to silt
Drainage	Well-drained	Dominantly well-drained to poorly drained during lake highstands	Dominantly well-drained but poorly drained proximal to springs
Immobile elements	Zr > Ti: dominated by Serengeti Plain	Ti ≈> Zr: mixed Serengeti Plain and basalts	Ti > Zr: dominated by basalts
Calcium Carbonate	High, includes calcrete	Low	Low, except for detrital grains
Zeolites	Generally low	Abundant, filling root pores and channels, replacing tephra grains	Very abundant, replacing tephra grains and filling root pores
Shrink-swell (vertic) features	Absent	Abundant, slickensides and wedge peds	Absent
Redoximorphic features	Absent	Rare	Rare on fan, abundant near springs

1986; Ashley 2007; Barboni et al. 2010; Beverly et al. 2014, and references therein).

These methods can also be applied to a basin-wide reconstruction of the paleocatena or paleo-CZ (Figs. 9.5 and 9.6B). By analyzing these paleosols in a more holistic method that includes field relationships and careful description of paleosol features such as ped structure or redoximorphic features, stable isotopes, micromorphology, bulk geochemistry, and grain size (Table 9.11), a detailed reconstruction of the paleoenvironment in which hominins were living can be created (Fig. 9.6B; Ashley et al. 2014). Differences in parent material and topography have a great effect on paleosol development at Olduvai Gorge irrespective of climate. Paleo-Aridisols formed in the quartzo-feldspathic sediment on the fluvial plain to the west of paleo-Lake Olduvai (Fig. 9.2A–B) in contrast to the paleo-Vertisols formed in the Mg-rich smectitic clay along the lake margin (Fig. 9.2D–F). The topography of the basin has a great effect on the water table, and further from the lake, soil drainage increased and soils were generally redder in color due to oxidation of Fe (Ashley and Driese 2000; Sikes and Ashley 2007; Ashley et al. 2014). This example illustrates profoundly variegated soils that result from variations in climate, organisms, relief, parent material, and time (*cl, o, r, p, t*). Without careful study and correlation using tephrostratigraphy, the changes in paleosol types could easily be attributed to changes in one specific soil-forming factor such as climate – the paleo-Vertisol high in OM would have formed a wetter climate and carbonate-rich paleo-Aridisols would have formed in a drier climate. In actuality, these paleosols formed coevally on the landscape and the changes in relief and parent material were predominant controls on soil formation. With careful study, climate can be teased out of this complex signal, but climate cannot be the assumed cause of paleosol heterogeneity on the landscape (Figs. 9.5, 9.6A–B; Ashley and Driese 2000; Sikes and Ashley 2007; Ashley et al. 2014; Beverly et al. 2014).

Future Prospects

The addition and refinement of paleosol proxies for climate and environment will advance paleopedology-based paleoecological reconstructions in the near future. The recent development of instrumentation capable measuring of clumped isotopes (Δ_{47}) in paleosol carbonates has allowed for great advances in reconstructing paleotemperature using paleosols. Ongoing research suggests that clumped isotopes may be seasonally biased and sensitive to soil hydroclimate (Passey et al. 2010; Gallagher and Sheldon 2016), and presents opportunities for future research. In addition, the development of the capability to measure ^{17}O in addition to ^{16}O and ^{18}O presents new and interesting prospects in understanding the relationship between evaporation and precipitation using triple oxygen isotopes (Passey et al. 2014). New microanalytical techniques on pedogenic carbonate using U-series dating and laser ablation ICP-MS and ion microprobe to measure variations in $\delta^{18}O$ and $\delta^{13}C$ are also pushing forward the field of paleopedology (Oerter et al. 2016).

Larger modern soil data sets with more quantitative soil attributes will yield new models for predicting both environmental factors and soil properties. This may improve both climate prediction and the ability to more rigorously classify paleosols using Soil Taxonomy (Nordt and Driese 2013). Larger datasets may also provide a means for developing multiple models designed for specific soil characteristics (e.g., soil order, drainage, parent material). Using a data-driven, multivariate statistical approach applied to a large soil dataset, Stinchcomb et al. (2016) showed that both MAP and MAT can be predicted simultaneously for a diverse range of soils forming across a large range of climatic conditions. This is especially useful when the paleopedologist has little environmental data at the outset of a project. Greater effort in estimating fundamental soil properties also has a bright future in the field of paleopedology. Soil pH and Eh proxies were used to construct a "paleo-Pourbaix diagram" used in the assessment of soil nutrient cycling and ecosystem response from Late Cretaceous and early Paleocene paleosols in the Western Interior of North America (Nordt et al. 2011). Similar efforts from East African and other locations could yield new insight into soil-plant-water interactions accompanying evolution.

Key Terms

Catena – A series of closely-related soils that manifest unique properties due to lateral changes in soil-forming conditions that vary as a function of slope. The catena concept has been applied to paleosols as their position relates to valley bottom features such as springs and channels (see pedofacies).

Critical Zone (CZ) – Near-surface environment in which complex interactions involving rock, soil, water, air, and living organisms regulate the natural habitat and determine the availability of life-sustaining resources (National Research Council 2001).

Edaphic – Influenced, produced, or related to soil, typically in reference to mineralogical or chemical characteristics that affect biota.

Eluviation – The transport and removal of soil solutes or material (mineral or organic) from an upper soil horizon. The material is commonly clay to silt size or dissolved in pore water.

Illuviation – The introduction of soil solutes or material (mineral or organic) into a soil horizon from an overlying horizon. The material is commonly clay to silt size.
Lithologic discontinuity – The contact between two different sediment types or genetically unrelated sediments that is usually marked by a distinct surface. When present, time may be unrepresented in the stratigraphic record.
Master horizon – A dominant soil horizon within the soil profile. Master horizons include: O, A, E, B, C, R, and K. The B and C master horizons are the most common in the sedimentary record. Master horizons also may have subordinate horizon modifiers describing secondary characteristics within that soil horizon.
Paleo-Critical Zone (paleo-CZ) – The deep-time equivalent of a modern CZ that is primarily studied using proxies for reconstructing past biogeochemical cycles related to environmental factors. The paleo-CZ is also known as a Deep Time Critical Zone (DTCZ). The DTCZ concept was originally confined to pre-Quaternary rocks and sediments (Nordt and Driese 2013). The concept is extended to the Quaternary (<2.6 Ma) in this chapter.
Paleoecology – The study of animal and/or plant interactions within ancient environments. Studies may incorporate direct fossil evidence or rely on proxy methods.
Paleopedology – The study of paleosols.
Paleosol – A soil that formed on a landscape of the past or under soil-forming conditions no longer operating at a given locality. Paleosols can occur as buried, exhumed, lithified, and/or relict soils. Paleosols are also known as fossil soils.
Parent material – The material (sediment or rock) from which a soil forms.
Pedogenic – Adjective used for soil, soil-forming processes, or products of soil formation (e.g., pedogenic minerals)
Pedology – A subdiscipline of soil science that emphasizes the study of soil formation, especially as it relates to environmental factors (e.g., climate, organisms, relief, parent material, and time).
Soil – a depth profile consisting of distinct layers that result from the physical, chemical, and biological alteration (i.e., weathering) of pre-existing parent materials into a more stable form.
Soil facies or pedofacies – lateral variation in soil development caused by variation in environmental factors of soil formation (i.e., topography).
Soil horizon – A layer within the soil profile that is different from the overlying and underlying layers, usually based on physical features such as color and texture.
Surface horizon – The zone within a soil profile dominated by organic matter inputs from the vegetative cover (e.g., A horizon).
Subordinate horizon indicators – Lower-case letters that are appended to master horizon designations that add distinction to the master horizon (see Table 9.2 for a list of subordinate horizon indicators).
Subsurface horizon – Also called the mineral soil, it is a zone where chemical reactions and physical actions cause the mobilization of ions and fine-grained particles, including clays, metals, and organo-metallic complexes (e.g., B horizon).
Translocation – The movement (usually downward) of chemical and physical constituents through the percolation of water in pores or via biotic agents.

