19. PALEOSOLS AND CHANGES IN CLIMATE AND VEGETATION ACROSS THE EOCENE/ OLIGOCENE BOUNDARY

by Gregory J. Retallack

ABSTRACT

Fossil soils in alluvial sequences, like those of Badlands National Park, South Dakota, can be evidence for changes in climate and vegetation during Tertiary geological time, supplementing other geological records of paleoclimate, such as eolian sediments, fossil leaves and the isotopic composition of foraminifera. In the Badlands of South Dakota, the depth of calcic horizons in paleosols when compared with depths in soils of the present Great Plains of North America, indicates late Eocene (38 Ma) mean annual rainfall in excess of 1000 mm, declining by Oligocene time (32 Ma) to 500-900 mm, then later in that epoch to 450-500 mm by 30.5 Ma and to 250-450 mm by 29.5 Ma. Drier climate of late Oligocene time also may explain the decreased abundance of kaolinite and increased abundance of smectite and then illite in paleosols higher in the sequence, as well as changes in hue from red to brown and then yellow, and changes in texture from clayey to silty. Climatic deterioration across the Eocene/ Oligocene boundary is prominently displayed in the appearance of these colorfully banded badlands.

Although few fossil plants are preserved in the Badlands of South Dakota, the paleosols contain abundant drab-haloed root traces and their profile form and soil structures also are evidence of changing vegetation: moist forests of 38 Ma giving way to dry forests by 34 Ma, to dry woodland by 33 Ma, to wooded grassland with streamside gallery woodland by 32 Ma and large areas of open grassland by 30 Ma. Some changes in assemblages of fossil mammals of Badlands National Park can be related to the progressively drier conditions and more open vegetation. The long term evolutionary trend toward the kinds of cursorial limbs and high crowned teeth well displayed in modern mammalian faunas of grasslands was initiated during this time, but in these respects Eocene and Oligocene faunas remained more like modern faunas of woodland than those of grassland. Mid-Tertiary paleoclimatic cooling and drying was an impetus for evolution of the grassland biome in continental interiors.

Local volcanism and tectonic uplift in the Black Hills and Rocky Mountains to the west also have affected this record in paleosols of mid-Tertiary paleoclimatic change, but not to the extent of obscuring it. A more critical limitation on this sequence as a paleoclimatic record is non-deposition during long periods of dry climate, as indicated by abrupt changes in the nature, development and spacing of paleosols. These limitations may be overcome when more is learned about other mid-Tertiary sequences of paleosols in Oregon, Argentina, Egypt, Mongolia and Kazakhstan.

INTRODUCTION

The demonstration by Dokuchaev (1883) that soil types such as the Russian Chernozem (or Mollisols in modern terminology) were similar over many different kinds of parent rocks marked the liberation of soil science as a discipline, independent of geology and agronomy. In this example and others gathered by this pioneering Soviet school of soil geography (Glinka, 1931), climate and vegetation were seen as more important factors in determining the nature of soils, than parent material, topographic setting or time for formation. These other factors are known to be significant, even limiting in certain cases, so that each factor must be considered in trying to understand how a particular soil may have formed (Jenny, 1941). There is now a great body of published information on how surface soils form (Birkeland, 1984). This can be applied to understanding the paleoenvironmental significance of soils buried in sedimentary sequences, including many sequences of paleosols across the Eocene/Oligocene boundary.

Late Eocene and Oligocene terrestrial sequences with mammalian fossils have been known for more than a century, but paleosols in them have been reported only in the last few decades (Schultz et al., 1955; Pomerol, 1964; Pettyjohn, 1966; Morand et al., 1968). The Badlands of South Dakota are typical of such paleosol sequences: subtly banded, colorfully variegated, red and green, bioturbated, clayey rocks. Paleosols in one part of the Badlands of South Dakota have now been studied in detail (Retallack, 1983a,b), and feature prominently in this chapter.

In addition to such red and variegated beds, paleosols also are common in coal measures that accumulated in swamps and marshes (Fastovsky and McSweeney, 1987). However, these are less useful guides to ancient climate because their formation is dominated by waterlogging that isolates them from climatic influences. Paleosols also are found in many sequences of evaporites. Precipitation of salts in exposed tidal flats and playa lakes results in a kind of soil called a Salorthid (of Soil Survey Staff, 1975), and is a very different process than the accumulation of salts within lake or ocean water (Warren, 1989). Because so little research has been reported on Eocene and Oligocene paleosols beyond the Badlands of South Dakota, a preliminary assessment of paleosols in many parts of the world can only be made by attempting to read between the lines of existing geological reports on red beds, coal measures and evaporites. A brief discussion of this kind is offered in this chapter, not to present realistic paleoclimatic interpretations, but to suggest the potential of such studies elsewhere in the world.

Paleopedological studies of Tertiary non-marine rocks now are becoming commonplace (Singer and Nkedi-Kizza, 1980; Singler and Picard, 1981; Andreis, 1981; Allen, 1986; Retallack, 1983a,b, 1985, 1986b; Fastovsky and McSweeney, 1987; Bown and Kraus, 1981, 1987), but their recent publication dates and lack of worldwide coverage reflect past problems with expertise and interest in paleosols. Few geologists have been trained in soil science, and few soil scientists have examined sedimentary rock sequences. Many paleosols have been altered substantially during burial, and many of the sensitive analytical approaches of soil science are not appropriate for them. Much remains to be done, but these problems now are being addressed on a wide front. Paleosols can be recognized within sedimentary rocks by three broad classes of diagnostic features: root traces, soil horizons and soil structure (Retallack, 1988a). They are best characterized by a combination of petrographic, mineralogical and geochemical techniques (Retallack, 1983a, 1986b). The alteration of paleosols after burial is in many ways like that of other rocks (Nesbitt and Young, 1989; Retallack, 1990, 1991). There are a variety of approaches for measuring sections, mapping and naming of individual paleosols (Retallack, 1983a), deep weathering zones of paleosols (Senior and Mabbutt, 1979) and natural assemblages of paleosols (Bown and Kraus, 1987). Interpretation of paleosols may be made in two distinct ways: by identification of a paleosol in a classification of modern soils and by study of particular paleoenvironmentally significant features of paleosols (Retallack, 1983a, 1990; Fastovsky and McSweeney, 1987). The second interpretive approach is emphasized in this chapter.

