

Cenozoic Paleoclimate on Land in North America

Gregory J. Retallack

*Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403, U.S.A.
(e-mail: gregr@uoregon.edu)*

ABSTRACT

Paleotemperature and paleoprecipitation over the past 40 m.yr. can be inferred from the degree of chemical weathering and depth of carbonate nodules in paleosols of Oregon, Montana, and Nebraska. Paleosol records show that late Eocene (35 Ma), middle Miocene (16 Ma), late Miocene (7 Ma), and early Pliocene (4 Ma) warm climatic episodes were also times of a wet climate in Oregon, Montana, and Nebraska. Oregon and Nebraska were humid during warm-wet times, but Montana was no wetter than subhumid within the rain shadow of intermontane basins. Global warm-wet paleoclimatic spikes steepened rather than flattened geographic gradients of Rocky Mountain rain shadows. Long-lived mountain barriers created dusty dry basins with sedimentation rates high enough to preserve Milankovitch-scale (100–41 kyr) global paleoclimatic variation in some sequences of paleosols. Greenhouse warm-wet climates indicated by paleosols were also peaks of diversity for North American plants and animals and coincided with advances in coevolution of grasses and grazers. Paleosol records differ from global compilations of marine foraminiferal oxygen and carbon isotopic composition, due to competing influences of global ice volume and C₄ grass expansion. Paleosol records support links between global warming and high atmospheric CO₂.

Online enhancement: table.

Introduction

Paleosols are abundant and high-resolution archives of climates past (Retallack et al. 2004a) because climatic change affects soils in ways predictable from studies of pedogenesis (Retallack 2001b). Warming accelerates chemical reactions within soils, especially removal of alkali and alkaline earth cations by hydrolysis. Wet climates have ample moisture for hydrolysis of common nutrient cations (Ca²⁺, Mg²⁺, Na⁺, K⁺) from soil minerals in the most weathered parts of clayey soils (argillic or Bt horizons). The wetter and warmer the climate, the more base depleted are soils (Sheldon et al. 2002). Dry climates, on the other hand, have insufficient water to flush cations from soil into groundwater, with the result that alkaline earth cations (Ca²⁺, Mg²⁺) accumulate within subsurface calcite or dolomite nodules (soil calcic or Bk horizons). The drier the climate, the closer this subsurface horizon of pedogenic carbonate is to the surface. (Retallack 2005). These relationships within soils have been used to derive transfer func-

tions for estimating paleoclimatic variables from paleosols of comparable parent material composition and degree of development (Sheldon et al. 2002; Retallack 2005).

This study extends these techniques to paleosols formed during the past 40 m.yr. in Oregon, Nebraska, and Montana (figs. 1, 2). Paleosols provide a new view of terrestrial Cenozoic biotic and climatic evolution to supplement information from fossil plants (Gregory and Chase 1992; Wolfe 1994; Meyer and Manchester 1997) and mammals (Barnosky and Carrasco 2002; Janis et al. 2004) and global marine oxygen and carbon isotopic records (Miller et al. 1987; Zachos et al. 2001a).

Material and Methods

Paleosols are abundant in Cenozoic localities of North America that are famous for their fossil mammals and plants (fig. 1). This study is based on 892 observations of depth and spread of carbonate nodules and 162 major-element chemical analyses of clayey horizons in paleosols of eastern Oregon and southeastern Washington, considerably ex-

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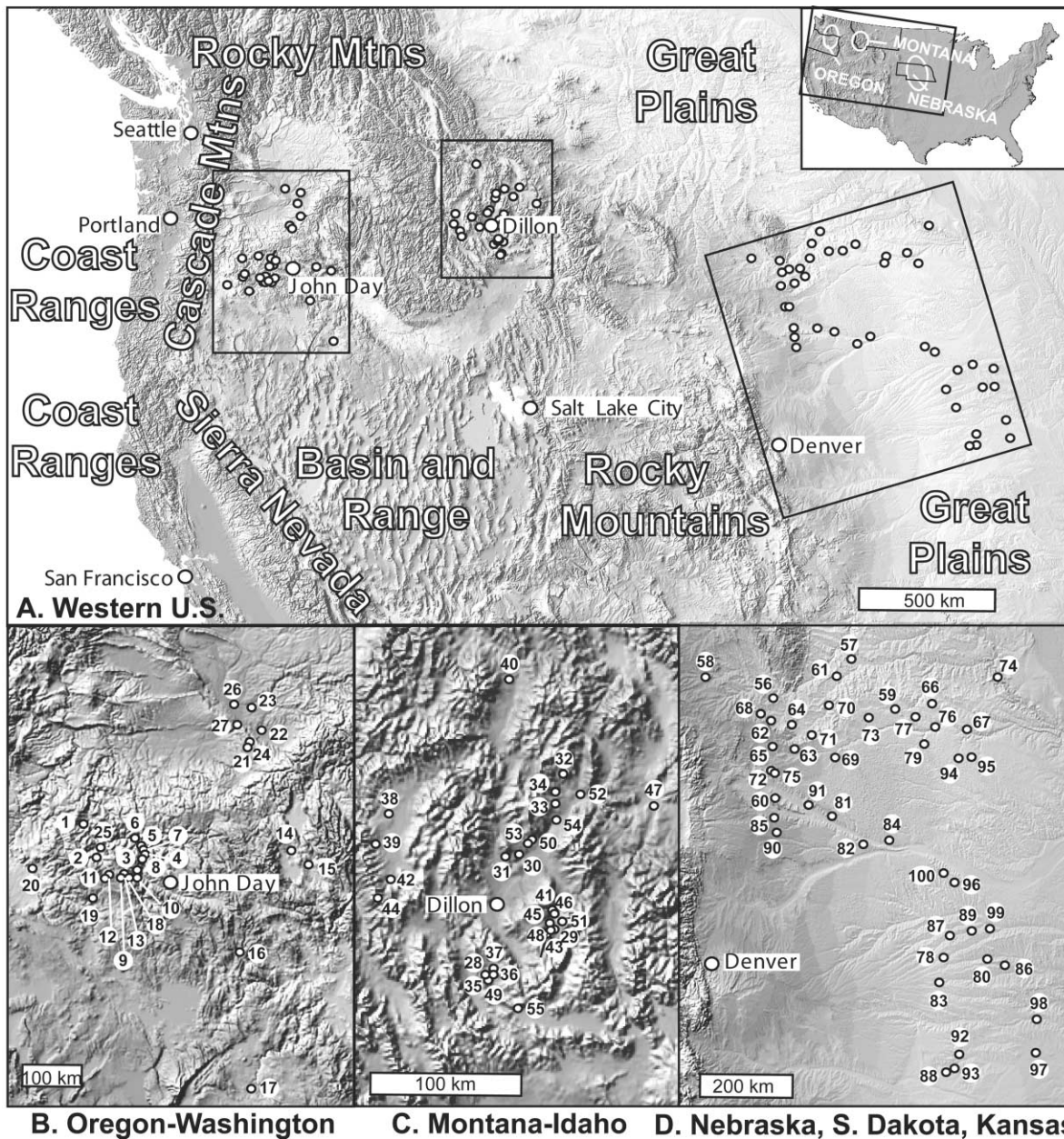


Figure 1. Examined localities for Cenozoic paleosols in central Oregon and southeastern Washington (*lower left*), southwestern Montana and eastern Idaho (*lower center*), and Nebraska, South Dakota, and Kansas (*lower right*). Base map is North America Color Shaded Relief DTED0 from Web site <http://nhdgeo.usgs.gov>, downloaded January 15, 2005. Examined localities are numbered as follows: 1, Clarno, OR; 2, Painted Hills, OR; 3, Longview Ranch airport, OR; 4, Roundup Flat, OR; 5, Kimberly, OR; 6, Spray, OR; 7, Bone Creek, OR; 8, Picture Gorge, OR; 9, Rock Creek, OR; 10, Mascall Ranch, OR; 11, Antone, OR; 12, Spanish Gulch, OR; 13, Birch Creek, OR; 14, Unity, OR; 15, Ironside, OR; 16, Juntura, OR; 17, Rome, OR; 18, Rattlesnake Creek, OR; 19, Logan Butte, OR; 20, Prineville, OR; 21, McKay Reservoir, OR; 22, Richland, WA; 23, Taunton, WA; 24, Pendleton, OR; 25, Sutton Mountain, OR; 26, Washtucna, WA; 27, Helix, OR; 28, Douglass Draw, MT; 29, Anderson Ranch, MT; 30, Diamond O Ranch, MT; 31, McCarty's Mountain, MT; 32, Easter Lily Mine, MT; 33, Little Pipestone Creek, MT; 34, Pipestone Springs, MT; 35, Hough Draw, MT; 36, Matador Ranch, MT; 37, White Hills, MT; 38, Mill Point, MT; 39, Everson Creek, MT; 40, Deer Lodge, MT; 41, Belmont Park Ranch, MT; 42, Maiden Creek, MT; 43, Virginia Springs, MT; 44, Railroad Canyon, ID; 45, Sweetwater Canyon, MT; 46, Barton Gulch, MT; 47, Madison Buffalo Jump, MT; 48, Timber Hill, MT; 49, Sage Creek, MT; 50, Beaverhead, MT; 51, Robb Creek, MT; 52, Whitehall, MT; 53, Giem Ranch, MT; 54, Silverstar,

tending published data (Retallack et al. 2000; Retallack 2004a, 2004b). Yet another 1194 observations of paleosol carbonate and 48 major-element chemical analyses of paleosols in southwestern Montana and adjacent Idaho are entirely new (downloadable Excel files available at the author's Web site: <http://www.uoregon.edu/~gregr/>). A third set of 677 carbonate paleosols and 48 major-element chemical analyses from Nebraska and adjacent states of South Dakota, Wyoming, and Kansas also considerably extends published analyses (Retallack 1983; Terry 2001). The 2763 paleosols and 258 chemical analyses compiled and newly reported here include only those with calcic or argillic horizons sufficiently developed to be indicative of paleoclimate; weakly developed paleosols were ignored for this analysis and have been described only for selected areas (Retallack 1983; Retallack et al. 2000).