Key References

1. *Soils in Archaeological Research* (Holliday 2004) is devoted to soil and paleosol investigations as they relate to the archaeological record. This textbook is geared towards Quaternary paleopedologists and geoarchaeologists and has a chapter devoted to soils in paleoenvironmental reconstructions.
2. *The Nature and Properties of Soils* (Brady and Weil 2008) is a comprehensive introduction to modern soil science with an emphasis on chemical, physical, and mineralogical characterization of weathering as it pertains to the modern U.S. Soil Survey program.
3. *Soils and Geomorphology* (Birkeland 1999) is a well-known textbook on the dynamics of soil formation related to the five environmental factors: climate, organisms, relief, parent material, and time. This book is frequently used by modern and paleopedologists.
4. *Guidelines for analysis and description of soil and regolith thin sections* (Stoops 2003) is widely available text that serves as a step-by-step guide for analyzing and describing soil thin-sections and includes examples and a CD with images. This book is partially based on Bullock et al. (1985), which is more difficult to find.
5. *Interpretation of Micromorphological Features of Soils and Regoliths* (Stoops et al. 2010) is a textbook with excellent color images illustrating features that can be identified in soils and paleosols.
6. *Stable isotope evidence for hominin environments in Africa* (Cerling 2014) is a chapter summarizing how stable isotopes can be used to reconstruct hominin environments.
7. *Environment and Climate of Early Human Evolution* (Levin 2015) is another summary of the use of stable isotopes to reconstruct hominin paleoenvironments.
8. *Quantitative paleoenvironmental and paleoclimatic reconstruction using paleosols* (Sheldon and Tabor 2009) is an extensive summary of the state of paleosol research and is good place to start for those interested in utilizing paleosols for paleoenvironmental and paleoclimatic research.

9. *Paleosols as Indicators of Paleoenvironment and Paleoclimate* (Tabor and Myers 2015) is an update to the Sheldon and Tabor (2009) summary of paleosol research.
10. *Revisitation of Concepts in Paleopedology: Transactions of the Second International Symposium on Paleopedology* (Follmer 1998) is a collection of abstracts, essays and reports by various authors on numerous facets of paleosols, following the first paleopedology symposium (Yaalon 1971).
11. *Weathering, Soils & Paleosols* (Martini and Chesworth 1992) is an edited volume that is a blend of pedology, classic geology (especially mineralogy and geochemistry) and geomorphology that discusses present and past weathering processes. Although this book offers an intriguing and unique view of paleopedology, it contains little content on paleosols as paleoecological indicators.
12. *A Colour Guide to Paleosols* (Retallack 1997) is a hard-to-find introductory book on paleopedology that contains remarkable paleosol images and accessible text on the basics of the field, including paleosol identification, description and interpretation.
13. *Soils of the Past: An introduction to paleopedology* (Retallack 2001) is a paleopedology textbook that introduces concepts and techniques in the field and provides a chapter-by-chapter overview of paleosols recorded through Earth history and their paleoenvironmental importance.

References

Amundson, R., Richter, D. D., Humphreys, G. S., Jobbagy, E. G., & Gaillardet, J. (2007). Coupling between biota and Earth materials in the critical zone. *Elements, 3,* 327–332.

An, Z. S., & Porter, S. C. (1997). Millennial-scale climatic oscillations during the last interglaciation in central China. *Geology, 25,* 603–606.

Anderson, S. P., Dietrich, W. E., & Brimhall, G. H. (2002). Weathering profiles, mass-balance analysis, and rates of solute loss: linkages between weathering and erosion in a small, steep catchment. *Geological Society of America Bulletin, 114,* 1143–1158.

Ashley, G. M. (2007). Orbital rhythms, monsoons, and playa lake response, Olduvai Basin, equatorial East Africa (ca. 1.85–1.74 Ma). *Geology, 35,* 1091–1094.

Ashley, G. M., & Driese, S. G. (2000). Paleopedology and paleohydrology of a volcaniclastic paleosol interval: implications for Early Pleistocene stratigraphy and paleoclimate record: Olduvai Gorge, Tanzania. *Journal of Sedimentary Research, 70,* 1065–1080.

Ashley, G. M., Barboni, D., Domínguez-Rodrigo, M., Bunn, H. T., Mabulla, A. Z. P., Diez-Martin, F., et al. (2010). Paleoenvironmental and paleoecological reconstruction of a freshwater oasis in savannah grassland at FLK North, Olduvai Gorge, Tanzania. *Quaternary Research, 74,* 333–343.

Ashley, G. M., Deocampo, D. M., Kahmann-Robinson, J. A., & Driese, S. G. (2013). Groundwater-fed wetland sediments and paleosols: it's all about water table. In S. G. Driese & L. C. Nordt (Eds.), *New frontiers in paleopedology and terrestrial paleoclimatology* (SEPM Special Publication No. 104) (pp. 47–62). Tulsa: SEPM.

Ashley, G. M., Beverly, E. J., Sikes, N. E., & Driese, S. G. (2014). Paleosol diversity in the Olduvai Basin, Tanzania: effects of geomorphology, parent material, depositional environment, and groundwater on soil development. *Quaternary International, 322,* 66–77.

Aslan, A., & Autin, W. J. (1998). Holocene flood-plain soil formation in the southern lower Mississippi Valley: implications for interpreting alluvial paleosols. *Geological Society of America Bulletin, 110,* 433–449.

Atchley, S. C., Nordt, L. C., Dworkin, S. I., Ramezani, J., Parker, W. G., Ash, S. R., et al. (2013). A linkage among Pangean tectonism, cyclic alluviation, climate change, and biologic turnover in the Late Triassic: the record from the Chinle Formation, Southwestern United States. *Journal of Sedimentary Research, 83,* 1147–1161.

Atchley, S., Nordt, L., & Dworkin, S. I. (2004). Eustatic controls on alluvial sequence stratigraphy: a possible example from the Cretaceous-Tertiary transition of the Tornillo Basin, Big Bend National Park, West Texas, U.S.A. *Journal of Sedimentary Research, 74,* 391–404.

Aziz, H. A., Hilgen, F. J., van Luijk, G. M., Sluijs, A., Kraus, M. J., Pares, J. M., et al. (2008). Astronomical climate control on paleosol stacking patterns in the upper Paleocene-lower Eocene Willwood Formation, Bighorn Basin, Wyoming. *Geology, 36,* 531–534.

Bae, C. J. (2013). Archaic Homo sapiens. *Nature Education Knowledge, 4*, 4.

Barboni, D., Ashley, G. M., Domínguez-Rodrigo, M., Bunn, H. T., Mabulla, A., & Baquedano, E. (2010). Phytoliths infer dense and heterogeneous paleovegetation at FLK North and surrounding localities during upper Bed I time, Olduvai Gorge, Tanzania. *Quaternary Research, 74,* 344–354.

Barnaby, R. J., & Rimstidt, J. D. (1989). Redox conditions of calcite cementation interpreted from Mn and Fe contents of authigenic calcites. *Geological Society of America Bulletin, 101,* 795–804.

Berke, M. A. (2017). Reconstructing Terrestrial Paleoenvironments Using Sedimentary Organic Biomarkers. In D. A. Croft, S. W. Simpson, & D. F. Su (Eds.), *Methods in paleoecology: Reconstructing Cenozoic terrestrial environments and ecological communities* (pp. 121–149). Cham: Springer.

Berke, M. A. (2018). Reconstructing terrestrial paleoenvironments using sedimentary organic biomarkers. In D. A. Croft, D. F. Su & S. W. Simpson (Eds.), *Methods in paleoecology: Reconstructing Cenozoic terrestrial environments and ecological communities* (pp. 121–149). Cham: Springer.