CLIMATICALLY SENSITIVE FEATURES OF PALEOSOLS

A variety of features of surface soils show a clear and significant relationship, or climofunction, to climatic variables. In general, tropical soils are more red in hue, more clayey, less rich in organic matter, and more deeply and thoroughly weathered than soils of temperate climates (Birkeland, 1984). Unfortunately however, none of these generalizations are very useful for interpreting paleoclimate from paleosols for two reasons. Some of these features (reddening, increased clay content, greater depth and intensity of weathering) are also related to the time available for soil formation and mean annual rainfall, as well as former temperature. Some of these features (reddening, depletion of organic matter) also are known to be imposed during burial alteration of paleosols

(Retallack, 1983a, 1990). From the large number of climate–soil relationships now known, a few are sufficiently independent of other soil–forming factors and unaffected by burial to be useful in interpreting paleoclimate from paleosols (Retallack, 1990). Only a few of these can be outlined here.

One useful climofunction for estimating mean annual rainfall from paleosols is the depth to the top of the calcareous nodular (calcic or Bk) horizon, which is deeper in wetter climates. This has been demonstrated for surface soils in several different parts of the world: North American Great Plains (Jenny, 1941; Ruhe, 1984), Mojave Desert of the North American southwest (Arkley, 1963), Serengeti Plains of Tanzania (de Wit, 1978) and Indo–Gangetic Plains of India and Pakistan (Sehgal et al., 1968). This general relationship is remarkably consistent among soils of different climate, parent material and time for formation (Ruhe, 1984). The depth to the calcic horizon should not to be confused with the depth of leaching of carbonate in non–calcareous soils of humid climates (Birkeland, 1984), nor with the depth to calcareous stringers and horizons of very old soils (Kubiena, 1970), nor with the depth to pore–filling simple cements of the kind produced in groundwater calcretes (Retallack, 1991). Carbonate nodules within paleosols are robust in the face of alteration after burial even up to greenschist-grade metamorphism (Retallack, 1985). The main problems with using depth to the nodular horizon as an indicator of paleoclimate are compaction of the upper part of paleosols during burial and erosion of the upper part of paleosols prior to burial. Corrections for compaction can be made using geological information on depth of burial and standard compaction curves (Baldwin and Butler, 1985). An assessment of surface erosion can be based on preservation of root traces and soil structures (especially granular peds) typical of surface horizons of paleosols. A greater limitation is the variance of the relationship in surface soils. In soils of the midwestern United States the relationship between depth to the calcic horizon (D in cm) and mean annual precipitation (P in cm) is $D = 2.5(P-30.5)$. Slightly different regressions of this kind of data have been found elsewhere. Until these studies are extended and integrated into a coherent whole for a variety of different conditions, the interpreted ranges of former precipitation given here and previously (Retallack, 1983a, 1986a) correspond to the total range of values observed in postglacial soils with calcic horizons at comparable depths.

Another potential indicator of mean annual rainfall is the nature of clay minerals in paleosols. In general, sepiolite and palygorskite are found in desert climates, smectite in dry climates and kaolinite in wet climates (Weaver, 1989). The kinds of clay mineral produced also depend on the nature of the parent material and the duration of soil formation among other factors, but most of these variables can be controlled by examining the changing proportions of clay minerals within paleosol profiles or between paleosol profiles of comparable degree of development. Far more serious for the paleoclimatic interpretation of paleosol clays is the diagenetic alteration of smectite to illite during burial, especially beyond about 2 km and 200°C (Eberl et al., 1990).

In extremely dry climates, there is so little rain that soils and lakes become salty from the evaporation of groundwater. Gypsum is a common soil evaporite, and may also form horizons whose depth from the surface of the soil reflects mean annual rainfall (Dan and Yaalon, 1982). Unfortunately, such evaporite minerals are readily leached from rocks during burial, and so not always preserved in paleosols, although pseudomorphs of them are sometimes found (Warren, 1989).

Strongly seasonal wet–dry climates induce characteristic soil features in clayey soils: the hummock and swale surface (gilgai microrelief) and cracking and deformation of surface horizons (mukkara structure of Paton, 1974) found in Vertisols. These kinds of structures are very robust during burial, and are known in very ancient paleosols, in some cases metamorphosed to within the greenschist facies (Allen, 1986; Retallack, 1986b).

PALEOSOLS IN BADLANDS NATIONAL PARK, SOUTH DAKOTA

One place where paleosols spanning the Eocene/Oligocene boundary have been studied in detail is in the Pinnacles area of Badlands National Park, in southwestern South Dakota, U.S.A. (Figure 19.1). These colorful, clayey,