Site selection for a time series from each region was a compromise between the need to obtain paleosols closely spaced in time and representative of Eocene to Holocene and the need for sites not so geographically dispersed as to sample different local climates. The greatest temporal resolution was found in Oregon (fig. 2A), which has a sequence of tuffs and lavas much thicker than those of Montana and Nebraska. The Montana sequence (fig. 2B) is almost as thick as Oregon's and is predominantly loessic and tuffaceous (Hanneman and Wideman 1991; Hanneman et al. 1994). Nebraska, in contrast, has a condensed sequence of loess and tuff with complex cut-and-fill sequences (fig. 2C). The Nebraska record is truncated at the base by a regional disconformity of late Eocene (Duchesnean) age on Late Cretaceous sediments (Terry 2001). In contrast, Oregon and Montana sequences extend further back into the Cenozoic than the sequences examined here (Retallack et al. 2000; Nichols et al. 2001).

The sites examined were mostly fossil mammal sites, constrained in age by mammalian biostratigraphy (Woodburne 2004), but precise ages were

assigned to each paleosol by linear interpolation of magnetostratigraphic and radiometric tie points within the specific measured sections of paleosols (table A1, available in the online edition or from the *Journal of Geology* office). These dated tie points were plotted against stratigraphic level for each of the 100 sites examined, and the numerical age (Ma) of each paleosol was interpolated by a site-specific linear regression (for graphical examples, see Retallack et al. 2000, 2004b; Retallack 2004a, 2004b). Some sites such as Oregon's Painted Hills (site 2 in figs. 1, 2; table A1) and South Dakota's Badlands National Park (site 57 in figs. 1, 2; table A1) have hundreds of paleosols; other sites, such as Nebraska's Kimball (site 90 in figs. 1, 2; table A1), have only one paleosol. The paleosols from all of the sites in each of the three regions examined were ordered by numerical age (Ma).

Three variables were measured for each paleosol: depth to Bk horizon (D in cm), thickness of soil with carbonate nodules (H in cm), and maximum size of carbonate nodules (N in cm). In addition, selected clayey (Bt) horizons were sampled and analyzed by x-ray fluorescence to determine chemical index of alteration without potash ($C = 100 \times \text{mAl}_2\text{O}_3 / [\text{mAl}_2\text{O}_3 + \text{mCaO} + \text{mMgO} + \text{mNa}_2\text{O}]$, where m is moles), and salinization index ($S = [\text{mK}_2\text{O} + \text{mNa}_2\text{O}] / \text{mAl}_2\text{O}_3$, where m is moles). These various field and chemical measures are related to mean annual precipitation (P in mm), seasonality (M as difference between wettest and driest monthly means, in mm), mean annual temperature (T in °C), and duration of formation of paleosols (A in kyr), by the following equations derived from large databases for modern soils (Sheldon et al. 2002; Retallack 2005):

$$P = 137.24 + 6.45D - 0.013D^2, \quad (1)$$

with $R^2 = 0.52$, $\text{SE} = \pm 147$ mm;

$$P = 221e^{0.0197C}, \quad (2)$$

MT; 55, Lima Dam, MT; 56, Toadstool Park, NE; 57, Badlands National Park, SD; 58, Douglas, WY; 59, Craven Creek, NE; 60, Scottsbluff, NE; 61, Sharps, SD; 62, Pants Butte, NE; 63, Pink Schoolhouse, NE; 64, Smiley Canyon, NE; 65, Agate, NE; 66, Mission, SD; 67, Norden, NE; 68, Eagle Crags, NE; 69, Hemingford, NE; 70, Chadron, NE; 71, Marsland, NE; 72, Mitchell, NE; 73, Martin, SD; 74, South Bijou Hill, SD; 75, Olcott Hill, NE; 76, Valentine, NE; 77, Big Spring, SD; 78, Morland, KS; 79, Merritt Dam, NE; 80, Ellis, KS; 81, Lisco, NE; 82, Ash Hollow, NE; 83, Scott Lake, KS; 84, Ogallala, NE; 85, Harrisburg, NE; 86, Hays, KS; 87, Clayton, KS; 88, Crooked Creek, KS; 89, Almena, KS; 90, Kimball, NE; 91, Broadwater, NE; 92, Meade, KS; 93, Borchers Badlands, KS; 94, Ainsworth, NE; 95, Long Pine, NE; 96, Eustis, NE; 97, Pratt, KS; 98, Great Bend, KS; 99, Phillipsburg, KS; 100, Bignell Hill, NE.

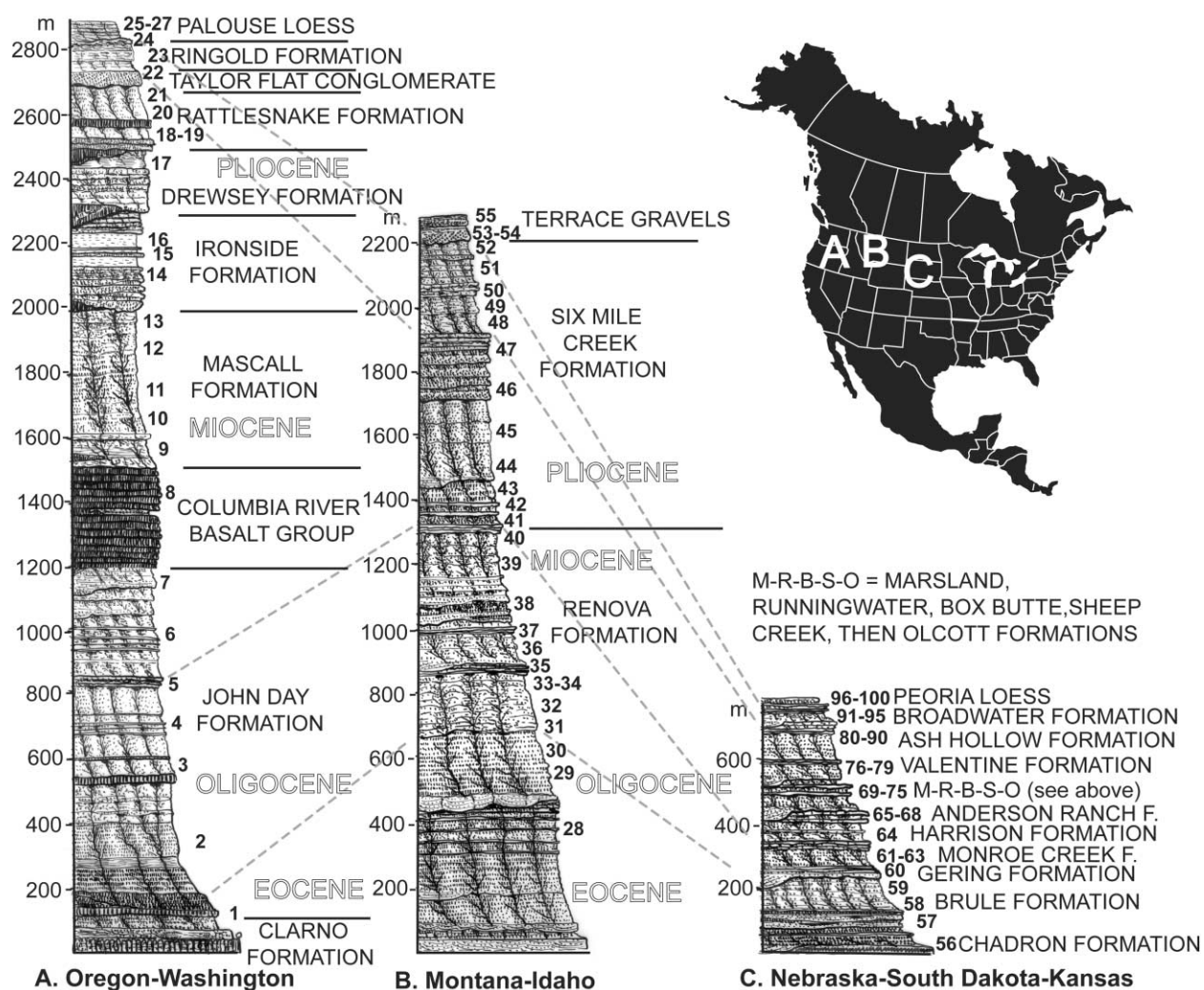


Figure 2. Outcrop sketches (not lithological logs) of the composite stratigraphic succession in Oregon, Montana and Nebraska (after Schultz and Stout 1980; Retallack 1983, 2004a, 2004b; Retallack et al. 2000; Nichols et al. 2001). Tuffaceous clayey badlands with paleosols in all three sections are divided by erosionally resistant paleochannel sandstones. Vertically jointed basalt and lava flows are mostly in the Oregon sequence. Stratigraphic levels of the 100 paleosol sites of figure 1 are shown. For details of the temporal range of sites, see table A1, available in the online edition or from the *Journal of Geology* office.

with $R^2 = 0.72$, $SE = \pm 182$ mm;

$$T = -18.5S + 17.3, \quad (3)$$

with $R^2 = 0.37$, $SE = \pm 4.4^\circ\text{C}$;

$$M = 0.79H + 13.71, \quad (4)$$

with $R^2 = 0.58$, $SE = \pm 22$ mm;

$$A = 3.92N^{0.34}, \quad (5)$$

with $R^2 = 0.57$, $SE = \pm 1.8$ kyr.