Beverly, E. J., Ashley, G. M., & Driese, S. G. (2014). Reconstruction of a Pleistocene paleocatena using micromorphology and geochemistry of lake margin paleo-Vertisols, Olduvai Gorge, Tanzania. *Quaternary International, 322–323,* 78–94.

Beverly, E. J., Driese, S. G., Peppe, D. J., Arellano, L. N., Blegen, N., Faith, J. T., et al. (2015a). Reconstruction of a semi-arid Late Pleistocene paleocatena from the Lake Victoria region, Kenya. *Quaternary Research, 84,* 368–381.

Beverly, E. J., Driese, S. G., Peppe, D. J., Johnson, C. R., Michel, L. A., Faith, J. T., et al. (2015b). Recurrent spring-fed rivers in a Middle to Late Pleistocene semi-arid grassland: implications for environments of early humans in the Lake Victoria Basin, Kenya. *Sedimentology, 62,* 1611–1635.

Bestland, E. A., & Retallack, G. J. (1993). Volcanically influenced calcareous paleosols from the Kiahera Formation, Rusinga Island, Kenya. *Journal of the Geological Society of London, 148*, 1067–1078.

Bestland, E. A., Retallack, G. J., & Swisher III, C. C. (1997). Stepwise climate change recorded in Eocene-Oligocene paleosol sequences from central Oregon, *The Journal of Geology, 105*, 153–172.

Birkeland, P. W. (1999). *Soils and geomorphology*. Oxford: Oxford University Press.

Blake, G. R., & Hartge, K. H. (1986). Bulk Density. In A. Klute (Ed.), *Methods of soil analysis: Part I. physical and mineralogical methods* (pp. 363–375). Madison: Soil Science Society of America Inc.

Blokhuis, W. A., Kooistra, M. J., & Wilding, L. P. (1990). Micromorphology of cracking clayey soils (Vertisols). In L. A. Douglas (Ed.), Soil micromorphology: a basic and applied science, *Developments in Soil Science, 19*, 123–148.

Bown, T. M., & Kraus, M. J. (1987). Integration of channel and floodplain suites in aggrading fluvial systems, I. Developmental sequence and lateral relations of lower Eocene alluvial paleosols, Willwood Formation, Bighorn Basin, Wyoming. *Journal of Sedimentary Petrology, 57*, 587–601.

Brady, N. C., & Weil, R. R. (2008). *The nature and properties of soil* (14th ed).

Brantley, S. L., Goldhaber, M. B., & Ragnarsdottir, K. V. (2007). Crossing disciplines and scales to understand the Critical Zone. *Elements, 3,* 307–314.

Brimhall, G. H., & Dietrich, W. E. (1987). Constitutive mass balance relations between chemical composition, volume, density, porosity, and strain in metasomatic hydrochemical systems: results on weathering and pedogenesis. *Geochimica et Cosmochimica Acta, 51,* 567–587.

Brimhall, G. H., Chadwick, O. A., Lewis, C. J., Compston, W., Williams, I. S., Danti, K. J., et al. (1991a). Deformational mass in invasive processes transport and soil evolution. *Science, 255,* 695–702.

Brimhall, G. H., Lewis, C. J., Ford, C., Bratt, J., Taylor, G., & Warin, O. (1991b). Quantitative geochemical approach to pedogenesis: importance of parent material reduction, volumetric expansion, and eolian influx in laterization. *Geoderma, 51,* 51–91.

Bradley, R. S. (1999). *Paleoclimatology: Reconstructing climates of the quaternary*. San Diego: Academic Press.

Brewer, R. (1976). *Fabric and mineral analysis of soils*. New York: Robert E. Krieger Publishing Company.

Bronger, A., & Heinkele, T. (1989). Micromorphology and genesis of paleosols in the Luochuan loess section, China: pedostratigraphic and environmental implications. *Geoderma, 45,* 123–143.

Bullock, P., FeDoroff, N., Jungerius, A., Stoops, G., Tursina, T., & Babel, U. (1985). *Handbook for soil thin section description*. Wolverhampton: Waine Research Publications.

Campisano, C. J., & Feibel, C. S. (2008). Depositional environments and stratigraphic summary of the Pliocene Hadar Formation at Hadar, Afar Depression, Ethiopia. In J. Quade & J. G. Wynn (Eds.), The geology of early humans in the Horn of Africa. *Geological Society of America Special Paper, 446*, 179–201.

Cerling, T. E. (1984). The stable isotopic composition of modern soil carbonate and its relationship to climate. *Earth and Planetary Science Letters, 71*, 229–240.

Cerling, T. E. (2014). Stable isotope evidence for hominin environments in Africa. In T. E. Cerling (Ed.), *Treatise on geochemistry, Vol. 14: Archaeology and anthropology* (pp. 157–67). Oxford: Pergamon.

Cerling, T., & Quade, J. (1993). Stable carbon and oxygen isotopes in soil carbonates. *Geophysical Monograph Series 78.*

Cerling, T. E., & Hay, R. L. (1986). An isotopic study of paleosol carbonates from Olduvai Gorge. *Quaternary Research, 25,* 63–78.

Cerling, T. E., Bowman, J. R., & O'Neil, J. R. (1988). An isotopic study of a fluvial-lacustrine sequence: the Plio-Pleistocene Koobi fora sequence, East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology, 63,* 335–356.

Cerling, T. E., Harris, J. M., & Passey, B. H. (2003). Diets of East African bovidae based on stable isotope analysis. *Journal of Mammalogy, 84,* 456–470.

Cerling, T. E., Levin, N. E., Quade, J., Wynn, J. G., Fox, D. L., Kingston, J. D., et al. (2010). Comment on the paleoenvironment of *Ardipithecus ramidus*. *Science, 328,* 1105.

Cerling, T. E., Wynn, J. G., Andanje, Sa, Bird, M. I., Korir, D. K., Levin, N. E., et al. (2011). Woody cover and hominin environments in the past 6 million years. *Nature, 476,* 51–56.

Chadwick, O. A., & Chorover, J. (2001). The chemistry of pedogenic thresholds. *Geoderma, 100,* 231–353.

Chadwick, O. A., Brimhall, G. H., & Hendricks, D. M. (1991). From a black to a gray box—a mass balance interpretation of pedogenesis. *Geomorphology, 3,* 369–390.

Chen, S. T., Smith, S. Y., Sheldon, N. D., & Strömberg, C. (2015). Regional-scale variability in the spread of grasslands in the late Miocene. *Palaeogeography, Palaeoclimatology, Palaeoecology, 437,* 42–52.

Cleveland, D. M., Atchley, S. C., & Nordt, L. C. (2007). Continental sequence stratigraphy of the Upper Triassic (Norian Rhaetian) Chinle strata, northern New Mexico, U.S.A.: allocyclic and autocyclic origins of paleosol-bearing alluvial successions. *Journal of Sedimentary Research, 77,* 909–924.

Cleveland, D. M., Nordt, L. C., Dworkin, S. I., & Atchley, S. C. (2008). Pedogenic carbonate isotopes as evidence for extreme climatic events preceding the Triassic-Jurassic boundary: implications for the biotic crisis? *Geological Society of America Bulletin, 120,* 1408–1415.

Clyde, W. C., Gingerich, P. D., Wing, S. L., Rohl, U., Westerhold, T., Bowen, G., et al. (2013). Bighorn Basin Coring Project (BBCP): a continental perspective on early Paleogene hyperthermals. *Scientific Drilling*, 21–31.

Cotton, J. M., & Sheldon, N. D. (2012). High-resolution isotopic record of C_4 photosynthesis in a Miocene grassland. Palaeogeography, Palaeoclimatology, Palaeoecology, 337–338, 88–98.