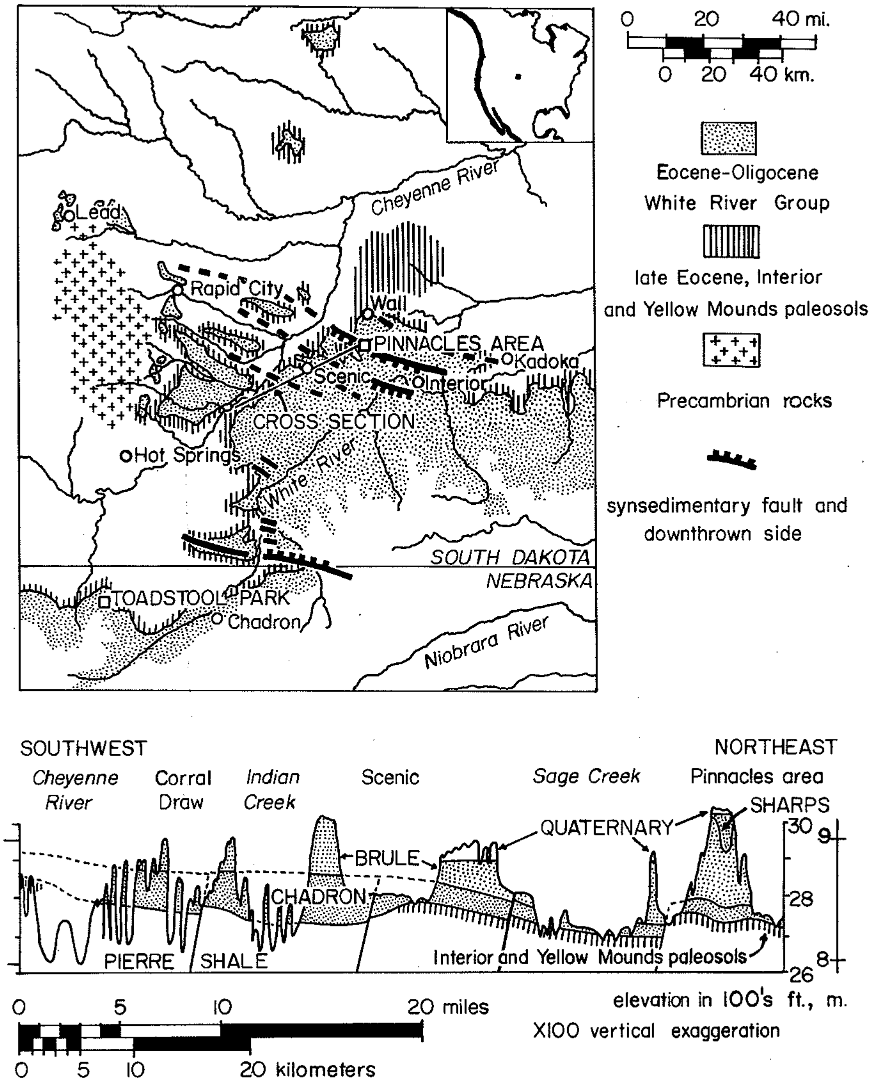


FIGURE 19.1. Map and cross section of Eocene–Oligocene alluvium of the White River Group, of extensive late Eocene Interior and Yellow Mounds paleosols developed in Cretaceous marine sediments, and of Precambrian rocks of the Black Hills uplift, in northeastern Nebraska and southwestern South Dakota (adapted from Retallack, 1988b).

non–marine rocks are exposed in a long ragged wall between high prairie to the north and the valley of the White River to the south. They have long been famous for their great variety of well preserved fossil mammals (Emry et al., 1987) and as the foremost example of badlands weathering (Schumm, 1962). It is now known that their subtle color banding reflects a large number of successive buried soil horizons (Figure 19.2). These paleosols have been described in detail, and used as guides to local Oligocene paleoenvironments (Retallack, 1983a,b), the evolution of ancient grassland ecosystems (Retallack, 1982, 1984b, 1988, 1990), completeness of stratigraphic sections (Retallack, 1984a) and factors in deposition of this volcanic–alluvial sedimentary sequence (Retallack, 1986a). This chapter looks beyond these issues to the record in paleosols of global climate change.

A particularly impressive break in the nature of these buried soils in the Badlands is between

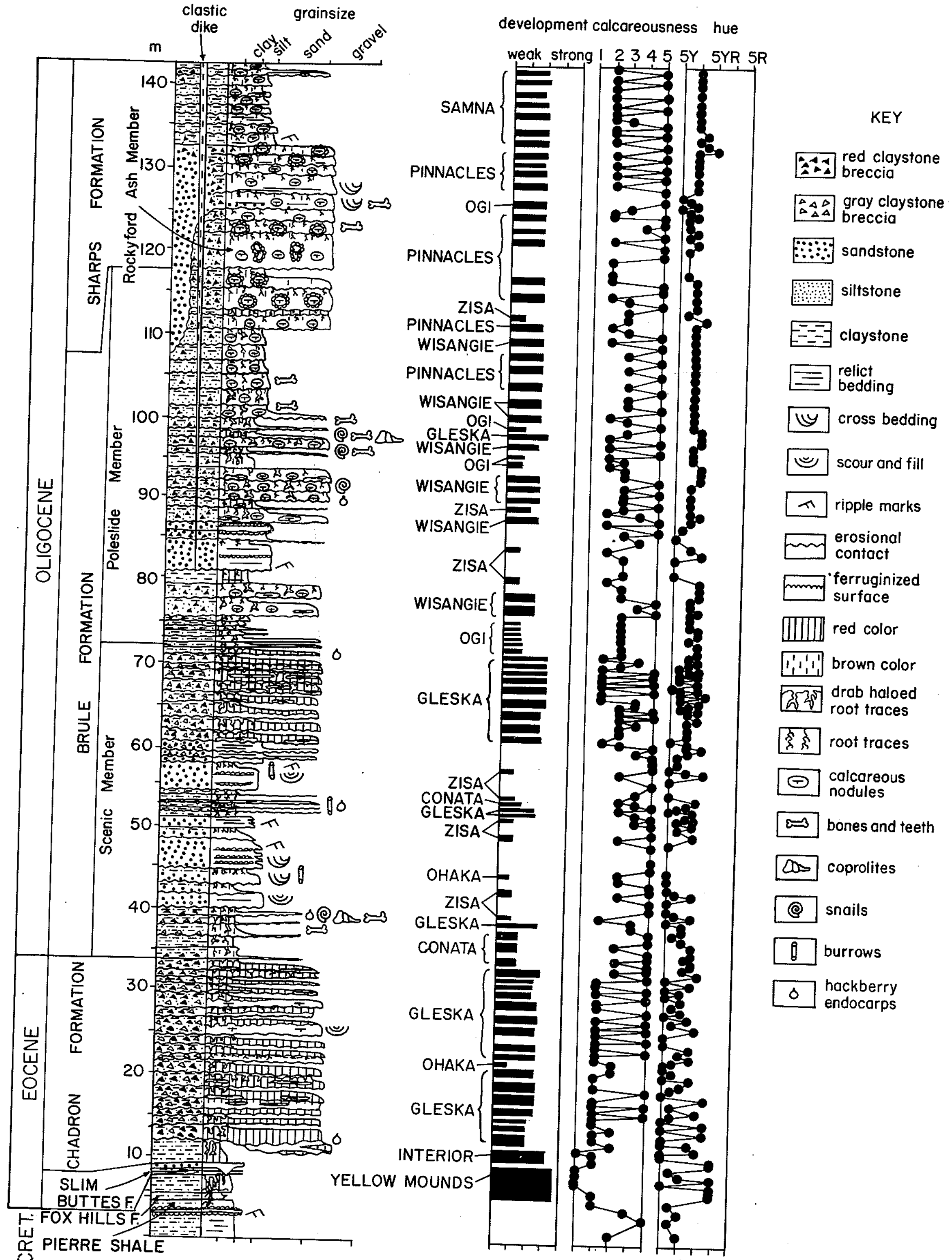


FIGURE 19.2. A measured section of paleosols in the Pinnacles area of Badlands National Park, South Dakota (from Retallack, 1990). Black boxes indicate position of paleosols and their width is their degree of development (in scale of Retallack, 1988a). Calcareousness by acid testing in field (using scale of Retallack, 1988a). Hue from olive to red after Munsell color charts.