Age of soils was calculated to evaluate comparability of the duration, within 2–12 kyr, of soil formation. Comparison with modern soils also requires correction of depth to calcic horizon in paleosols for burial compaction, using standard algorithms (Sheldon and Retallack 2001) and geological estimates of overburden (table A1). Such compactional adjustments do not have a large effect on paleoprecipitation estimates from depth to pedogenic carbonate, because paleoclimatic variation is clear from depth variation in the field. Correction for higher ancient atmospheric CO_2 levels was not applied, because increases from two to 11 times the

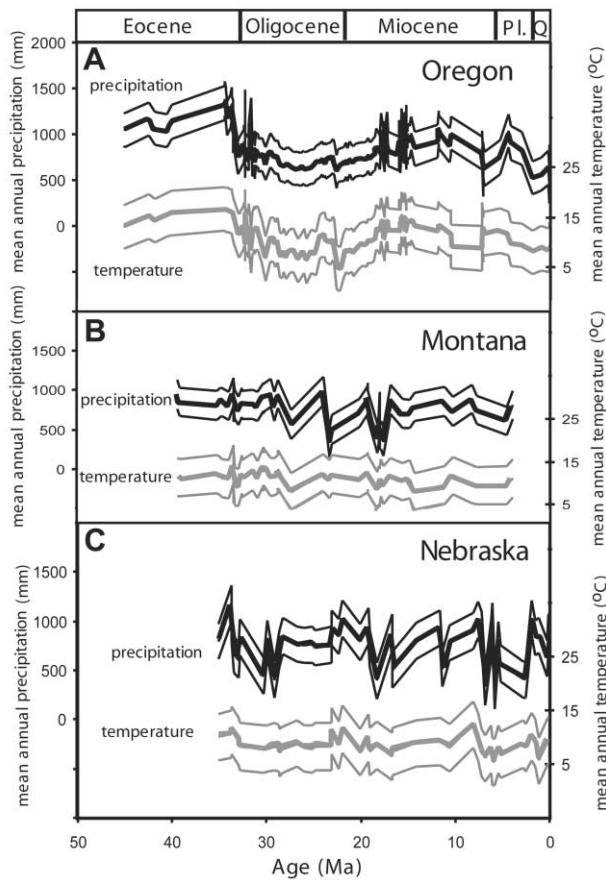


Figure 3. Time series of mean annual precipitation and mean annual temperature inferred from individual paleosols using transfer functions for geochemical composition of Bt horizons (for data, see table A1, available in the online edition or from the *Journal of Geology* office and author's Web site). Flanking lines are 1 SE of the transfer function (Sheldon et al. 2002).

present level have been modeled (by McFadden and Tinsley [1985]) to increase depth to pedogenic carbonate by only 5 cm. This results in a difference in mean annual precipitation of only 25 mm for carbonate depth of 50 cm and 18 mm for carbonate depth of 100 cm: both are well within the standard error of equation (1) above. Stomatal index studies of fossil ginkgo leaves, as well as other CO₂ paleobarometers, indicate Cenozoic CO₂ levels no more than 10 times present levels (Retallack 2002; Demicco et al. 2003).

Warmer Times Were Also Wetter

Paleosol proxies reveal that times of paleoclimatic warmth were also times of high precipitation. This is indicated not only by the roughly parallel tracks

of paleosol proxies of mean annual precipitation and mean annual temperature (fig. 3) but by significant correlation between these proxies within the paleosols studied (fig. 4). Paleoprecipitation estimates from chemical composition of paleosols agree with estimates independently obtained from depth to Bk horizon (fig. 5). Each time series includes base-depleted or deep calcic paleosols at

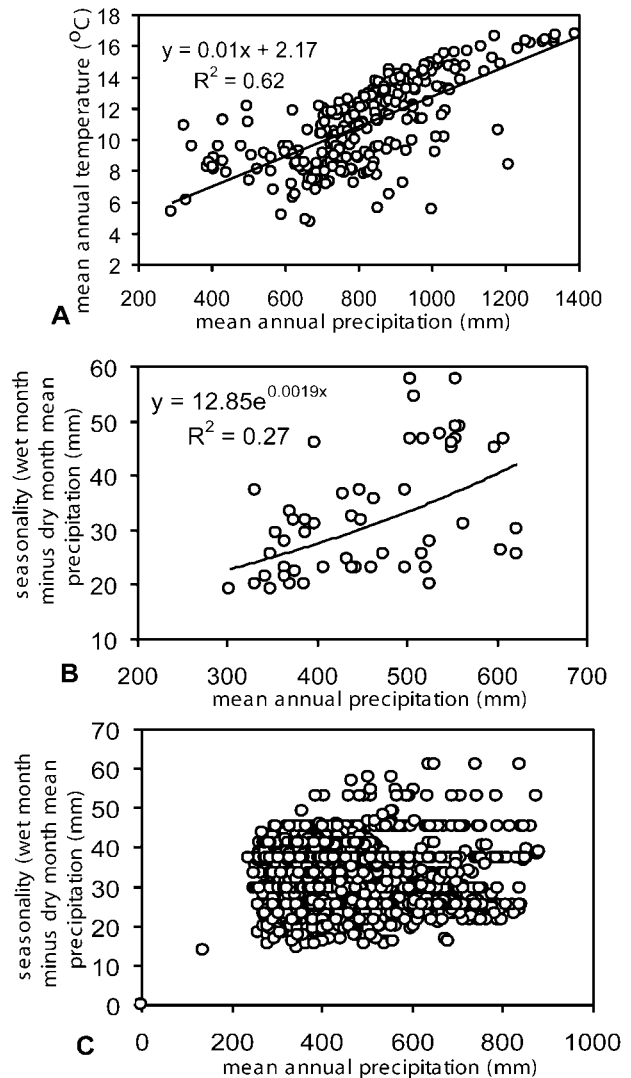


Figure 4. A, Correlations between mean annual precipitation and mean annual temperature, estimated from paleosol Bt horizon chemistry; B, C, mean annual range of precipitation and mean annual precipitation from carbonate nodules in paleosols from whole data set (C) and only data from closest to the monsoonlike paleoclimatic region of the Gulf Coast (B). Data for B are from Meade County, Kansas (sites 88, 92, 93 of figs. 1, 2; table A1, available in the online edition or from the *Journal of Geology* office).

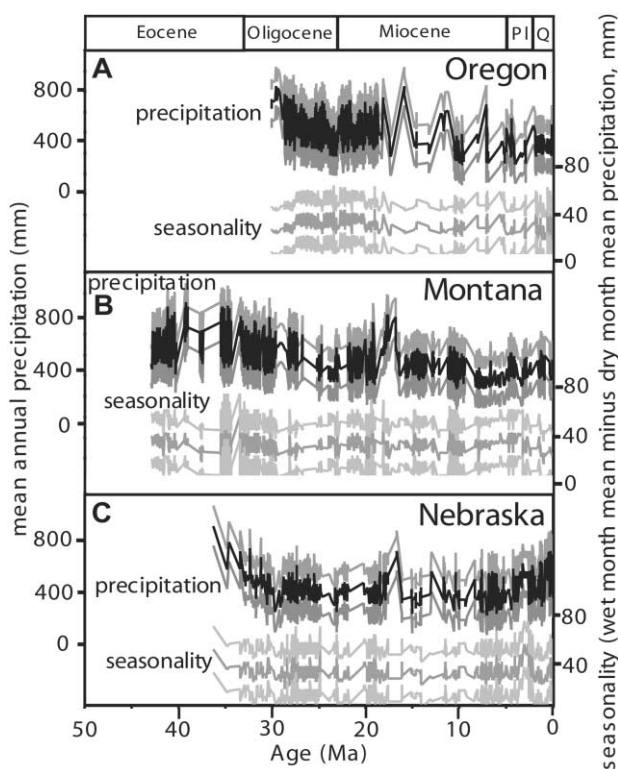


Figure 5. Time series of mean annual precipitation and of seasonality (difference between means of wettest and driest months) from depth to Bk and thickness of paleosol with carbonate nodules from Oregon, Montana, and Nebraska (for data, see table A1, available in the online edition or from the *Journal of Geology* office and author's Web site). Calcic soils can reveal only dry climate because humid-climate soils are noncalcareous (Retallack 2001b). Flanking lines are 1 SE of the transfer function (Retallack 2005).

global warm-wet spikes (fig. 6): latest Eocene (35 Ma), middle Miocene (16 Ma), late Miocene (7 Ma), and early Pliocene (4 Ma). There are thick caliche caprocks in each region at unusually dry and cool times (fig. 6): terminal Oligocene (23.5 Ma), late Miocene (10 Ma), and terminal Miocene (5 Ma).

In addition, paleosol proxies indicate long-term paleoclimatic volatility during times of generally warm conditions, such as the late Eocene (35 Ma) and middle Miocene (16 Ma). These times had warm-wet transients but also cold-dry spells on timescales of only a few million years (figs. 3, 5). This is particularly striking in the middle Miocene (Barstovian) Sheep Creek Formation of southern Sioux County, Nebraska (Skinner et al. 1977), where deep calcic clayey paleosols alternate with shallow calcic silty paleosols. Late Eocene and middle Miocene were also peaks of paleosol diversity

(Retallack et al. 2000; Bestland and Forbes 2003), in contrast with long periods of limited paleosol diversity and less dramatic paleoclimatic fluctuation within equally densely sampled intervals (Retallack 2004b; Retallack et al. 2004a).

A terminal Paleocene paleoclimatic warm spike may have been a time of more seasonal as well as warm and wet paleoclimate (Retallack 2005), but correlated seasonality, warmth, and precipitation increases are not apparent from the data presented here (figs. 3, 5). Seasonality shows no correlation with precipitation (fig. 4C), even in the sites closest to the Gulf Coast, which now has monsoonlike summer humidity and rainfall (data from most southern localities near Meade, Kansas, plotted only in fig. 4B). Most of these paleosols formed north of monsoonal tropical regions where seasonality effects are most marked. Paleocene paleosols (Retallack 2005) and their unusually thick enclosing synorogenic conglomerates (Talling et al. 1994) may indicate monsoon-inducing paleorelief similar to that of Neogene Pakistan and Kenya (Retallack 1991). In contrast, post-Laramide paleogeography of North America many have been more subdued and oriented orthogonally to warm waters of the Gulf of Mexico (Poage and Chamberlain 2001), so that paleoclimate was less extremely monsoonal.

North American Desert Shrubland Is Ancient

One surprising result of this paleosol survey was the discovery of paleosols (fig. 7) similar to those now forming under desert shrubland in Montana during the late Eocene (35 Ma), which was a time of unusually warm and wet paleoclimate elsewhere in North America (figs. 3, 5). These paleosols not only had weakly pedal structure, low clay content, limited chemical weathering, and shallow calcic horizons of aridland soils but contained trace fossils of distinctive back-filled burrows (ichnogenus *Taenidium*; fig. 7D), similar to those made by obligate-sagebrush cicadas (*Okanagia*) in modern and Pleistocene soils of eastern Washington (O'Geen and Busacca 2001). The antiquity of desert shrubland is also indicated by late Eocene pollen of aridland taxa such as Mormon tea (*Ephedra*) and greasewood (*Sarcobatus*) in Montana, Wyoming, and Colorado (Leopold et al. 1992). Comparable desert shrubland paleosols, which presumably supported very different plant species, are found in rocks of Paleocene to Pennsylvanian age in Utah and Colorado (Retallack 1995, 2001a, 2005).