Cremeens, D. L., Hart, J. P., & Darmody, R. G. (1998). Complex pedostratigraphy of a terrace fragipan at the Memorial Park site, central Pennsylvania. *Geoarchaeology, 13,* 339–359.

Driese, S., & Foreman, J. L. (1992). Paleopedology and paleoclimatic implications of Late Ordovician vertic paleosols, Juniata Formation, southern Appalachians. *Journal of Sedimentology Research, 62,* 71–83.

Driese, S. G., & Mora, C. I. (1993). Physio-chemical environment of pedogenic carbonate formation in Devonian vertic paleosols, central Appalachians, USA. *Sedimentology, 40,* 199–216.

Driese, S. G., Mora, C. I., Cotter, E., & Foreman, J. L. (1992). Paleopedology and stable isotope chemistry of Late Silurian vertic Paleosols, Bloomsburg formation, central Pennsylvania. *Journal of Sedimentary Research, 62*(5), 825–841.

Driese, S. G., Mora, C. I., Stiles, C. A., Joeckel, R. M., & Nordt, L. C. (2000). Mass-balance reconstruction of a modern Vertisol: implications for interpreting the geochemistry and burial alteration of paleo-Vertisols. *Geoderma, 95,* 179–204.

Driese, S. G., Li, Z.-H., & Horn, S. P. (2005). Late Pleistocene and Holocene climate and geomorphic histories as interpreted from a 23,000 ^{14}C yr B.P. paleosol and floodplain soils, southeastern West Virginia, USA. *Quaternary Research, 63,* 136–149.

Driese, S. G., Peppe, D. J., Beverly, E. J., DiPietro, L. M., Arellano, L. N., & Lehmann, T. (2016). Paleosols and paleoenvironments of the early Miocene deposits near Karungu, Lake Victoria, Kenya. *Palaeogeography, Palaeoclimatology, Palaeoecology, 443,* 167–182.

Driese, S. G., & Ober, E. G. (2005). Paleopedologic and paleohydrologic records of precipitation seasonality from early Pennsylvanian “underclay” Paleosols, USA. *Journal of Sedimentary Research, 75,* 997–1010.

Driese, S. G., Medaris Jr, L. G., Ren, M., Runkel, A. C., & Langford, R. P. (2007). Differentiating pedogenesis from diagenesis in early terrestrial paleoweathering surfaces formed on granitic composition parent materials. *The Journal of Geology, 115*(4), 387–406.

Driese, S. G., Jirsa, M. A., Ren, M., Brantley, S. L., Sheldon, N. D., Parker, D., et al. (2011). Neoarchean paleoweathering of tonalite and metabasalt: implications for reconstructions of 2.69 Ga early terrestrial ecosystems and paleoatmospheric chemistry. *Precambrian Research, 189,* 1–17.

Driese, S. G., & Ashley, G. M. (2016). Paleoenvironmental reconstruction of a paleosol catena, the Zinj archeological level, Olduvai Gorge, Tanzania. *Quaternary Research*, *85*, 133–146.

Dworkin, S. I., Nordt, L., & Atchley, S. (2005). Determining terrestrial paleotemperatures using the oxygen isotopic composition of pedogenic carbonate. *Earth and Planetary Science Letters, 237,* 56–68.

Eidt, R. C. (1985). Theoretical and practical considerations in the analysis of anthrosols. In G. Rapp & J. A. Gifford (Eds.), *Archaeological geology* (pp. 155–190). New Haven: Yale University Press.

Fedo, C. M., Nesbitt, H. W., & Young, G. M. (1995). Unraveling the effects of potassium metasomatism in sedimentary rocks and paleosols, with implications for paleoweathering conditions and provenance. *Geology, 23,* 921–924.

Fedoroff, N., & Goldberg, P. (1982). Comparative micromorphology of two late Pleistocene paleosols (in the Paris Basin). *Catena, 9,* 227–251.

Fitzpatrick, E. A. (1984). *Micromorphology of soils*. London and New York: Chapman and Hall.

Fitzpatrick, E. A. (1993). *Soil microscopy and micromorphology*. New York: Wiley.

Ferring, C. R. (1992). Alluvial pedology and geoarchaeological research. In V. T. Holliday (Ed.), *Soils in archaeology* (pp. 1–40). Washington D. C., Smithsonian Institution Press.

Follmer, L. R. (1998). Preface. In L. R. Follmer, D. L. Johnson, & J. A. Catt (Eds), *Revisitation of concepts in paleopedology: Transactions of the Second International Symposium on Paleopedology. Quaternary International* (vol. 51/52, pp. 1–3).

Gallagher, T. M., & Sheldon, N. D. (2013). A new paleothermometer for forest paleosols and its implications for Cenozoic climate. *Geology, 41,* 647–650.

Gallagher, T. M., & Sheldon, N. D. (2016). Combining soil water balance and clumped isotopes to understand the nature and timing of pedogenic carbonate formation. *Chemical Geology 435*, 79–91.

Garrett, N. D., Fox, D. L., McNulty, K. P., Tryon, C. A., Faith, J. T., Peppe, D. J., et al. (2015). Stable isotope paleoecology of Late Pleistocene middle stone age humans from the equatorial East Africa, Lake Victoria basin, Kenya. *Journal of Human Evolution, 82,* 1–14.

Ghosh, P., Adkins, J., Affek, H., Balta, B., Guo, W., Schauble, E., et al. (2006). 13C–18O bonds in carbonate minerals: a new kind of paleothermometer. *Geochimica et Cosmochimica Acta, 70,* 1439–1456.

Gile, L. H. (1979). Holocene soils in eolian sediments of Baily County, Texas. *Soil Science Society of America Journal, 43,* 994–1003.

Gile, L. H., Peterson, F. F., & Grossman, R. B. (1966). Morphological and genetic sequences of carbonate accumulation in desert soils. *Soil Science, 101*.

Gulbranson, E. L., Montanez, I. P., & Tabor, N. J. (2011). A proxy for humidity and floral province from paleosols. *Journal of Geology, 119,* 559–573.

Guthrie, R. L., & Witty, J. E. (1982). New designations for soil horizons and layers and the new Soil Survey Manual. *Soil Science Society of America Journal, 46*(2), 443–444.

Han, J. T., Fyfe, W. S., & Longstaffe, F. J. (1998). Climatic implications of the S5 paleosol complex on the southernmost Chinese Loess Plateau. *Quaternary Research, 50,* 21–33.

Harnois, L. (1988). The CIW index: a new chemical index of weathering. *Sedimentary Geology, 55,* 319–322.

Harris, W., & White, N. (2008). X-ray diffraction techniques for soil mineral identification. *Methods of soil analysis part 5—Mineralogical methods*. Madison: Soil Science Society of America.

Hasiotis, S. T. (2007). Continental ichnology: fundamental processes and controls on trace-fossil distribution. In W. Miller III (Ed.), *Trace fossils—Concepts, problems, prospects* (pp. 268–284). Elsevier Press.

Hasiotis, S. T., Platt, B. F., Hembree, D.I., & Everhart, M. (2007a). The trace-fossil record of vertebrates. In W. Miller III (Ed.), *Trace fossils—Concepts, problems, prospects* (pp. 196–218). Elsevier Press.

Hasiotis, S. T., Kraus, M. J., & Demko, T. M. (2007b). Climate controls on continental trace fossils. In W. Miller III (Ed.), *Trace fossils—Concepts, problems, prospects* (pp. 172–195). Elsevier Press.

Hasiotis, S. T., & Honey, J. G. (2000). Paleohydrologic and stratigraphic significant of crayfish burrows in continental deposits: examples from several Paleocene Laramide basins in the Rocky Mountains. *Journal of Sedimentary Research, 70,* 127–139.

Hembree, D. I., Platt, B. F. & Smith, J. J. (Eds.) (2014). *Experimental approaches to understanding fossil organisms: Lessons from the Living*. Netherlands: Springer.