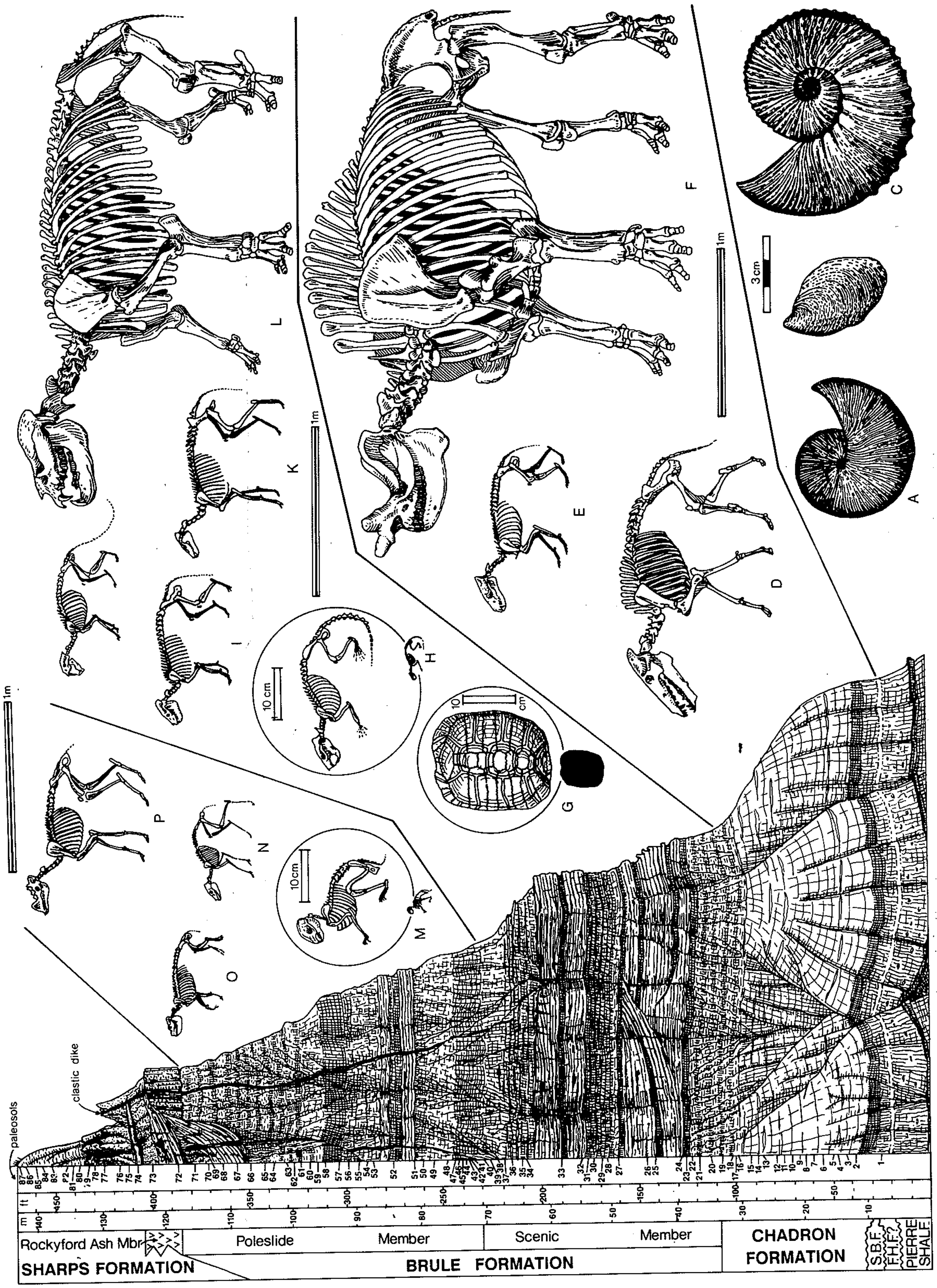
the pink and green banded, clayey Chadron Formation and the brown to white, nodular, silty Brule Formation (Figure 19.3). This lithological contact was until recently considered mid–Oligocene in age (Prothero et al., 1983), but reassessment of radiometric dating and magnetostratigraphy of Oligocene continental deposits in the western United States now indicate that it is the local Eocene/Oligocene boundary and some 34 million years old (Swisher and Prothero, 1990). This and other boundaries between rock units in the Badlands are now known to reflect paleoclimatic changes, as revealed by studies of the depth of carbonate nodules, clay content, clay minerals and other features of paleosols in the sequence.

The depths to calcareous nodular horizons and layers within paleosols were recorded in the field during section measuring (Figure 19.4), and show a decline in paleosols higher within the section. Linear regressions of these data were only done within rock units, because of evidence from paleosols (Retallack, 1986a) and magnetostratigraphy (Prothero et al., 1983) for profound disconformities between rock units. The tabular calcareous (petrocalcic) horizons low in the sequence are of a kind that are not necessarily related to climate in surface soils (Kubiena, 1970), but they do continue the relationship of shallowing with time. The relationship also extrapolates back to the oldest paleosol on the unconformity with Cretaceous marine rocks. This paleosol has been leached of fossiliferous marine, calcareous nodules to a depth of 511 cm. For this paleosol and the non–calcareous profile above it, mean annual rainfall was probably in excess of 1000 mm, by comparison with surface soils in the Great Plains (Jenny, 1941; Ruhe, 1984) and elsewhere (Arkley, 1963; de Wit, 1978; Sehgal et al., 1968). The depth of petrocalcic horizons in the late Eocene Chadron Formation should not be trusted as a paleoclimatic indicator, but these paleosols probably received less than 1000 mm and more than 500 mm. For the Scenic Member of the early Oligocene Brule Formation, there also are petrocalcic horizons, indicating mean annual rainfall from about 900 mm to 500 mm. Rainfall then declined to 450 to 500 mm for the Poleslide Member of the Brule Formation and 350 to 450 mm for the Sharps Formation. These estimated ranges are the total range of precipitation found in surface soils of the Great Plains with nodules at comparable depths to the paleosols.

These results do not appear compromised by erosion of the paleosols prior to deposition. Surface horizons with fine root traces and granular ped structure are well preserved in many of the paleosols (Retallack, 1983a, 1990). Nor were these results compromised by compaction. The maximum thickness of the Eocene–Oligocene White River Group is 195 m, and the cumulative maximum thicknesses of other Oligocene and younger formations in this region total 361 m (Martin, 1983). If a marine, non–calcareous shale were buried to a depth of 500 m it would be compacted to about 90% of its former thickness (Baldwin and Butler, 1985). This seems a very unlikely maximal compaction at the base of the Chadron Formation, because these younger sediments did not form a single layer, but filled local paleovalleys created as the nearby Black Hills continued to rise, as has been especially well documented in nearby northwestern Nebraska (Schultz and Stout, 1980; Swinehart et al., 1985).

Another indicator of climatic change is the nature of clay minerals in the paleosols, as determined by x–ray diffraction studies (L.G.