Montana has been dry within an intermontane rain shadow since at least 43 Ma, indicated by late Eocene (Uintan) paleosols in Douglass Draw (site

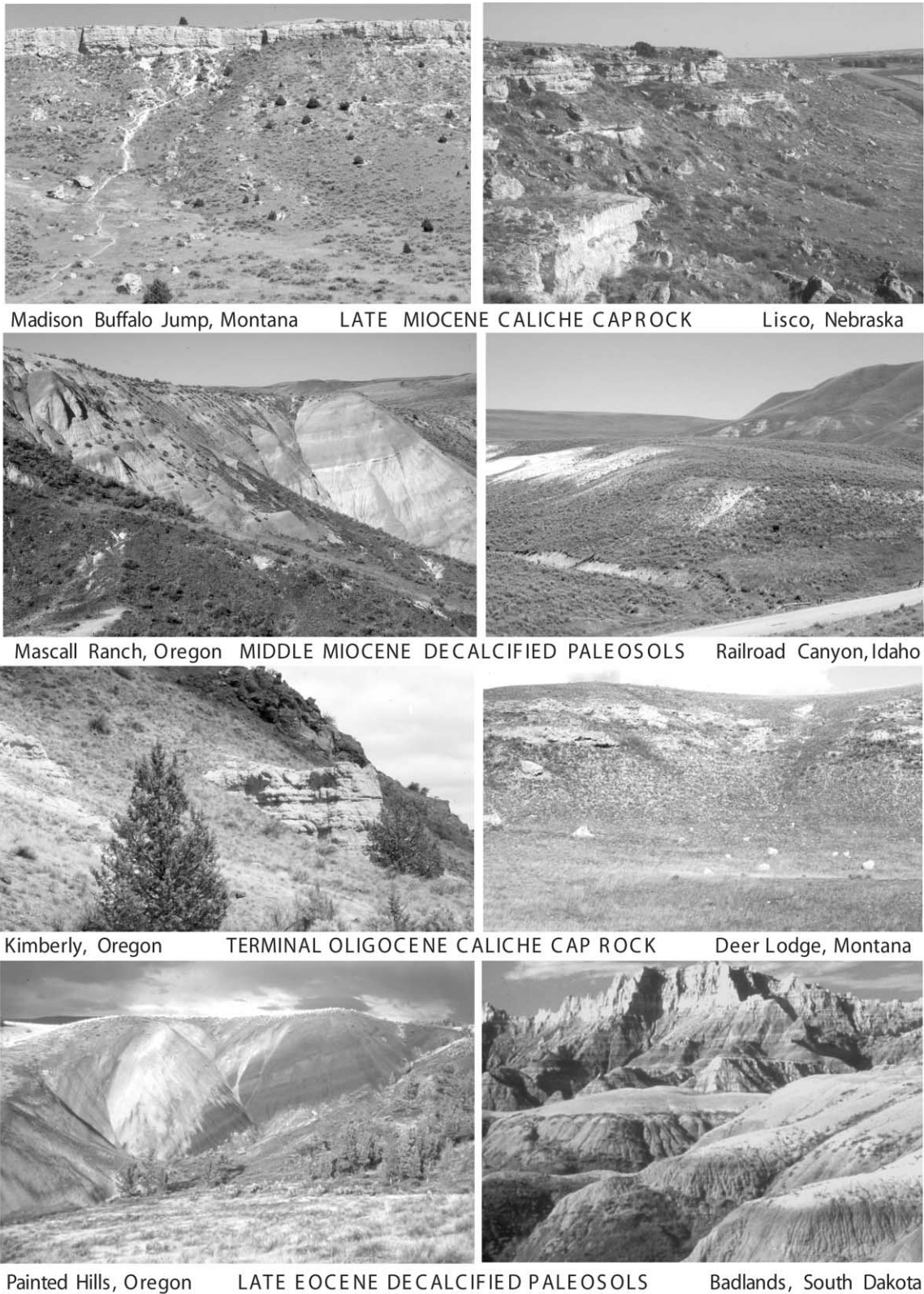


Figure 6. Field views of decalcified paleosols of wet climatic phases (*bottom and upper middle*), and of caliche caprocks of dry climatic phases (*top and lower middle*). Paleosol records of paleoclimate are widely exposed throughout the western United States.

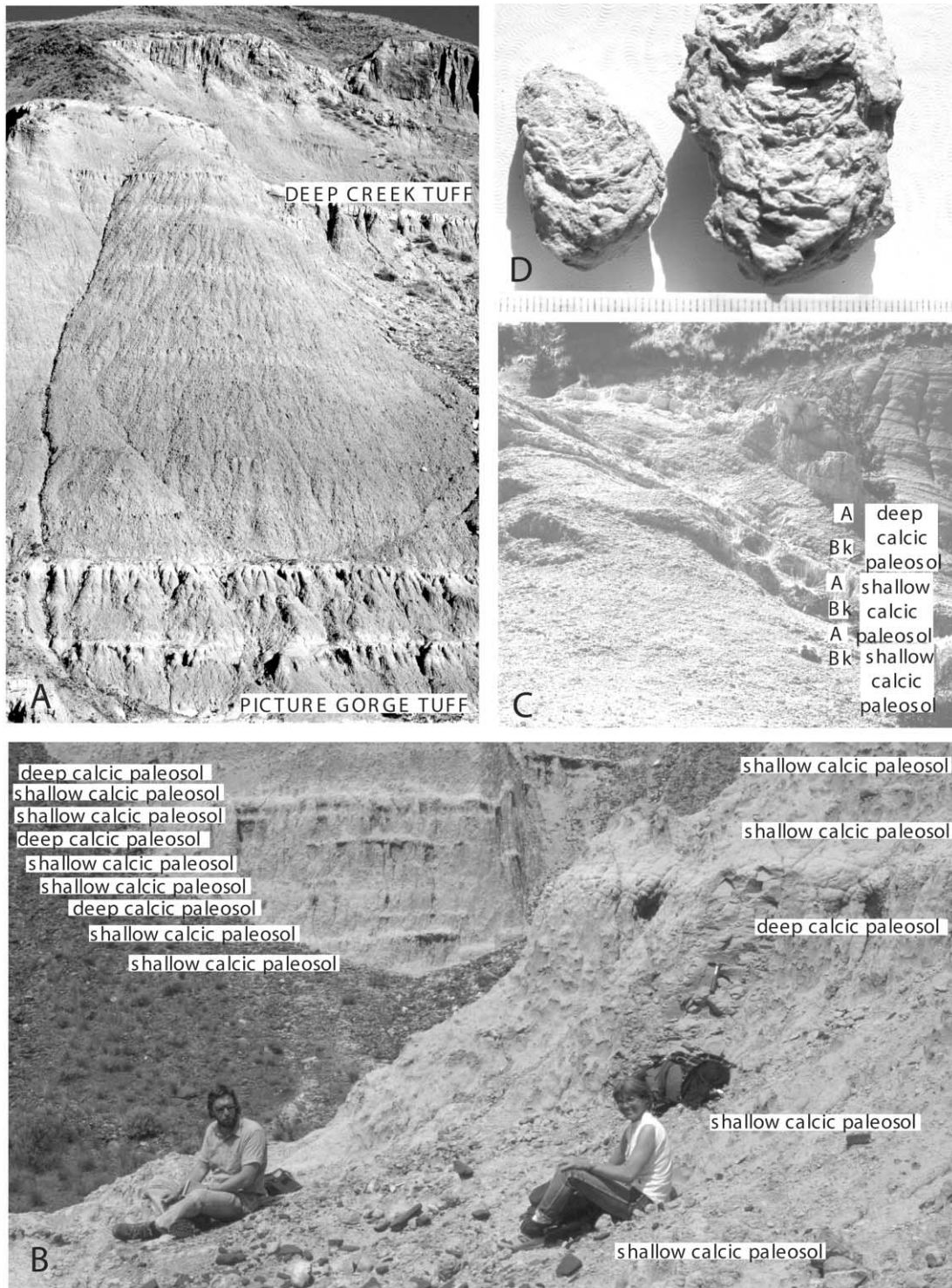


Figure 7. Milankovitch triplets of paleosols in the field (A–C) and a *Taenidium* burrow (D) from a shallow calcic paleosol in Montana taken as indication of desert shrubland vegetation (following O’Geen and Busacca 2001). Obliquity (40 kyr) triplets of paleosols at Foree (B; Ted Fremd and Andrea Mindszenty for scale) and in a 62-m section of 76 paleosols and 23 triplets between dated tuffs at Longview Ranch airport (A), Late Oligocene, John Day Formation of Oregon (Retallack 2004a; Retallack et al. 2004a). The *Taenidium* burrow (D) and two shallow calcic paleosols and a deep calcic paleosol (C) are in the late Eocene, Renova Formation, Little Pipestone Springs, Montana (Tabrum et al. 1996; Nichols et al. 2001). Scale at base of D is graduated in millimeters.

28 of fig. 1; table A1) that have shallow calcic horizons within loesslike tuffaceous sediments. This is a minimum age for the rain shadow in Montana because shallow calcic paleosols were found in Bridgerian sandstones of the lower Renova Formation (Nichols et al. 2001) and in Late Cretaceous to Paleocene red conglomerates of the Beaverhead Conglomerate (Dyman et al. 1995), but local biostratigraphic, radiometric, and paleomagnetic dating is inadequate to demonstrate continuously arid conditions so far back in time. In contrast, calcareous paleosols and loesslike tuffaceous sediments date back only 35 Ma in Nebraska and 30 Ma in Oregon (figs. 5, 7). Similarly, the Chinese loess record of aridity extends back only 22 Ma (Guo et al. 2002). These differences are consistent with the idea that noncalcareous, clayey, red paleosols of near-coastal Nebraska, Oregon, and China were influenced by moist maritime air masses from the Gulf Coast and Pacific Ocean early during the Cenozoic, but these air masses retained little moisture by the time they reached intermontane southwest Montana.