Henkes, G. A., Passey, B. H., Grossman, E. L., Shenton, B. J., Perez-Huerta, A., & Yancey, T. E. (2014). Temperature limits for preservation of primary calcite clumped isotope paleotemperatures. *Geochimica et Cosmochimica Acta, 139,* 362–382.

Holliday, V. T. (2004). *Soils in archaeological research*. Oxford: Oxford University Press.

Holliday, V. T. (2006). A history of soil geomorphology in the United States. In B. P. Warkentin (Ed.), *Footprints in the soil: People and ideas in soil history* (pp. 187–254). Amsterdam: Elsevier.

Holliday, V. T., & Gartner, W. G. (2007). Soil phosphorus and archaeology: a review and comparison of methods. *Journal of Archaeological Science, 34,* 301–333.

Holloway, R. L., Broadfield, D. C., & Yuan, M. S. (2004). *The human fossil record Vol. 3*. Wiley.

Hover, V. C., & Ashley, G. M. (2003). Geochemical signatures of paleodepositional and diagenetic environments: a STEM/AEM study of authigenic clay minerals from an arid rift basin, Olduvai Gorge, Tanzania. *Clays and Clay Minerals, 51,* 231–251.

Huggett, R. J. (1998). Soil chronosequences, soil development, and soil evolution: a critical review. *Catena, 32,* 155–172.

Hyland, E. G., Sheldon, N. D., Van der Voo, R., Badgley, C., & Abrajevitch, A. (2015). A new paleoprecipitation proxy based on soil magnetic properties: implications for expanding paleoclimate reconstructions. *Geological Society of America Bulletin*.

Jenny, H. (1941). *Factors of soil formation: a system of quantitative pedology*. McGraw-Hill book company, Inc.

Jenny, H. (1980). *The soil resource. Origin and behavior*. Berlin: Springer.

Kemp, R. A. (1999). Micromorphology of loess-paleosol sequences: a record of paleoenvironmental change. *Catena, 35,* 179–196.
Kraus, M. J. (1987). Integration of channel and floodplain suites in aggrading fluvial systems, II. Vertical relations of alluvial paleosols. *Journal of Sedimentary Petrology, 57*(4), 602–612.
Kraus, M. J. (1997). Lower Eocene alluvial paleosols: pedogenic development, stratigraphic relationships, and paleosol/landscape associations. *Palaeogeography, Palaeoclimatology, Palaeoecology, 129,* 387–406.
Kraus, M. J. (1999). Paleosols in clastic sedimentary rocks: their geologic applications. *Earth-Science Review, 47,* 41–70.
Kraus, M. J., & Brown, T. M. (1988). Pedofacies analysis; a new approach to reconstructing ancient fluvial sequences. In J. Reinhard & W. R. Sigleo (Eds.), *Paleosols and weathering through geologic time: Principles and applications* (Geological Society of America Special Paper 216) (pp. 143–152). Denver: Geological Society of America.
Kraus, M. J., & Hasiotis, S. T. (2006). Significance of different modes of rhizolith preservation to interpreting paleoenvironmental and paleohydrologic settings: examples from Paleogene paleosols, Bighorn Basin, Wyoming, U.S.A. *Journal of Sedimentary Research, 76,* 633–646.
Kubiena, W. (1970). Micromorphology of polygenetic soils and paleosoils in polar regions. *Annales de Edafologia y Abrobiologia*, 845–856.
Leighton, M. M. (1937). The significance of profiles of weathering in stratigraphic archaeology. In G. G. MacCurdy (Ed.), *Early Man* (pp. 163–172). New York: Lippincott.
Leopold, M., Völkel, J., Dethier, D., Huber, J., & Steffens, M. (2011). Characteristics of a paleosol and its implication for the Critical Zone development, Rocky Mountain Front Range of Colorado, USA. *Applied Geochemistry, 26,* S72–S75.
Lepre, C. J., Quinn, R. L., Joordens, J. C. a, Swisher, C. C., & Feibel, C. S. (2007). Plio-Pleistocene facies environments from the KBS Member, Koobi Fora Formation: implications for climate controls on the development of lake-margin hominin habitats in the northeast Turkana Basin (northwest Kenya). *Journal of Human Evolution, 53*, 504–514.
Levin, N. E. (2015). Environment and climate of early human evolution. *Annual Review of Earth and Planetary Sciences, 43,* 405–429.
Levin, N. E., Cerling, T. E., Passey, B. H., Harris, J. M., & Ehleringer, J. R. (2006). A stable isotope aridity index for terrestrial environments. *Proceedings of the National Academy of Sciences, USA, 103,* 11201–11205.
Levin, N. E., Quade, J., Simpson, S. W., Semaw, S., & Rogers, M. J. (2004). Isotopic evidence for Plio-Pleistocene environmental change at Gona, Ethiopia. *Earth and Planetary Science Letters, 219,* 93–110.
Levin, N., Brown, F. H., Behrensmeyer, A. K., Bobe, R., & Cerling, T. E. (2011). Paleosol carbonates from the Omo Group: isotopic records of local and regional environmental change in East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology, 307,* 75–89.
Lüdecke, T., Schrenk, F., Thiemeyer, H., Kullmer, O., Bromage, T. G., Sandrock, O., et al. (2016). Persistent C3 vegetation accompanied Plio-Pleistocene hominin evolution in the Malawi Rift (Chiwondo Beds, Malawi). *Journal of Human Evolution, 90,* 163–175.
Ludvigson, G. A., Gonzalez, L. A., Fowle, D. A., Roberts, J. A., Driese, S. G., Villarreal, M. A., et al. (2013). Paleoclimatic applications and modern process studies of pedogenic siderite. In S. G. Driese & L. C. Nordt (Eds.), *New frontiers in paleopedology and terrestrial paleoclimatology* (SEPM Special Publication No. 104) (pp. 47–62). Tulsa: SEPM.
Ludvigson, G. A., Gonzalez, L. A., Metzger, R. A., Witzke, B. J., Brenner, R. L., Murillo, A. P., et al. (1998). Meteoric sphaerosiderite lines and their use for paleohydrology and paleoclimatology. *Geology, 26,* 1039–1042.
Lüdecke, T., Schrenk, F., Thiemeyer, H., Kullmer, O., Bromage, T. G., Sandrock, O., et al. (2016b). Persistent C3 vegetation accompanied Plio-Pleistocene hominin evolution in the Malawi Rift (Chiwondo Beds, Malawi). *Journal of Human Evolution, 90,* 163–175.
Lukens, W. E., Driese, S. G., Peppe, D. J., & Loudermilk, M. (2017a). Sedimentology, stratigraphy, and paleoclimate at the late Miocene Coffee Ranch fossil site in the Texas Panhandle. *Palaeogeography, Palaeoclimatology, Palaeoecology, 48*5, 361–376.
Lukens, W. E., Lehmann, T., Peppe, D. J., Fox, D. L., Driese, S. G., & McNulty, K. P. (2017b). The early Miocene Critical Zone at Karungu, Western Kenya: an equatorial, open habitat with few primate remains. *Frontiers in Earth Science, 5*, 87.
Lukens, W. E., Nordt, L. C., Stinchcomb, G. E., Driese, S. G., & Tubbs, J. D. (2018). Reconstructing pH of Paleosols Using Geochemical Proxies. *The Journal of Geology, 126*(4), 427–449.
Mack, G. H., James, W. C., & Monger, H. C. (1993). Classification of paleosols. *Geological Society of America Bulletin, 105,* 129–136.
Machette, M. N. (1985). Calcic soils of the southwestern United States. In *Geological Society of America Special Paper 203* (pp. 1–21).
Machel, H. G., Mason, R. A., Mariano, A. N., & Mucci, A. (1991). Causes and emission of luminescence in calcite and dolomite. In C. E. Barker and O. C. Kopp (Eds.), *Luminescence microscopy and spectroscopy: Quantitative and qualitative applications (SC25)* (pp. 9–25). Tulsa: SEPM.
Magill, C. R., Ashley, G. M., & Freeman, K. H. (2013a). Ecosystem variability and early human habitats in eastern Africa. *Proceedings of the National Academy of Sciences, USA, 110*, 1167–1174.
Magill, C. R., Ashley, G. M., & Freeman, K. H. (2013b). Water, plants, and early human habitats in eastern Africa. *Proceedings of the National Academy of Sciences, USA, 110,* 1175–1180.
Maher, B. A. (1998). Magnetic properties of modern soils and Quaternary loessic paleosols: paleoclimatic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology, 137,* 25–54.
Maher, B. A., & Thompson, R. (1995). Paleorainfall reconstructions from pedogenic magnetic susceptibility variations in the Chinese loess and paleosols. *Quaternary Research, 44,* 383–391.
Marin-Spiotta, E., Chaopricha, N. T., Plante, A. F., Diefendorf, A. F., Mueller, C. W., Grandy, A. S., et al. (2014). Long-term stabilization of deep soil carbon by fire and burial during early Holocene climate change. *Nature Geosci, 7,* 428–432.
Maxbauer, D. P., Feinberg, J. M., & Fox, D. L. (2016a). Magnetic mineral assemblages in soils and paleosols as the basis for paleoprecipitation proxies: a review of magnetic methods and challenges. *Earth-Science Review, 155,* 28–48.
Maxbauer, D. P., Feinberg, J. M., Fox, D. L., & Clyde, W. C. (2016b). Magnetic minerals as recorders of weathering, diagenesis, and paleoclimate: a core-outcrop comparison of Paleocene-Eocene paleosols in the Bighorn Basin, WY, USA. *Earth and Planetary Science Letters, 452,* 15–26.
Maynard, J. B. (1992). Chemistry of modern soils as a guide to interpreting Precambrian paleosols. *Journal of Geology, 100*, 279–289.
Medaris Jr, L. G., Driese, S. G., & Stinchcomb, G. E. (2017). The Paleoproterozoic Baraboo paleosol revisited: quantifying mass fluxes of weathering and metasomatism, chemical climofunctions, and atmospheric pCO_2 in a chemically heterogeneous protolith. *Precambrian Research, 301*, 179–194.
Mentzer, S. M. (2014). Microarchaeological approaches to the identification and interpretation of combustion features in prehistoric archaeological sites. *Journal of Archaeoogical Method and Theory, 21,* 616–668.
Michel, L. A., Driese, S. G., Nordt, L. C., Breecker, D. O., Labotka, D. M., & Dworkin, S. I. (2013). Stable-Isotope geochemistry of