FIGURE 19.3. (next page) The Eocene–Oligocene boundary is currently recognized in the Pinnacles area of Badlands National Park, South Dakota, between the clayey, pink, mound–like paleosols of the Chadron Formation with their titanothere–dominated fauna, and the silty, brown, cliff–forming paleosols and paleochannels of the Brule Formation with their oreodon–dominated fauna: A–C, are Late Cretaceous molluscs, *Hoploscaphites nicolleti* (A), *Tenuipteria fibrosa* (B), and *Discoscaphites cheyennensis* (C); D–F are late Eocene mammals, *Archaeotherium mortoni* (D), *Hyaenodon horridus* (E), and *Brontops robustus* (F); G–L are early Oligocene tortoise (G) and mammals, *Stylemys nebrascensis* (G), *Ischyromys typus* (H), *Merycoidodon culbertsoni* (I), *Hoplophoneus primaevus* (J), *Mesohippus bairdi* (K) and *Metamynodon planifrons* (L); M–P are mid–Oligocene mammals, *Palaeolagus haydeni* (M), *Leptomeryx evansi* (N), *Leptauchenia decora* (O) and *Protoceras celer* (P; from Retallack, 1983b).


paleosols
clastic dike
Rockyford Ash Mbr
SHARPS FORMATION
Poleslide Member
Scenic Member
BRULE FORMATION
CHADRON FORMATION
S.B.F.
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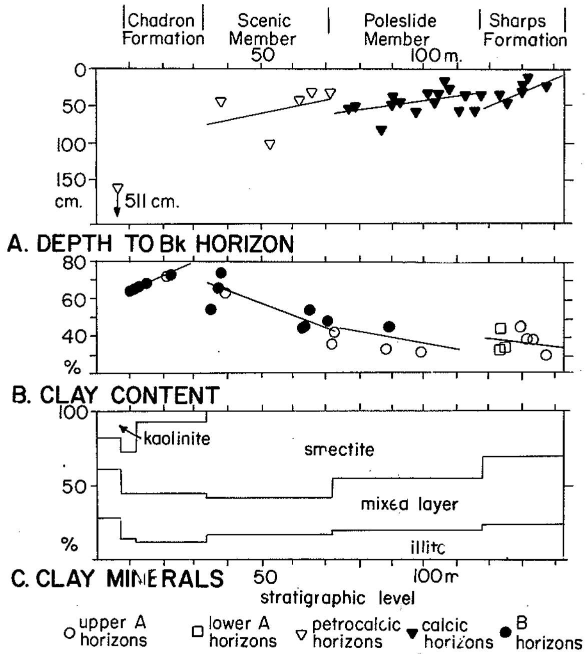

FIGURE 19.4. Indices of stepwise climatic drying in a measured section of paleosols in the Pinnacles area, Badlands National Park, South Dakota (from Retallack, 1990; data from Retallack, 1983a, 1986a): A, depth to calcareous horizons, both nodular (calcic) and continuous (petrocalcic), as measured in the field: B, clay content of paleosol horizons, determined by point counting petrographic thin sections: C, proportion of clay minerals in each recognized rock unit, by averaging many x-ray diffractometer traces for each unit. Correlation coefficients (r) for regression lines in A, are right to left, 0.72, 0.40, 0.41.

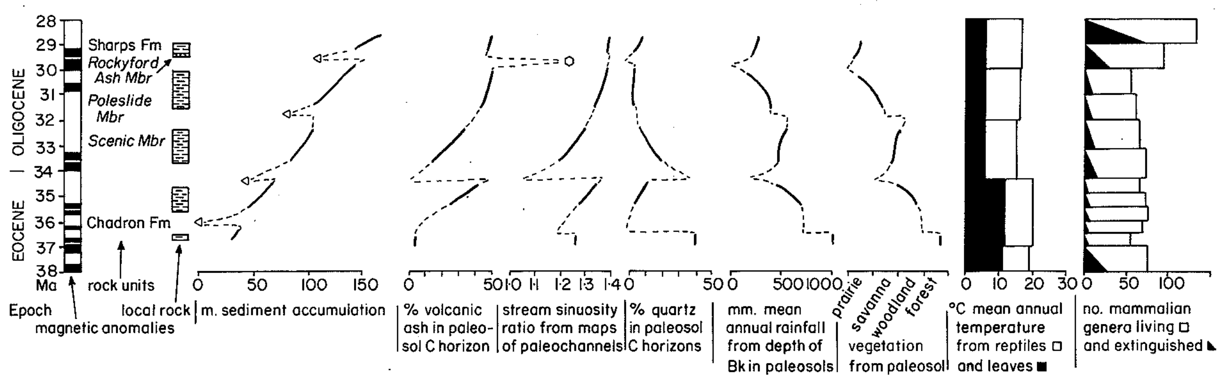

FIGURE 19.5. Estimated controls on sedimentation and erosional episodes in a sequence of paleosols in the Pinnacles area, Badlands National Park, South Dakota (adapted from primary data of Wolfe, 1978; Hutchison, 1982; Retallack, 1986a, Stucky, 1990): changing mammalian generic diversity and extinctions (from taxonomic compilations), mean annual temperature (from reptile diversity and percentage entire-margined leaves), vegetation (from paleosol morphology), rainfall (from depth to paleosol calcareous horizons), stream sinuosity (measured from maps of paleochannels), far travelled alluvium (from point counting for quartz in paleosol C horizons), volcanic ash (from point counting for volcanic shards in paleosol C horizons), and sediment accumulation (from degree of development of paleosols), plotted for current understanding of the Oligocene time scale (Swisher and Prothero, 1990).

Schultz, 1961; Retallack, 1983a, 1986a). The two lowest paleosols in the sequence contain appreciable amounts of kaolinite, as would be expected in soils of humid climate. Paleosols higher in the sequence contain largely smectite, with increasing amounts of illite up-section (Figure 19.4). These less intensely weathered clays were formed in a relatively drier climate during Oligocene time.

These results are not compromised greatly by effects of parent material. The lowest two paleosols developed on smectitic marine shale. The rest of the sequence developed on smectitic alluvium and redeposited soil derived ultimately from rhyolitic airfall ash. This material is of very uniform composition chemically and mineralogically, and relic shards are found throughout the sequence above the basal paleosol (Retallack, 1983a). Nor are these results compromised by illitization of the clay minerals in this sequence. The shallow depth of burial of these deposits is well short of that required for illitization, and the up-profile increase in illite abundance is the reverse of that seen in studies of illitization in boreholes (Weaver, 1989). These results are compromised somewhat by differences in development of the paleosols. The basal two profiles are much better developed than those of the Chadron to Sharps interval, but many of these younger paleosols show comparable degree of development (within the categories of Retallack, 1988a). Stepwise changes in clay composition above the basal paleosols remain as a climatic signal.