Climatic Signal Enhanced by Tectonism and Volcanism

Climatic drying in North America has been attributed to uplift of mountain ranges created by terrane accretion and continental-arc volcanism (Kohn et al. 2002). Rain shadow effects of Cascade and Rocky Mountain uplift can now be inferred from comparison of paleoprecipitation time series of Oregon with that of Montana and of Montana with that of Nebraska (Sheldon and Retallack 2004). Rain shadow was calculated as the gradient of precipitation change ($\text{mm} \times \text{km}^{-1}$) between the studied regions for million-year increments of each paleoprecipitation time series, with allowance for significant post-Miocene (16 Ma) tectonic extension affecting distances between the regions (Atwater and Stock 1998). A high positive gradient indicates rapid eastward decline of precipitation away from the site. A low negative gradient indicates gradual increase of precipitation beyond the rain shadow toward humid eastern North America. During late Eocene (35 Ma) warm-wet times, only Montana was arid, and both Oregon and Nebraska were humid-subhumid, indicating a narrow rain shadow in the central Rocky Mountains. By the Oligocene (30 Ma), the rain shadow cast by the Cascades connected with intermontane rain shadows of the Rocky Mountains and western Great Plains, as it does today. Miocene warm-wet maxima narrowed the rain shadow to only the central Rocky Moun-

tains at least three times (fig. 8A). Maritime warm-wet air masses encroached upon the margins of the continent without large effects in the continental interior, thus steepening the rainfall gradient into the dry interior during warm-wet times.

Variation on ca. 2-m.yr. wavelengths in the extent of the western United States montane rain shadows (fig. 8A) is unlikely to be caused by tectonic forcing. Volcanic and other tectonic cycles are on longer temporal wavelengths. The Western Cascades volcanic arc was in place by 42 Ma (Retallack et al. 2000), well before the late Eocene (35 Ma) warm-wet spike and long before cooling and drying of central Oregon by 30 Ma. Synorogenic conglomerates in Montana (Dyman et al. 1995) and Utah (Talling et al. 1994) indicate mountains by the mid-Cretaceous, well before the paleosol records studied here. Greater than modern Late Cretaceous–Paleocene paleoelevation of the North American Cordillera is revealed by oxygen isotopic studies of authigenic and pedogenic carbonate, clays, and chert (Chamberlain and Poage 2000; Poage and Chamberlain 2001) and hydrogen isotopic studies of meteoric alteration (Mulch et al. 2004). Foliar physiognomy of fossil plants indicate that the Col-

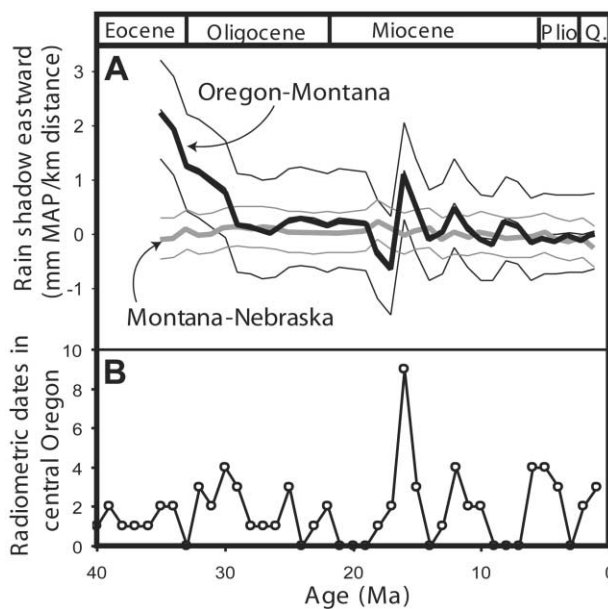


Figure 8. A, Time series of Cascade and Rocky Mountain rain shadow variation, estimated from comparison of mean annual precipitation for million-year increments from Oregon to Montana and Montana to Nebraska (from paleosol Bt chemical data); B, number of radiometric dates as a proxy for volume of volcanic eruptions in the central Cascade Range of Oregon (data from McBirney et al. 1974; Retallack et al. 2004b).

orado Front Range had a paleoelevation comparable with that of today (2500 m) during the late Eocene (35 Ma; Gregory and Chase 1992) and that the Nevada Basin and Range was higher (3000 m) during the middle Miocene (15–16 Ma) than today (2000 m; Wolfe et al. 1997). Orographic rain shadows lengthen and shorten with global paleoclimatic change on million-year timescales, but the mountains themselves rise and fall on timescales of tens of millions of years.

Other indications that paleosols recorded global paleoclimate rather than tectonic or volcanic signals come from paleosol records of Milankovitch-scale paleoclimatic fluctuations (figs. 7, 9), known to be produced by global orbital variations (Retallack et al. 2004a). Orbital obliquity cycles (41 kyr) are the most common Milankovitch fluctuations observed in paleosol sequences of appropriate exposure and age control (fig. 9). Such fine temporal

resolution is more common in thick intermontane Oregon and Montana records than in the condensed Nebraska record (fig. 9D). Paleosol sequences of Badlands National Park, South Dakota (Retallack 1983, 1986), have only a single paleosol for each eccentricity cycle (100 kyr). Climatic variation at these timescales and frequencies are cued to variations in Earth's orbit such as precession, obliquity, and eccentricity and so are global, not local.

Volcanic activity did not induce rain shadows either, because peaks in volcanic activity inferred from abundance of radiometric ages (fig. 8B) coincide with warm-wet rather than cold-dry paleoclimatic episodes (figs. 3, 5). Marked middle Miocene (16 Ma) warmth and humidity in Oregon coincides in time with voluminous eruptions of flood basalts in the Columbia River and Snake River basins (fig. 8B), which were low-relief, valley-filling flows with little potential to cast a rain shadow (Humphreys

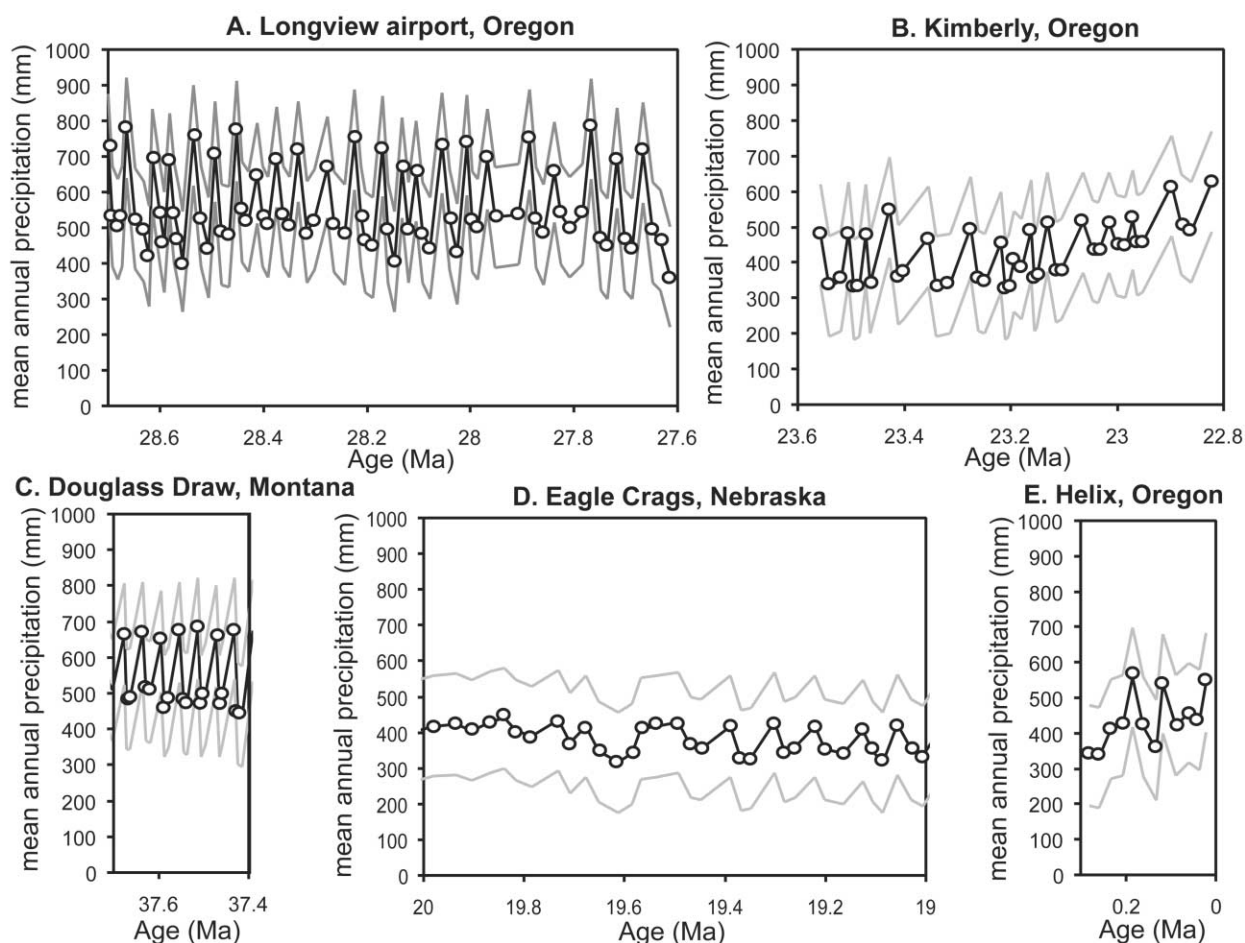


Figure 9. Milankovitch-scale variation in mean annual precipitation inferred from well-dated (table A1, available in the online edition or from the *Journal of Geology* office) sequences of depths to Bk horizons at a variety of sites, all plotted to the same temporal scale. Gray flanking lines are 1 SE from transfer function (Sheldon et al. 2002).

et al. 2000). These flood basalts and the late Eocene (35 Ma) volcanism that initiated the Western Cascades Range (Retallack et al. 2004b) may have induced climatic warming by releasing greenhouse gases such as carbon dioxide and water vapor (Courtillot 2002) or by intruding and mobilizing methane from clathrate reservoirs (Svensen et al. 2004) or heating of coal seams (McElwain et al. 2005). Also plausibly inducing climatic greenhouse effects were biomass destruction from asteroid impacts, such as the 35-Ma Chesapeake Crater (Poag et al. 2003) and the 16-Ma Ries and Steinheim craters (Stoeffler et al. 2002). Another possibility is that greenhouse paleoclimates were produced by accelerated sea-floor spreading, although compilations of spreading rate do not document significant change at 35 and 16 Ma (Rowley 2002). Both 35 and 16 Ma were peaks of warmth, evident from a network of deep-sea cores (Zachos et al. 2001a), and were thus global paleoclimatic perturbations.