Vertisols formed on marine limestone and implications for deep-time paleoenvironmental reconstructions. *Journal of Sedimentary Research, 83,* 300–308.

Michel, L. A., Peppe, D. J., Lutz, Ja, Driese, S. G., Dunsworth, H. M., Harcourt-Smith, W. E. H., et al. (2014). Remnants of an ancient forest provide ecological context for Early Miocene fossil apes. *Nature Communications, 5,* 1–9.

Mintz, J. S., Driese, S. G., Breecker, D. O., & Ludvigson, G. A. (2011). Influence of changing hydrology on pedogenic calcite precipitation in Vertisols, Dance Bayou, Brazoria County, Texas, U.S.A.: implications for estimating paleoatmospheric pCO_2. *Journal of Sedimentary Research, 81,* 394–400.

Moore, D. M., & Reynolds, R. C. (1997). *X-Ray diffraction and the identification and analysis of clay minerals*. New York: Oxford University Press.

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.

Morrison, R. B. (1967). Principles of Quaternary soil stratigraphy. In R. B. Morrison & H. E. Wright (Eds.), *Quaternary soils* (pp. 1–69). Reno: University of Nevada Desert Research Institute, Center for Water Resources Research.

Myers, T. S., Tabor, N. J., & Rosenau, N. A. (2014). Multiproxy approach reveals evidence of highly variable paleoprecipitation in the Upper Jurassic Morrison Formation (western United States). *Geological Society of American Bulletin, 126,* 1105–1116.

National Research Council. (2001). *Basic research opportunities in Earth science*. Washington, D. C.: National Academies Press.

National Research Council. (2010). *Understanding climate's influence on human evolution*. Washington, D. C.: National Academies Press.

Nettleton, W. D., Olson, C. G., & Wysocki, D. A. (2000). Paleosol classification: problems and solutions. *Catena, 41,* 61–92.

Nesbitt, H. W., & Young, G. M. (1982). Earth Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature, 299,* 715–717.

Nordt, L. C. (2001). Stable C and O isotopes in soils: applications for archaeological research. In P. Goldberg, V. Holliday & R. Ferring (Eds.), *Earth-sciences and archaeology* (pp. 419–445).

Nordt, L. C., & Driese, S. D. (2010a). New weathering index improves paleorainfall estimates from Vertisols. *Geology, 38,* 407–410.

Nordt, L. C., & Driese, S. G. (2010b). A modern soil characterization approach to reconstructing physical and chemical properties of paleo-Vertisols. *American Journal of Science, 310,* 37–64.

Nordt, L. C., & Driese, S. G. (2013). Application of the Critical Zone concept to the deep-time sedimentary record. *The Sedimentary Record, 11*, 4–9.

Nordt, L. C., Dworkin, S. I., & Atchley, S. C. (2011). Ecosystem response to soil biogeochemical behavior during the Late Cretaceous and early Paleocene within the western interior of North America. *Geological Society of America Bulletin, 123,* 1745–1762.

Nordt, L. C., Hallmark, C. T., Driese, S. G., Dworkin, S. I., & Atchley, S. C. (2012). Biogeochemical characterization of a lithified paleosol: implications for the interpretation of ancient Critical Zones. *Geochimica et Cosmochimica Acta, 87,* 267–282.

Nordt, L. C., Hallmark, C. T., Driese, S. G., & Dworkin, S. I. (2013). Multianalytical pedosystem approach to characterizing and interpreting the fossil record of soils. In S. G. Driese & L. C. Nordt (Eds.), *New frontiers in paleopedology and terrestrial paleoclimatology* (SEPM Special Publication No. 104) (pp. 47–62, 89–108). Tulsa: SEPM.

Nordt, L., Orosz, M., Driese, S., & Tubbs, J. (2006). Vertisol carbonate properties in relation to mean annual precipitation: implications for paleoprecipitation estimates. *Journal of Geology, 114,* 501–510.

Nordt, L. C., Wilding, L. P., Lynn, W. C., & Crawford, C. C. (2004). Vertisol genesis in a humid climate of the coastal plain of Texas, U.S.A. *Geoderma, 122,* 83–102.

Oerter, E. J., Sharp, W. D., Oster, J. L., Ebeling, A., Valley, J. W., Kozdon, R., et al. (2016). Pedothem carbonates reveal anomalous North American atmospheric circulation 70,000–55,000 years ago. *Proceedings of the National Academy of Sciences, USA, 113*, 919–924.

Passey, B. H., Levin, N. E., Cerling, T. E., Brown, F. H., & Eiler, J. M. (2010). High-temperature environments of human evolution in East Africa based on bond ordering in paleosol carbonates. *Proceedings of the National Academy of Sciences, USA, 107*, 11245–11249.

Passey, B. H., Hu, H., Ji, H., Montanari, S., Li, S., Henkes, G. A., et al. (2014). Triple oxygen isotopes in biogenic and sedimentary carbonates. *Geochimica et Cosmochimica Acta, 141,* 1–25.

Pettijohn, F. J., Potter, P. E., & Siever, R. (1987). *Sand and sandstone* (2nd ed.). Springer-Verlag.