Evidence of soil cracking was seen in many of the paleosols in the form of soil clods (peds in the terminology of soil science) surrounded by clay skins (argillans), but deep cracking was uncommon. The zeolite clinoptilolite was common in several paleosols high within the exposed sequence, but this rare associate of evaporitic soils of very dry climates could have formed during burial of the paleosols (Retallack, 1983a). Evidence for very dry and highly seasonal climates was localized to boundaries between the rock units in the Badlands. For example, the Chadron Formation immediately below the Brule Formation is riddled with chalcedony-replaced pseudomorphs of gypsum and chalcedony-filled deep cracks (Honess, 1923; Lawler, 1923; Retallack, 1983a). Both these features have been documented in saline playa lakes and surrounding soils (Eugster, 1969). The prominent clastic dikes that penetrate as deeply as 60 m from the top of the exposed sequence of Sharps Formation (Figure 19.3), are most like the giant desiccation cracks seen in playa lakes and soils of the desert southwestern United States (Retallack, 1983a). A less extreme climatic drying may have created calcareous rhizoconcretions with partial chalcedony filling, rare for the sequence as a whole, within the upper part of the second paleosol above the base of the sequence in the Pinnacles area. Such features are found in desert soils (Retallack, 1983a). Magnetostratigraphic (Prothero et al., 1983) and completeness studies of these paleosol sequences (Retallack, 1984a, 1986a), have shown that the boundaries between recognized rock units are disconformities reflecting periods of erosion at least as long as the periods of net deposition that created each rock unit. Thus some of these periods of dry climate were also times erosion.

Other features of the paleosols such as clay content and hue may also reflect climatic change. Clay content was determined by point counting petrographic thin sections (Figure 19.4), and hue by comparison of freshly excavated rock chips with a Munsell color chart (Figure 19.2). More clayey and red paleosols were formed in humid to subhumid paleoclimates of the lower part of the Badlands sequence, but more silty and brown paleosols in semiarid paleoclimates of the upper part of the sequence.

Paleoclimatic interpretation of clay content and hue of these paleosols must remain imprecise, for a variety of reasons. Estimation of clay content from petrographic thin sections systematically overestimates clay abundance of paleosols and soils (Murphy and Kemp, 1984) and the color of rock chips is not directly comparable with the hue of either wet or dry soils. Both clay content and color reflect paleotemperature, duration of soil formation, and former vegetation of the paleosols, as well as climatic drying (Birkeland, 1984). Indeed there is evidence from fossil floras and reptiles in the western United States for cooling at the Eocene/Oligocene boundary (Wolfe, 1978; Hutchison, 1982). The color of these paleosols has certainly been altered during burial by the subsurface decomposition of most soil organic matter (Stevenson, 1969) and by the dehydration and recrystallization to coarser grainsize of ferric hydroxide minerals (Walker, 1967). Nevertheless, relative changes in clay content within this sequence, which is uniformly altered in this way, may be interpreted as a muted signal of relative paleoclimatic change, in support of other more robust paleoclimatic indicators.

Also broadly compatible with indications of paleoclimatic change are likely changes in vegetation and animals with time in this sequence. The two, thick, red, kaolinitic basal paleosols contain a large, deeply–penetrating root traces and well–differentiated clayey subsurface (Bt) horizons, as in surface soils under forests. Silicified wood has been found at this stratigraphic level (Clark et al., 1967), and it contains growth rings as evidence of some climatic seasonality.

Moderately developed, green–pink, calcareous, clayey paleosols of the late Eocene Chadron Formation also contain abundant, large, drab–haloed root traces and blocky, angular peds. The drab haloes were probably produced by bacterial reduction of organic matter associated with the roots during shallow burial (Retallack, 1990), but their size and distribution, as well as the overall profile form of these paleosols, are of kinds now found under dry woodland vegetation. A few paleosols in the Chadron Formation are thin and gray, with only fine root traces and some relict bedding, as in soils of herbaceous vegetation early in the ecological succession to colonize disturbed ground (Retallack, 1983a). Fossil logs, walnuts (*Juglans siouxensis*) and hackberry (*Celtis hatcheri*) endocarps from the Chadron Formation provide additional evidence of dicot trees (Chaney, 1925; Manchester, 1987), but the rarity and exceptional preservation of these remains limit their usefulness in paleoecological reconstruction (Retallack, 1984a).

Grey–pink paleosols with large root traces also persist in the Brule Formation, though closely associated with paleochannels. Thin paleosols with fine root traces and relict bedding also are found in the Brule Formation, but here they are red as well as gray in color. The distinctive brown, nodular, silty appearance of much of the Brule Formation is due to the appearance of new kinds of paleosols with abundant fine root traces, as well as scattered large ones. These paleosols have granular ped structure, and a subsurface (Bk) horizon of calcareous nodules. These paleosols are most like those of wooded grassland or wooded shrubland in the modern southern Great Plains of North America (Aandahl, 1982). Vegetation during deposition of the Brule Formation has been reconstructed as a mosaic of streamside early successional vegetation and gallery woodland, with more open wooded grassland away from streams. Hackberry endocarps (*Celtis hatcheri*) also are found in the Brule Formation as evidence of trees (Retallack, 1984a,b). Fossil grasses (*Stipidium* sp.) were not found here, but have been recorded as rare fossils from the Brule Formation in nearby Colorado (Galbreath, 1974).

A single paleosol with large, drab–haloed root traces has been found interbedded with a deeply incised sandstone paleochannel in the Sharps Formation at the top of the exposed sequence, but most of the paleosols at this stratigraphic level lack large root traces. They are thin profiles, with abundant fine root traces and granular peds, over a very shallow and moderately developed (Bk) horizon of calcareous nodules, as in dry parts of the modern Great Plains and intermontane west (Aandahl, 1982). By this time open rangelands were probably widespread. No plant fossils of any kind have yet been found from the Sharps Formation.