Mountain building and volcanic activity greatly improved resolution of the paleoclimatic signal of paleosols by creating rapidly subsiding intermontane basins and long-term dusty and dry rain shadows. Thick volcanoclastic sequences in Oregon and Montana have fine temporal resolution (Retallack et al. 2004a) and abundant dateable tuffs and lavas (Fritz et al. 1989, 1992; Retallack et al. 2000; Retallack 2004a, 2004b). Intermontane basins with high sediment accumulation rates provide unusually detailed records of global climatic fluctuations (fig. 9).

Comparison with Fossil Plant Record

Paleoclimate revealed by paleosols agrees with that inferred from fossil plants. In central Oregon, mean annual precipitation (MAP) and temperature (MAT) estimated from leaf physiognomic characters for the early Oligocene (32 Ma) Bridge Creek flora (1000–1500 mm MAP, 3°–9°C MAT; Meyer and Manchester 1997) and middle Miocene (16 Ma) Mascall flora (ca. 1270 mm MAP, ca. 17°C MAT; Chaney and Axelrod 1959) are comparable with estimates from associated paleosols (figs. 3, 5). In southwestern Montana, the late Eocene (36 Ma) Horse Prairie flora has been taken to indicate MAP of 790 mm and MAT of 12.5°C (Axelrod 1998), again comparable with indications from paleosols of the same age, known from McCarty's Mountain, Montana (site 31 of fig. 1; table A1). The Miocene (12 Ma) Kilgore flora of Nebraska has been taken as evidence of 760–890 mm MAP and frost-free temperatures (Axelrod 1985), again similar to paleosols near Valentine, Nebraska (site 76, fig. 1; table A1).

Paleobotanical estimates of paleoclimate are few not only because they are time consuming (Gregory and Chase 1992; Wolfe 1994) but because fossil leaves and pollen of suitable preservation require lakes and swamps, which were uncommon and unrepresentative of the vegetation of Cenozoic aridlands of western North America (Leopold et al. 1992).

More voluminous information about Cenozoic vegetation change comes from paleosols: their profile form and root traces (Retallack 1983, 2004a, 2004b; Retallack et al. 2000; Terry 2001), carbon isotopic composition of pedogenic carbonate (Fox and Koch 2003), and A horizon accumulations of opal phytoliths (Strömberg 2002, 2004). All lines of evidence indicate open grassy vegetation well back into the late Eocene. Isotopic values of pedogenic carbonate carbon are unusually high for paleosols and indicate at least 20% C₄ plants (largely warm-climate grasses) back to 38 Ma in the Great Plains (fig. 10D), with a dramatic increase after 6.4 Ma (Fox and Koch 2003). Many North American Miocene grasses were and are C₃ plants, with isotopically low values indistinguishable from those of woody plants (MacFadden 2000; Smith and White 2002). Comparison of Cenozoic phytolith assemblages from the Great Plains with the dicot/grass ratio of assemblages from Africa indicate open shrub steppe back to 35 Ma (Strömberg 2002). Phytoliths of grasses constitute 38%–54% of late Eocene (35 Ma) phytoliths (Strömberg 2004), although some of these (5%–8% of total) are primitive bambusoid grasses that today are generally found in shade. Other rare (2%–3% of total) phytoliths are blocky polyhedrons similar to those of sage (*Artemisia*; Blinnikov et al. 2002), though Strömberg (2004) interprets them as evidence of forest. By the middle Miocene (16 Ma), bambusoid grasses drop out entirely, sagebrush phytoliths remain low in abundance, and grasses dominate phytolith assemblages of the Great Plains (fig. 10E).

Paleosols themselves are also evidence of vegetation, and the relative abundance of different pedotypes gives a detailed record of vegetation change over the past 45 m.yr. (fig. 10A–10C). Sod grassland paleosols, for example, have abundant fine root traces and crumb ped structure, whereas bunch grassland paleosols have fine root traces and other traces of grasslands, such as earthworm chimneys and dung beetle boli, but also have a granular ped structure (Retallack 2004b). Sagebrush paleosols, in contrast, have very shallow calcic horizons, platy ped structure, and burrows of cicadas (fig. 7D; O'Geen and Busacca 2001; Retallack et al. 2004a). Woodland and forest are distinguished by size of

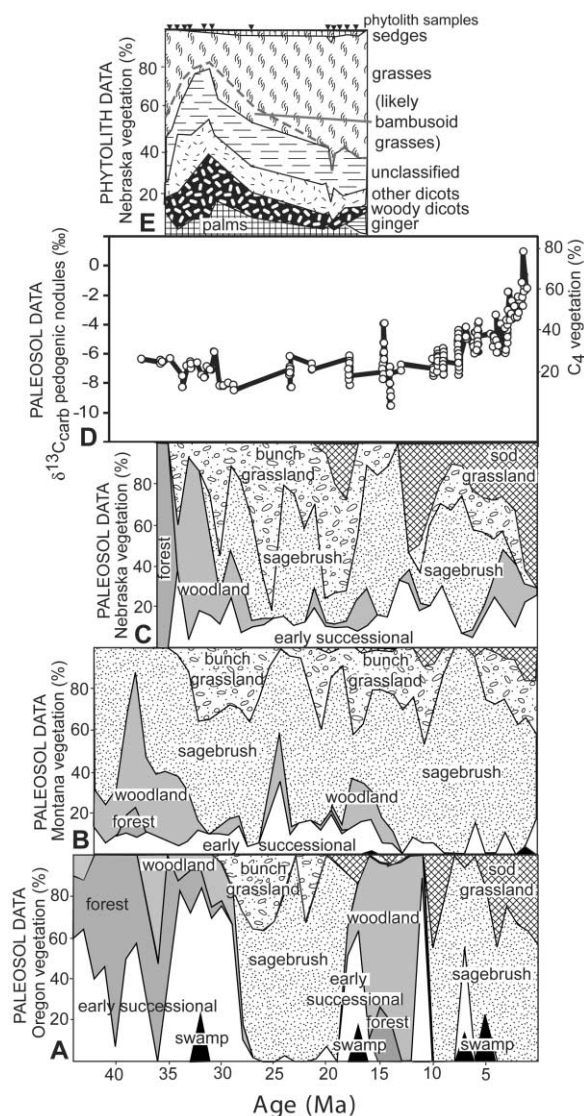


Figure 10. Time series of North American Cenozoic vegetation, inferred from proportion of paleosols by running mean of 2-million-yr increments in Oregon (A), Montana (B), and Nebraska (C), compared with isotopic composition of pedogenic carbonate and inferred proportion of C_4 grasses in the Great Plains (D), and relative abundance of silica phytoliths in Nebraska (E). Data for A–C are in table A1, available in the online edition or from the *Journal of Geology* office and author's Web site; data for D are from Lander (1990), Gardner et al. (1992), and Fox and Koch (2003); data for E are from Strömberg (2004).

root traces and by geochemical and clay mineralogical proxies for soil fertility (Retallack et al. 2000). Swamp paleosols are lignites and coals, and early successional soils have root traces as well as abundant sedimentary bedding (Retallack 2001b).

Montanan paleosols are evidence for the antiquity and abundance of sagebrush vegetation (fig. 10B). As in comparable paleosol sequences in Oregon (Retallack et al. 2004a), sagebrush paleosols formed in dry phases of Milankovitch-scale obliquity cycles (fig. 9C). In contrast, Nebraska and Oregon have mainly forested paleosols as late as 35 and 37 Ma, respectively, and mainly woodland paleosols as late as 31 and 30 Ma, respectively (fig. 10A, 10C). Middle Miocene warm-wet climates coincided with resurgence of woodland in Oregon but lesser woodland expansion in Montana and Nebraska.

These new compilations of Cenozoic rangeland vegetation (fig. 10) do not support the widely held supposition (e.g., by Zachos et al. 2001a; Kohn et al. 2002) that grassland ecosystems spread with paleoclimatic drying and cooling. Dry-cool conditions coincide with sagebrush expansion. Evidence from phytoliths, pedogenic carbon isotopic values, fine root traces, and granular ped structure indicates that bunch grasslands expanded in Nebraska and Montana during the latest Eocene (35–34 Ma; Terry 2001) and in Oregon during the early Oligocene (30–27 Ma; Retallack et al. 2000), which were both wet and warm times (figs. 3, 5). An expansion of bunch grassland at the expense of sagebrush is seen in all three areas during the minor warm-wet spike of 25–24 Ma (fig. 10A–10C). Short sod grasslands are evident from grass phytoliths, a minor spike in C_4 carbon isotopic signatures, fine root traces, crumb structure, and shallow calcic horizons. Short sod grasslands appear and expand in all three areas during the early to middle Miocene (19–16 Ma; Retallack 2004b), again at a wet-warm spike (figs. 3, 5). Tall sod grassland evident from deep fine root traces, crumb peds, and widespread appearance of C_4 carbon isotopic signatures in both teeth and pedogenic nodules appear during another Miocene (7–6 Ma) warm-wet spike (Retallack et al. 2002). Sagebrush scrub expanded in cool dry climates, not grasslands.

Bunch grassland and then sod grassland ecosystems appeared for the first time among more ancient kinds of vegetation and then sequentially expanded their range despite fluctuating paleoclimate. Successive appearances and expansion of bunch grassland at 35–34 Ma, sod-forming short grassland at 19–16 Ma, and sod-forming tall grassland at 7–6 Ma were each followed by long-term climatic drying and cooling during the Oligocene (33–23 Ma), late Miocene (15–10 Ma), and Pliocene–Pleistocene (5–0 Ma), respectively. These data favor the view that grasslands evolved at a pace determined not by paleoclimatic drying but by grass-

grazer coevolution that enhanced grassland soil carbon storage, water retention, and albedo (Retallack 2001a).