Poppe, L. J., Paskevich, V. F., Hathaway, J. C., & Blackwood, D. S. (2001). A laboratory manual for X-ray powder diffraction. *US Geological Survey Open-File Report, 1*(041), 1–88.

Prochnow, S. J., Nordt, L. C., Atchley, S. C., & Hudec, M. R. (2006). Multi-proxy paleosol evidence for Middle and Late Triassic climate trends in eastern Utah. *Palaeogeography, Palaeoclimatology, Palaeoecology, 232,* 53–72.

Pimentel, N. L., Wright, V. P., & Azevedo, T. M. (1996). Distinguishing early groundwater alteration effects from pedogensis in ancient alluvial basins: examples form the Palaeogene of southern Portugal. *Sedimentary Geology, 105,* 1–10.

Qadir, M., & Schubert, S. (2002). Degradation processes and nutrient constraints in sodic soils. *Land Degradation & Development, 13*(4), 275–294.

Quade, J., Levin, N., Semaw, S., Stout, D., Renne, P., Rogers, M., et al. (2004). Paleoenvironments of the earliest stone toolmakers, Gona, Ethiopia. *Geological Society of America Bulletin, 116,* 1529.

Quade, J., Eiler, J., Daëron, M., Achyuthan, H. (2013). The clumped isotope geothermometer in soil and paleosol carbonate. *Geochimica Cosmochimica Acta, 105*, 92–107.

Quinn, R. L., Lepre, C. J., Wright, J. D., & Feibel, C. S. (2007). Paleogeographic variations of pedogenic carbonate d13C values from Koobi Fora, Kenya: implications for floral compositions of Plio-Pleistocene hominin environments. *Journal of Human Evolution, 53,* 560–573.

Quinn, R. L., Lepre, C. J., Feibel, C. S., Wright, J. D., Mortlock, R. A., Harmand, S., et al. (2013). Pedogenic carbonate stable isotopic evidence for wooded habitat preference of early Pleistocene tool makers in the Turkana Basin. *Journal of Human Evolution, 65,* 65–78.

Rasmussen, C., & Tabor, N. J. (2007). Applying a quantitative pedogenic energy model across a range of environmental gradients. *Soil Science Society of America Journal, 71*(6), 1719–1729.

Rasmussen, C., Southard, R. J., & Horwath, W. R. (2005). Modeling energy inputs to predict pedogenic environments using regional environmental databases. *Soil Science Society of America Journal, 69*(4), 1266–1274.

Rawls, W. J. (1983). Estimating soil bulk density from particle size analyses and organic matter content. *Soil Science, 135,* 123–125.

Retallack, G. J. (1983). Late Eocene and Oligocene paleosols from Badlands National Park, South Dakota. *Geological Society of America. Special Papers, 193,* 82.

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

Retallack, G. J., Wynn, J. G., Benefit, B. R., & Mccrossin, M. L. (2002). Paleosols and paleoenvironments of the middle Miocene, Maboko Formation, Kenya. *Journal of Human Evolution, 42*(6), 659–703.

Retallack, G. J. (1994). The environmental-factor approach to the interpretation of paleosols. In R. J. Luxmoore & J. M. Bartels (Eds.), *Factors of soil formation: A fiftieth anniversary retrospective* (pp. 31–64). Madison: Soil Science Society of America.

Retallack, G. J. (2001). *Soils of the past: An introduction to paleopedology* (2nd ed.). Oxford: Blackwell Science Ltd.

Retallack, G. J. (2005). Pedogenic carbonate proxies for amount and seasonality of precipitation in paleosols. *Geology, 33,* 333–336.

Retallack, G. J., James, W. C., Mack, G. H., & Monger, H. C. (1993). Classification of paleosols: discussion and reply. *Geological Society of America Bulletin, 105,* 1635–1637.

Retallack, G. J., Orr, W. N., Prothero, D. R., Duncan, R. A., Kester, P. R., & Ambers, C. P. (2004). Eocene-Oligocene extinction and paleoclimatic change near Eugene. *Oregon. Geological Society of America Bulletin, 116,* 817.

Retallack, G. J., & Huang, C. (2010). Depth to gypsic horizon as a proxy for paleoprecipitation in paleosols of sedimentary environments. *Geology, 38,* 403–406.

Richter, D. deB, & Yaalon, D. H. (2012). "The changing model of soil" revisited. *Soil Science Society of America Journal, 76*, 766–778.

Rosenau, N. A., Tabor, N. J., Elrick, S. D., & Nelson, W. J. (2013). Polygenetic history of paleosols in Middle-Upper Pennsylvanian cyclothems of the Illinois Basin, U.S.A.: Part I. Characterization of paleosol types and interpretations of pedogenic processes. *Journal of Sedimentary Research, 83,* 606–636.

Ruhe, R. V. (1965). Quaternary paleopedology. In H. E. Wright, & D. G. Frey (Eds.), The Quaternary of the United States (pp. 755–764). Princeton, NJ: Princeton University Press.

Saxton, K. E., & Rawls, W. J. (2006). Soil water characteristic estimates by texture and organic matter for hydrologic solutions. *Soil Science Society of America Journal, 70,* 1569–1578.

Schaetzl, R. J., & Thompson, M. L. (2015). *Soils*: Cambridge University Press.

Schoeneberger, P. J., Wysocki, D. A., Benham, E. C., & Broderson, W. D. (2012). *Field book for describing and sampling soils, version 3.0.* Lincoln: Natural Resources Conservation Service, National Soil Survey Center.

Schwertmann, U. (1993). Relations between iron oxides, soil color, and soil formation. In J. M. Bigham & E. J. Ciolkosz (Eds.), *Soil color* (Special Publication No. 31) (pp. 51–69). Madison: Soil Science Society of America.

Sheldon, N. (2003). Pedogenesis and geochemical alteration of the Picture Group subgroup, Columbia River basalt, Oregon. *Geological Society of America Bulletin, 115,* 1377–1387.

Sheldon, N. D. (2005). Do red beds indicate paleoclimatic conditions?: a Permian case study. *Palaeogeography, Palaeoclimatology, Palaeoecology, 228,* 305–319.

Sheldon, N. D. (2006). Precambrian paleosols and atmospheric CO_2 levels. *Precambrian Research, 147,* 148–155.

Sheldon, N. D., & Retallack, G. J. (2001). Equation for compaction of paleosols due to burial. *Geology, 29,* 247–250.

Sheldon, N. D., & Retallack, G. J. (2004). Regional paleoprecipitation records from the late eocene and oligocene of North America. *The Journal of Geology, 112,* 487–494.

Sheldon, N. D., & Tabor, N. J. (2009). Quantitative paleoenvironmental and paleoclimatic reconstruction using paleosols. *Earth-Science Review, 95,* 1–52.

Sheldon, N. D., Retallack, G. J., & Tanaka, S. (2002). Geochemical climofunctions from North American soils and application to paleosols across the Eocene-Oligocene boundary in Oregon. *Journal of Geology, 110,* 687–696.

Sikes, N. E., Potts, R., & Behrensmeyer, A. K. (1999). Early Pleistocene habitat in member 1 Olorgesailie based on paleosol stable isotopes. *Journal of Human Evolution, 37,* 721–746.

Sikes, N. E., & Ashley, G. M. (2007). Stable isotopes of pedogenic carbonates as indicators of paleoecology in the Plio-Pleistocene (upper Bed I), western margin of the Olduvai Basin, Tanzania. *Journal of Human Evolution, 53*, 574–594.

Snell, K. E., Thrasher, B. L., Eiler, J. M., Koch, P. L., Sloan, L. C., & Tabor, N. J. (2013). Hot summers in the Bighorn Basin during the early Paleogene. *Geology, 41,* 55–58.

Soil Classification Working Group. (1998). *The Canadian system of soil classification* (3rd ed., Agriculture and Agri-Food Canada Publication 1646). Ottawa: NRC Research Press.