Stepwise changes are seen in the Badlands also in fossil vertebrate faunas. Studies involving careful sampling of many levels have shown that fossil mammalian faunas and individual mammalian species are relatively uniform from the bottom to the top of the recognized rock units, but there are profound changes in the fossil mammals between units (Clark et al., 1967; Prothero, 1985). Some of these faunal overturns may be related to climatic change. The boundary between the two basal paleosols in the Pinnacles area may correlate with the boundary between the Duchesnean and Chadronian North American land mammal "ages" (about 37 Ma old). This was a time, less significant than the Uintan/Duchesnean boundary (about 40 Ma), when many archaic forest mammals of the Eocene again became extinct and modern groups of mammals became established (Hutchison, 1982; Emry, et al., 1987; Stucky, 1990). The relevance of paleosols for this faunal shift would be better established by their study in localities that contain more abundant fossil mammals of this age than found in the Badlands. The Pinnacles section is however, relevant for interpreting the faunal change between the Chadron and Brule Formation, which corresponds to the boundary of the Chadronian and Orellan North American land mammal ages and the Eocene/Oligocene boundary (about 34 Ma). At this time alligators, many species of turtles, and the large titanotheres became extinct: only one specimen of a titanothere and a single small alligator have been found in the basal Brule Formation (Clark et al., 1967), which has an unusually low diversity of fossil turtles (Hutchison, 1982). Fossil land snails also were much reduced in diversity at this time (Evanoff et al., this volume). All these creatures would have been more dependent than other vertebrates on consistent supplies of water, which became scarce here at the Eocene/Oligocene boundary. In addition, reduced reptilian diversity could be explained by a drop from a mean annual temperature of 20° C to 15° C (Hutchison, 1982). A temperature drop of comparable magnitude at this time can be seen in montane floras of the Rocky Mountains: from 11° C to 6° C (Wolfe, 1978; Hutchison, 1982). Higher within the Brule Formation and within the Sharps Formation a similar assemblage of animals persists, but there are increases in the abundance and diversity of burrowing mammals and decreases in the overall body size of the mammals within the most prominent evolutionary lineages. Both trends may reflect a drier, cooler and less productive ecosystem through time.

Stepwise, episodic, long term, paleoclimatic change is clearly evident in this sequence of paleosols. By controlling the density of vegetation cover and landscape stability, climatic deterioration can also control rates of sediment accumulation, episodic periods of erosional downcutting and fluvial style. A variety of other factors, including tectonic uplift of the source terrane and volcanic input of ash also can determine the timing and nature of alluvial deposition (Retallack, 1986a). Accumulation of these sequences was due in part to an inbalance between tectonic subsidence and sediment influx from volcanic ash and stream flooding (Swinehart et al., 1985). However, exceptional uplift of the source region should introduce far–travelled components to the alluvium such as quartz and unusually voluminous volcanism should be evident from thick deposits of airfall ash unaltered by Oligocene soil formation. There is evidence that each of these occurred during at least one period of erosional downcutting, but the constant association of each erosional episode with evidence for climatic deterioration (Figure 19.5), implicates paleoclimate as an overiding extrinsic control on erosional episodes now dated at 40, 37, 34, 32 and 29.5 million years ago.

PALEOSOLS IN OTHER REGIONS

Paleosols are known now in several other Eocene–Oligocene sequences, and suspected in many more. This brief discussion of other areas for the purposes of illustrating the global scope of changes in paleosols at that time, is limited to areas in which at least a few paleosols have reported in more than cursory fashion.

Paleosols in the Chadron and Brule Formations change in appearance west from the Pinnacles area of Badlands National Park to Sheep Mountain Table in southwestern South Dakota, to Toadstool Park National Monument in northwestern Nebraska, and to the Lusk and Douglas areas of southeastern Wyoming and northeastern Colorado, northwest to Flagstaff Rim Wyoming and north to Slim Buttes, South Dakota (Schultz et al., 1955; Schultz and Stout, 1980; Singler and Picard, 1981; Retallack, personal observations). Proceeding westward, paleosols at the same stratigraphic level become more calcareous, silty and thinner, so that the Chadron Formation near Douglas and Slim Buttes is more like the Brule Formation in the Badlands National Park. A fossil flora found in the Chadron Formation at Flagstaff Rim is dominated by small–leaved plants that indicate dry climatic conditions (Wing, 1987). At any given time it was drier to the west from South Dakota into Wyoming and Colorado, and climatic drying through Eocene and Oligocene time lengthened the rain shadow cast by the Rocky Mountains and their active silicic volcanoes.

Abundant paleosols and fossil mammals of Eocene and Oligocene age also are known in the John Day Country of central Oregon (Retallack, 1981; Smith, 1988; Pratt, 1988). This thick sequence of volcanic and volcaniclastic rocks also has preserved a superb fossil record of fossil plants (Manchester, 1981; Manchester and Meyer, 1987). The thick, red, deeply weathered paleosols of the Eocene Clarno Formation are compatible with evidence from fossil plants for rainforest vegetation and climate (Smith, 1988). Red paleosols also are found in the overlying Oligocene John Day Formation, but are discernably less deeply weathered, and associated with a variety of brown and green, calcareous paleosols generally like those thought to have supported wooded grassland in the Pinnacles area of Badlands National Park. Fossil plants in the John Day Formation are of cool temperate climatic affinities (Wolfe, 1978), but do not show indications of dry climate, perhaps in part because they lived around lakes in whose deposits they are found. Nevertheless, Oregon probably became drier across the Eocene/Oligocene boundary, though remaining wetter than South Dakota at this time.

In southwestern Oregon (Bestland, 1987), northwestern California (Singer and Nkedi-Kizza, 1981) and central Washington (Gresens, 1981), there are thick deeply weathered paleosols of likely late Eocene age, and red paleosols with large root traces of Oligocene age. Much remains to be done on these paleosols and their climatically sensitive features.

In Argentina, grassland paleosols have been identified within volcaniclastic alluvium of Eocene age, or Mustersan in the local mammalian scheme (Spalletti and Mazzoni, 1978; Andreis, 1981). These are the oldest likely grassland paleosols known anywhere, and presumably reflect expansion over Patagonia of a rain shadow from the volcanic Andean Cordillera. Calcareous paleosols have been found in upper Oligocene, or local Deseadean, alluvial rocks in this region, but their paleoclimatic significance has not yet been established, and this sequence is not complete enough to preserve the Eocene/Oligocene boundary (MacFadden et al., 1985).

In the Paris Basin of France, there is a well-preserved late Eocene (probably about 40 Ma), podzolic paleosol (Pomerol, 1964). Similar Orthods now form under oligotrophic forest in humid climates (Retallack, 1990). This paleosol is followed by latest Eocene marine deposits and lacustrine evaporites, which latter indicate significant climatic drying. Oligocene coastal dune deposits of the Paris Basin include thin, humic paleosols (Morand et al., 1968). Such soils of swampy areas are relatively insensitive to climate because of their isolation by groundwater. Late Oligocene to early Miocene paleosols of Aquitaine have both clayey and calcareous (Bt and K) horizons of a kind found in strongly developed soils of dry woodland (Meyer et al., 1980).

In the Fayum Depression of Egypt there is a spectacularly well exposed sequence of paleosols that may span the Eocene/Oligocene boundary (Bown and Kraus, 1988; Rasmussen et al., this volume). This boundary may be better placed within the Jebel Qatrani Formation, rather than at its base as in the past [considering recent recalibration of the time scale by Swisher and Prothero (1990), and radiometric dates of Fleagle et al. (1986); see Rasmussen et al., this volume]. There are differences between the paleosols in the upper and lower part of this formation and those of underlying formations, but none yet documented that were obviously related to paleoclimatic change. There are marine intercalations in this sequence and many of the paleosols may have supported mangal vegetation buffered by marine influence from regional climate.