Comparison with Fossil Mammal Record

High precipitation and temperature generally correspond with high mammalian diversity and low precipitation and temperature with low diversity, so that Alroy et al. (2000), Barnosky and Carrasco (2002), and Prothero (2004) were surprised to find that mammalian diversity does not track the marine oxygen isotope curve of Zachos et al. (2001a). However, mammalian evolution does track paleosol-based estimates of MAP and MAT (fig. 11). Late Eocene aridland faunas of Montana include abundant small marsupials, insectivores, and rodents but are unique in lacking comparably sized primates common in other early Cenozoic communities of the American West (Nichols et al. 2001). Titanotheres megaherbivores and other archaic elements were most diverse during a late Eocene (35–34 Ma) warm-wet spike (Mihlbachler et al. 2004). This was followed by a long ramp of declining precipitation and temperature from 33 to 30 Ma, apparent in the paleosol record (figs. 3, 5). During this early Oligocene interval, titanotheres and other archaic taxa became extinct (Prothero 2004). The middle Miocene (19–16 Ma) warm-wet interval coincides with a remarkable adaptive radiation of ungulates (Janis et al. 2004). Some middle Miocene endemic clades diversified (Barnosky 2001), but diversification was enhanced by immigration of gelocid deer (*Pseudoceras*) and proboscideans (*Zygodon*) from Asia at about 16 Ma (Webb et al. 1995; Woodburne 2004), presumably through an unusually warm Beringian land bridge. Cooling, drying, and expansion of sod grasslands during the late Miocene to Pliocene (7–5 Ma) reduced ungulate diversity (Janis et al. 2002).

Climatic control of mammal faunas is also apparent on shorter timescales of millions and thousands of years. Martin (1994) and Meehan (1998) have argued that faunal overturn and iterative evolution of comparable ecomorphs between North American land mammal ages spaced at about 2.3-m.yr. intervals are correlated with climatically controlled cutting-and-filling cycles demonstrated by Schultz and Stout (1980) and Retallack (1986, 1998). Cycles of 2.4 m.yr. emerge from other paleoclimatic records (Olsen 1997) and orbital simulation of insolation variation (Laskar et al. 2004). On timescales of Milankovitch obliquity cycles (41 kyr), mammal communities are alternately dominated by different mouse-deer genera, *Nanotrugu-*

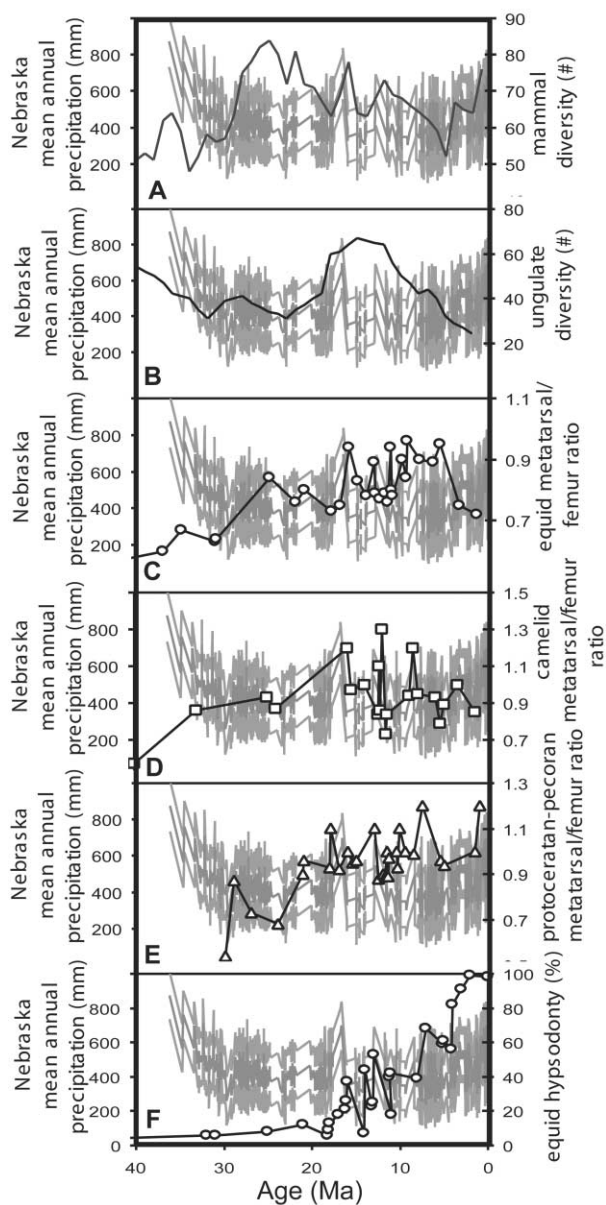


Figure 11. Time series for North American mammalian generic diversity (A), ungulate generic diversity (B), equid metatarsal/femur ratio (C), camelid metatarsal/femur ratio (D), protoceratan and pecoran metatarsal/femur ratio (E), and equid hypsodonty (F) compared to mean annual precipitation inferred from depth to Bk in Nebraska (gray in background of each frame from fig. 5C). Mammalian data sources: A, Janis et al. (2002); B, Alroy et al. (2000); C–E, Janis and Wilhelm (1993); F, MacFadden (1992).

lus in subhumid grassland paleosols and *Hypertragulus* in semiarid shrubland paleosols, some 105 times in 5.1 m.yr. of the upper John Day Formation of Oregon (Retallack et al. 2004a). The fossil record

of mammals follows paleoclimatic variation at all timescales observable from paleosols.

Grassland mammals and their various adaptations to grazing and running in open grassy vegetation also show patterns of first appearance that bear on the question of grass-grazer coevolution, which may have been a force for climate modification by virtue of lowered transpiration, higher albedo, and greater carbon storage than in earlier shrubland vegetation (Retallack 2001a). A significant faunal modernization, largely by immigration from Asia (Lucas et al. 2004), came at the late Eocene (39 Ma) temperature and precipitation maximum in Oregon and Montana (figs. 3, 5). A subsequent dip in temperature and precipitation from 38 to 35 Ma corresponds with marked increases in metatarsal/femur ratio, and thus running stride (cursoriality), in horses (equids of fig. 11C). This was followed by a long declining ramp of precipitation, temperature, and replacement of woodland by grassland from 40 to 33 Ma, when marked advances in cursoriality appeared in camels and deer-like protoceratids (fig. 11D, 11E). The middle Miocene (19–16 Ma) warm and wet interval coincides with a remarkable adaptive radiation of ungulates (fig. 11B), including the first horses with high-crowned (hypsodont) teeth of grazers (fig. 11F), and modest increases in cursoriality in horses, camels, and deer (fig. 11C–11E). Monodactyl horses appear during a late Miocene (7.5 Ma) warm-wet spike (Retallack et al. 2002). Cooling, drying, and expansion of sod grasslands during the latest Miocene to Pleistocene then reduced ungulate diversity (fig. 11B) and increased equid hypsodonty to modern levels (fig. 11F).

Similar to the record of rangeland vegetation (fig. 10), that of mammals (fig. 11) does not support the widely held supposition (e.g., by Zachos et al. 2001a; Kohn et al. 2002) that grassland ecosystems adapted to paleoclimatic drying and cooling. Key turning points in the evolution of hypsodonty, cursoriality, and monodactyly were at times of warm and wet climate (figs. 3, 5). These warm-wet crucibles of grassland coevolutionary innovation may have a biological rather than climatic explanation.

One explanation is that grassland evolution was favored by megaherbivores, such as Chadronian titanotheres and Barstovian gomphotheres. Similar to modern African elephants, these enormous mammals probably consumed great quantities of browse, tore down trees, and trampled undergrowth (Owen-Smith 1988). Grasses are colonizing plants with small seeds, basal tillering, underground rhizomes, and wind pollination adapted to such dis-

turbance (Retallack 2001a). Titanotheres, similar to large uinatheres before them, had been around for some time before the 34–35-Ma appearance of grassland and were extinct by 32 Ma (Janis 1997; Prothero 2004). Proboscids arrived too late, at 16 Ma, (Woodburne 2004) to explain the origin of sod grasslands at 19–17 Ma. Furthermore, large rhinos persisted throughout this time interval (Prothero 2005).

Another explanation is that grasslands were favored by innovations in tooth design, such as fully lophed teeth (lophodonty) and high-crowned teeth (hypsodonty). These are tooth designs adapted to more efficiently triturating leaves (lophodonty) and withstanding the abrasiveness of dusty and phytolith-rich grass leaves (hypsodonty). The rising proportions of mammalian genera with lophodont teeth during the Uintan (40 Ma), Duchesnean (39 Ma), Chadronian (37 Ma), and Orellan (32 Ma; Janis 1997) indicates an evolutionary process more protracted than the origin of bunch grasslands at 34–35 Ma. The Hemingfordian (17 Ma) rise of hypsodonty in equids (fig. 11F) postdates, and was probably a consequence of, sod grasslands dating back to 19 Ma (fig. 10).

Other biological explanations are that grasslands were favored by changes in behavior, such as Chadronian (34–35 Ma) origin of rumination (cud chewing) proposed using molecular clock techniques on digestive ribonucleases of extant ruminants and their relatives (Jermann et al. 1995). The fivefold increase in catalytic activity and its acquisition by a marked evolutionary radiation of ruminants at about 35 Ma can be taken as an adaptation to digestion of grasses, which are richer in cellulose, opal, and indigestible substances than other leaves (Benner et al. 2002). Early horses (*Mesohippus*) show lophodonty, wear patterns, and stable isotopic composition compatible with facultative C₃ grazing and browsing (Solounias and Semprebon 2002). Such herbivore pressure would elicit coevolutionary adaptations in grasses, such as basal tillering, telescoped meristems, underground rhizomes, and high opal content in the classical model of grass-grazer coevolution first proposed by Kowalevsky (1873).