Soil Survey Staff. (2006). *Keys to soil taxonomy* (10th ed.). Washington, D. C.: United States Department of Agriculture Natural Resources Conservation Service.

Soil Survey Staff. (2014a). *Illustrated guide to soil taxonomy* (1.0 ed.). Lincoln: United States Department of Agriculture Natural Resources Conservation Service.

Soil Survey Staff. (2014b). *Kellogg soil survey laboratory methods manual, Version 5.0*. (Soil Survey Investigations Report No. 42). Lincoln: United States Department of Agriculture Natural Resources Conservation Service.

Stiles, C. A., Mora, C. I., & Driese, S. G. (2001). Pedogenic iron-manganese nodules in Vertisols: a new proxy for paleoprecipitation? *Geology, 29,* 943–946.

Stiles, C. A., Mora, C. I., & Driese, S. G. (2003a). Pedogenic processes and domain boundaries in a Vertisol climosequence: evidence from titanium and zirconium distribution and morphology. *Geoderma, 116,* 279–299.

Stiles, C. A., Mora, C. I., Driese, S. G., & Robinson, A. C. (2003b). Distinguishing climate and time in the soil record: mass-balance trends in Vertisols from the Texas coastal prairie. *Geology, 31,* 331–334.

Stinchcomb, G. E., Driese, S. G., Nordt, L. C., DiPietro, L., & Messner, T. C. (2014). Early Holocene soil cryoturbation in northeastern USA: implications for archaeological site formation. *Quaternary International, 342,* 186–198.

Stinchcomb, G. E., Nordt, L. C., Driese, S. G., Lukens, W. E., Williamson, F. C., & Tubbs, J. D. (2016). A data-driven spline model designed to predict paleoclimate using paleosol geochemistry. *American Journal of Science, 316,* 746–777.

Stoops, G. (2003). Guidelines for analysis and description of soil and regolith thin sections. Madison: Soil Science Society of America, Inc.

Stoops, G., Marcelino, V., & Mees, F. (Eds.) (2010). *Interpretation of micromorphological features of soils and regoliths, 1st ed.* Netherlands: Elsevier.

Tabor, N. J., & Montañez, I. P. (2002). Shifts in late Paleozoic atmospheric circulation over western equatorial Pangea: insights from pedogenic mineral $\delta^{18}O$ compositions. *Geology, 30,* 1127–1130.

Tabor, N. J., & Myers, T. S. (2015). Paleosols as indicators of paleoenvironment and paleoclimate. *Annual Review of Earth and Planetary Science, 43*, 11.1–11.29.

Tabor, N. J., Montañez, I. P., Steiner, M. B., & Schwindt, D. (2007). $\delta^{13}C$ values of carbonate nodules across the Permian-Triassic boundary in the Karoo Supergroup (South Africa) reflect a stinking sulfurous swamp, not atmospheric CO_2. *Palaeogeography, Palaeoclimatology, Palaeoecology, 252,* 370–381.

Terry, D. O. (2001). Paleopedology of the Chadron Formation of northwestern Nebraska: implications for paleoclimatic change in the North America midcontinent across the Eocene-Oligocene boundary. *Palaeogeography, Palaeoclimatology, Palaeoecology, 168,* 1–38.

Torres, M. A., & Gaines, R. R. (2013). Paleoenvironmental and paleoclimatic interpretations of the late Paleocene Golder

Formations, southern California, U.S.A., based on paleosol geochemistry. *Journal of Sedimentary Research, 83,* 591–605.

Trendell, A. M., Nordt, L. C., Atchley, S. C., LeBlanc, S. L., & Dworkin, S. I. (2013a). Determining floodplain plant distributions and populations using paleopedology and fossil root traces: upper Triassic Sonsela Member of the Chinle Formation at Petrified Forest National Park, Arizona. *PALAIOS, 28,* 471–490.

Trendell, A. M., Atchley, S. C., & Nordt, L. C. (2013b). Facies analysis of a probable large-fluvial-fan depositional system: the Upper Triassic Chinle Formation at Petrified Forest National Park, Arizona, U.S.A. *Journal of Sedimentary Research, 83,* 873–895.

Ufnar, D. F., Ludvigson, G. A., González, L. A., Brenner, R. L., & Witzke, B. J. (2004). High latitude meteoric $\delta^{18}O$ compositions: Paleosol siderite in the Middle Cretaceous Nanushuk Formation, North Slope. *Alaska. Geological Society of America Bulletin, 116* (3/4), 463–473.

Ufnar, D. (2007). Clay coatings from a modern soil chronosequence: a tool for estimating the relative age of well-drained paleosols. *Geoderma, 141,* 181–200.

Valentine, K. W. G., & Dalrymple, J. B. (1976). Quaternary buried paleosols: a critical review. *Quaternary Research, 6,* 209–222.

Vepraskas, M. J. (1992). Redoximorphic features for identifying aquic conditions. *North Carolina State University Technical Bulletin 301.* Raleigh: North Carolina Agricultural Research Service.

Vepraskas, M. J. (2001). Morphological features of seasonally reduced soils. *Wetland soils: Genesis, hydrology, landscapes, and classification* (pp. 163–182). New York: Lewis Publishers.

Vepraskas, M. J., & Faulkner, S. P. (2001). Redox chemistry of hydric soils. In J. L. Richardson & M. J. Vepraskas (Eds.), *Wetland soils: Genesis, hydrology, landscapes, and classification* (pp. 85–105). New York: Lewis Publishers.

Waters, M. R. (1992). *Principles of geoarchaeology: A North American perspective*. Tucson: University of Arizona Press.

Waters, M. R., Forman, S. L., Jennings, T. A., Nordt, L. C., Driese, S. G., Feinberg, J. M., et al. (2011). The Buttermilk Creek Complex and the origins of Clovis at the Debra L. Friedkin Site. *Texas. Science, 331,* 1599–1603.

Weider, M., & Yaalon, D. H. (1982). Micromorphological fabrics and developmental stages of carbonate nodular forms related to soil characteristics. *Geoderma, 28*, 203–220.

White, T. D., Asfaw, B., Beyene, Y., Haile-Selassie, Y., Lovejoy, C. O., Suwa, G., et al. (2009). *Ardipithecus ramidus* and the paleobiology of early hominids. *Science, 326*(64), 75–86.

WoldeGabriel, G., Ambrose, S. H., Barboni, D., Bonnefille, R., Bremond, L., Currie, B., et al. (2009). The geological, isotopic, botanical, invertebrate, and lower vertebrate surroundings of *Ardipithecus ramidus*. *Science, 326*, 65, 65e1–65e5.

Wynn, J. G. (2000). Paleosols, stable carbon isotopes, and paleoenvironmental interpretation of Kanapoi, Northern Kenya. *Journal of Human Evolution, 39,* 411–432.

Wynn, J. G. (2004). Influence of Plio-Pleistocene aridification on human evolution: evidence from paleosols of the Turkana Basin, Kenya. *American Journal of Physical Anthropoloy, 123,* 106–118.

Wynn, J. G. (2007). Carbon isotope fractionation during decomposition of organic matter in soils and paleosols: implications for paleoecological interpretations of paleosols. *Palaeogeography, Palaeoclimatology, Palaeoecology, 251,* 437–448.

Wynn, J. G., & Feibel, C. S. (1995). Paleoclimatic implications of Vertisols within the Koobi Fora Formation, Turkana Basin, Northern Kenya. *Journal of Undergraduate Research, 6,* 34–42.

Yaalon, D.H., International Society of Soil Science, International Union for Quaternary Research. (1971). *Paleopedology: origin, nature, and dating of paleosols*. Jerusalem: International Society of Soil Science.

Zamanian, K., Pustovoytov, K., & Kuzyakov, Y. (2016). Pedogenic carbonates: forms and formation processes. *Earth-Science Reviews, 157*, 1–17.