Near Lake Zaisan in Soviet Kazakhstan, inner Asia, there is a sequence of paleosols and lake beds rich in fossil plants (Akmetiev et al., 1986). Drab–sideritic and pink–mottled Eocene paleosols are overlain by a sequence with calcareous nodular Oligocene paleosols. In nearby Mongolia, there are red and green, nodular upper Eocene paleosols and reddish–brown Oligocene paleosols (Dashvezeg and Devyatkin, 1986).

In southwestern Queensland, Australia, coal measures and sandstones of Paleocene to Eocene age have been deeply weathered by a thick lateritic paleosol of Oligocene age (Senior and Mabbutt, 1979). This change may reflect improved drainage across the Eocene/Oligocene boundary, but this area formerly at high latitude probably remained covered in wet forests.

OTHER LINES OF EVIDENCE FOR PALEOCLIMATIC CHANGE

Eocene and Oligocene paleosols in most parts of the world are little known compared to those of South Dakota, but they do demonstrate a potential for further studies of global climatic change. Although a global data base from paleosols is not yet available, it is instructive to compare the timing of paleoclimatic changes indicated by paleosols in Badlands National Park with those proposed on the basis of other lines of evidence in different parts of the world.

There is evidence of climatic drying from the amounts of eolian quartz in deep sea cores of the Pacific Ocean (Rea et al., 1985). The late Eocene (about 37 Ma) and Eocene–Oligocene transition (about 34 Ma) were times of significantly increased mass accumulation and weight percent of eolian quartz, although there were much more significant influxes of eolian quartz during early Miocene time. The Eocene–Oligocene transition was also a time when the mean grain size of eolian quartz in these cores began to climb to near modern values, which were attained during early Miocene time.

Also in Pacific Ocean cores, oxygen–isotopic studies of foraminifera have shown declines in ocean temperature during the late Eocene (about 37 Ma), followed by a very profound drop at the Eocene/Oligocene boundary (now dated about 34 Ma) and smaller drops during Oligocene time (32 Ma: Miller, this volume). Climatic cooling at similar times has been postulated to explain the declining percentage of entire–margined leaves in fossil floras (Wolfe, 1978) and diversity of turtles and crocodilians (Hutchison, 1982) in the western United States, where the Eocene–Oligocene cooling was much more profound than the other episodes. In northwest Europe, fossil palynofloras show a decline in land temperature at the Eocene/Oligocene boundary (Boulter, 1984). In Europe this was also a time of major extinctions of forest–adapted archaic species (Hartenberger, 1988), and their replacement with a more modern fauna with a size distribution suggestive of dry, open rangeland (Legendre, 1987).

At these critical times, there were also major marine regressions (Haq et al., 1987). The circum–Antarctic current was becoming well established as Australia drifted away from Antarctica and the central Tethys Ocean was much constricted as Africa began to make full contact with Europe. The ability of oceans to absorb and circulate heat was thus dramatically curtailed, as was hydrothermal activity and the generation of carbon dioxide along mid–ocean ridges (Owen and Rea, 1985). Antarctic ice caps presumably began to grow at these times, although they did not reach sea level until late Oligocene time (Kennett, 1982). These were times when the world stepped out of the torrid and more equable paleoclimate of middle Eocene times, closer to the highly zoned, glacial climates of the past two million years.

CONCLUSIONS

Evidence of stepwise climatic deterioration during late Eocene time (40 and 37 Ma), the Eocene/Oligocene boundary (34 Ma) and Oligocene time (32 and 29.5 Ma) is provided by a variety of paleoclimatically–sensitive features of paleosols in a detailed measured section in Badlands National Park, South Dakota. The most informative of these features were the changing proportions of clay minerals and the depth to the calcareous nodular horizons in the paleosols. This latter line of evidence by comparison with surface soils in the North American Great Plains indicates mean annual precipitation of more than 1000 mm for the early part of the late Eocene (local Duchesnean), somewhat drier for the latest Eocene (Chadronian), 500 to 900 mm for the early Oligocene (Orellan), 450 to 500 mm for the mid–Oligocene (Whitneyan) and 350 to 450 mm for the late Oligocene (Arikareean). Deep clastic dikes and pseudomorphs of gypsum at the Eocene/Oligocene and the mid/late Oligocene boundaries are evidence that climate was at times even drier than these estimates.

These rainfall estimates are compatible with evidence from fossil root traces and profile form of the paleosols for late Eocene forest, latest Eocene dry woodland, early Oligocene wooded grassland with gallery woodland and mid–Oligocene open grassland with very few trees along watercourses. Paleoclimatic deterioration at 40 Ma is not well understood in Badlands National Park. Although the other steps at 37, 34, 32 and 29 Ma appear to have been of equal magnitude, the paleoclimatic change with the greatest biological effects was the one at 34 Ma, which heralded the earliest known wooded grasslands in North America.

Climatic drying also could explain the stepwise decrease in clay content and change from red to yellow hue of the paleosols from the base to the top of the section, although climatic cooling and other paleoenvironmental factors may also have played a role in these features of the paleosols.

A limitation on the paleosol sequence in Badlands National Park as a record of paleoclimate is its profound disconformities, separating each of the major rock units. These long periods of non–deposition and erosion may also have been caused by climatic deterioration. Tectonically determined subsidence rates were critical in allowing sediment to accumulate at all, but periodic variations in accumulation rate do not all appear to be due to exceptional tectonic or volcanic activity, but rather to thinner vegetation of dry periods less able to stabilize the landscape against erosion.

Much more could be done to improve the quality of paleoclimatic data from paleosols of Badlands National Park, and to extract similar information from other sequences of paleosols of this age, especially in central Oregon, southern Argentina, Kazakhstan, and Mongolia. These paleoclimatic records are useful additions to those based on fossil reptiles in paleosols, on fossil leaves in lake deposits, on isotopic composition of marine foraminifera in deep sea cores and on amounts of eolian dust in the deep sea. Not only was the stepwise, mid–Tertiary paleoclimatic deterioration a matter of cooling, but also of drying in the interiors of large mid–latitude continents. Both these climatic coolings and desiccations may have been due to a redistribution of water into polar ice caps. Away from the poles, paleoclimatic changes ushered in early grasslands and pronounced modernization of mammalian faunas. Neither the polar ice caps, grasslands nor grazers were as extensive or highly evolved as they are today, but a series of climatic deteriorations around the Eocene/Oligocene boundary were important turning points in their geological history.

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