A later behavioral innovation plausibly responsible for the origin of sod grasslands was the Hemingfordian (20–16 Ma) origin of pack-hunting carnivores and its effect in aggregating large herds of herbivores. The prorean gyrus of carnivore brains, visible on skull interiors, is found in modern pack-hunting carnivores such as wolves but not solitary hunters such as foxes (Radinsky 1969, 1973). The

prorean gyros first appears in fossil dogs (*Osbornodon*) and cats (*Pseudaelurus*) as old as the Hemingfordian (Wang 1994; Rothwell 2001; Van Valkenburgh et al. 2003). Although Oligocene oreodonts are so common that they have been considered herd animals, no more than 10 skeletons have been found in a family group, which included a litter of pups, and they may have lived in underground dens (Sundell 2004). The oldest fossilized mammal herd is the 17 individual camels in the *Stenomylus* quarry at Agate, Nebraska (Peterson 1911). Also at Agate are the oldest known dung cakes (Retallack 1990); only pelletlike coprolites are known before that time (Retallack 2004b). Sod-forming grasses are generally richer in protein than bunch grasses (Ferdinandez and Coulman 2001), and high-protein graze increases moisture content of ruminant dung (Ireland-Perry and Stallard 1993). The oldest sod-forming grassland paleosols (Mollisols) appear during the early Miocene (19 Ma) at Agate, Nebraska (site 65, fig. 1; table A1), Maiden Creek, Montana (site 42), and Bone Creek, Oregon (site 7).

Clues to evolution of sod grassland ecosystems come from the "Savory solution" for pasture improvement by high-density grazing, as opposed to livestock roaming within large fenced enclosures that go to weed because only the most palatable plants are eaten (Savory and Butterfield 1999). In nature, sod grasslands are promoted by dense herds of large ungulates, which cluster for safety from predators and thus graze both palatable and unpalatable grasses. Cakelike dung left by the herd blankets the area, forcing a respite from grazing until the dung decays after a couple of weeks, and mainly sod-forming grasses resprout through this cover. By this view, the origin of sod grasslands is a biological process of coevolution in which basal tillering, protected meristems, and phytoliths of grasses coadapt to hypsodonty, hooves, and dung cakes of ungulates (table 1).

Comparison with Marine Oxygen Isotopic Record

Paleoclimatic records presented here from fossil soils, plants, and mammals do not agree with the oxygen isotopic record from benthic marine foraminifera, which shows different excursions and trends (Miller et al. 1987; Zachos et al. 2001a). Notable differences are rapid shifts during the early and late Oligocene and the sloping rather than flat overall trend in oxygen isotope values, not seen in the paleosol record of Oregon (asterisks in fig. 12A). As is well known (Zachos et al. 2001a), marine ox-

xygen isotopic values are compromised as proxies of paleotemperature by changes in (1) ocean currents, (2) terrestrial ice volume, (3) oceanic salinity, (4) crustal recycling, and (5) burial diagenesis.

Marine paleoclimates could have differed from terrestrial paleoclimates due to local redirection of ocean currents, such as opening of the Southern Ocean gateway south of Tasmania (Kennett and Exon 2004). Oxygen isotopic records from the tropical Pacific show subdued early and late Oligocene excursions (Pekar et al. 2002) compared with global compilation for the Oligocene dominated by Southern Ocean cores (Zachos et al. 2001a, 2001b), but an early Oligocene peak and late Oligocene trough in $\delta^{18}\text{O}$ values are global (Miller et al. 1987; Pekar et al. 2002). While absolute values vary slightly from place to place, marine oxygen isotopic trends appear to be global (Zachos et al. 2001a).

Abrupt early and late Oligocene shifts in oxygen isotope composition of marine foraminifera have long been attributed to thresholds in growth (at 33 Ma) and melting (at 25 Ma) of the Antarctic ice sheet, which is a large reservoir of isotopically light oxygen disconnected from the world ocean. Although ice sheet growth and decay are related to temperature, the relationship is nonlinear, with complex feedbacks (Zachos et al. 2001a). The effects of ice sheets on oxygen isotopic estimates of paleotemperature can be assessed by independent estimates of paleotemperature from Mg/Ca ratios of foraminifera (Lear et al. 2000). Such corrections indicate less pronounced early Oligocene cooling and late Oligocene warming than oxygen isotope data indicate, which is more in line with paleotemperature from paleosols presented here (fig. 3). Nevertheless, the Mg/Ca paleothermometer is compromised by variations in atmospheric carbon dioxide (Demichco et al. 2003), oceanic salinity (Lemarchand et al. 2000), and burial diagenesis (de Villiers 2003).

Salinity effects can also affect oxygen isotopic records of foraminifera, but this effect would have been unimportant for benthic foraminifera used in recent compilations (Zachos et al. 2001a) compared with planktic foraminifera. The boron isotopic proxy for past oceanic pH and inferred atmospheric carbon dioxide levels (Pearson and Palmer 2000) may reflect varying alkaline riverine runoff and oceanic salinity (Lemarchand et al. 2000). Changes in oceanic pH, salinity, and riverine runoff are also suggested by indications of varying atmospheric CO_2 from the stomatal index of fossil leaves of *Ginkgo* (Retallack 2002; Wynn 2003) and Lauraceae (Kürschner et al. 2002).

An additional distinctive feature of the marine

Table 1. Key Adaptations in the Coevolution of Grassland Ecosystems

Ecosystem type	Modern examples	Age of first appearance	Plant adaptations	Invertebrate adaptations	Mammal adaptations
Agro-ecosystems	Swiss Alps	Holocene (10 ka)	Large seeds and broad leaves; increased yield	Insecticide resistance; decreased yield	Domestication; ease of handling
Tall sod grassland	Tall grass prairie of Illinois	Late Miocene (7 Ma)	C ₄ photosynthetic pathway; CO ₂ conservation	Large and diverse dung beetles; condition soil	Large, hard hooves; increase grazing pressure
Short sod grassland	Short grass prairie of Colorado	Early Miocene (19 Ma)	Telescoped internodes; forming sod	Deep-burrowing earthworm; condition soil	Pack hunting and herds; increase grazing pressure
Bunch grassland	Mitchell grass plains of Australia	Late Eocene (35 Ma)	Hay; extensive silicification of grasses	Burrowing dung beetles; condition soil	Ruminant digestion; increase grazing pressure
Arid shrubland	Bluebush and mallee shrubland of Australia	Eocene and earlier (Carboniferous?)	Small, revolute and pilose leaves; reduced water loss of shrubs	Cicada burrows; aerate soil	Burrowing rodents; aerate soil

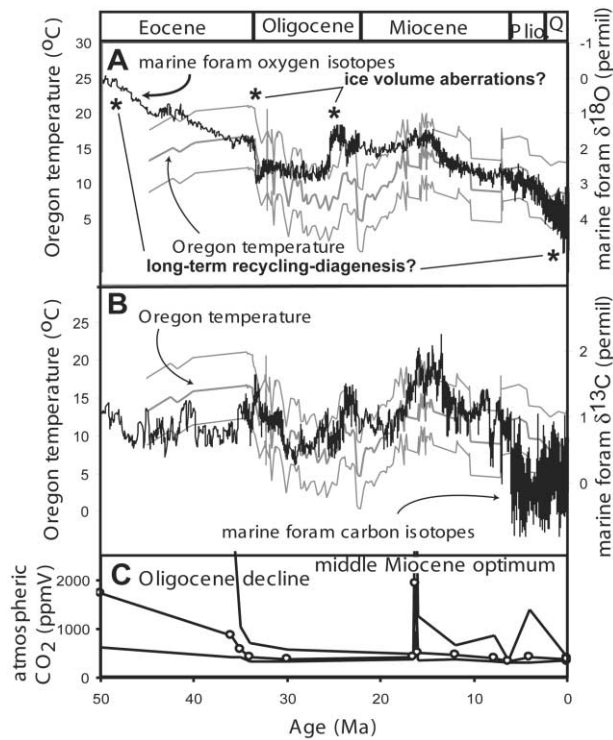


Figure 12. Time series of mean annual paleotemperature of Oregon (A and B in gray, from fig. 3A) compared with oxygen (A) and carbon (B) isotopic composition of deep-sea foraminifera (Zachos et al. 2001a), and with atmospheric CO_2 (C) estimated from stomatal index of *Ginkgo* (Retallack 2002). Error envelopes are all 1 SE.

oxygen isotopic record is its long-term trend to more negative values (lower $\delta^{18}\text{O}$, but beware conventional inverse plotting in fig. 12A), unlike relatively flat marine carbon isotope (fig. 12B) and paleosol records (figs. 3, 5). Increasingly negative oxygen isotope values back through time indicate progressively warmer temperatures to unrealistically high ocean temperatures (50°C) for the earliest Cambrian (542 Ma), when biogenic carbonate averaged -7.5‰ $\delta^{18}\text{O}$ (Veizer et al. 2000). Some of this long-term trend may be due to deep burial diagenesis, which affects the older part of the record more than the younger part (Mii et al. 1997). Also likely is crustal recycling of water by tectonic subduction, with preferential crustal sequestration of ^{18}O (Veizer et al. 2000).

The oxygen isotopic record of deep-sea foraminifera can also be compromised by burial diagenesis, which is more likely to affect oxygen than carbon isotopic composition (Mii et al. 1997). Partial dissolution of foraminiferal tests is also a serious problem for accurate oxygen isotopic com-

positions during the Cenozoic (de Villiers 2003). Even large shells and pedogenic nodules are not immune to diagenetic alteration of oxygen isotope compositions (Lander and Hay 1993; Mii et al. 1997).

Comparison with Marine Carbon Isotopic Record

Continental temperature fluctuations inferred from fossil soils, plants, and mammals appear to match late Miocene to Pleistocene but not Eocene and Oligocene marine carbon isotopic records (fig. 12B) from the same foraminifera. Peak atmospheric carbon dioxide levels and warmth at 35 and 16 Ma were times of foraminiferal carbonate carbon isotopic lows, but curiously, there also are dramatic drops in foraminiferal carbon isotopic values with late Miocene and Pleistocene cooling (fig. 12C). Isotopic composition of pedogenic carbonate on land (fig. 10D) also shows low values at 35 and 16 Ma but swings to very high values with late Miocene cooling and rise of C_4 grasses (Fox and Koch 2003). Long-term Cenozoic (Zachos et al. 2001a, 2001b), Mesozoic (Jenkyns 2003), and Paleozoic isotopic records (Saltzman 2005) reveal that oceanic carbonate has high carbon isotopic values during times of carbon burial, cooling, and falling carbon dioxide but low isotopic values during times of carbon oxidation, warming, and rising carbon dioxide. Exceptions to this rule during the late Miocene to Pleistocene may reflect the rise of C_4 grasses, an impressive demonstration of the scope of this global geochemical shift (Cerling et al. 1997).

Soils and their ecosystems participated in these greenhouse mechanisms. Paleoclimatic warm spikes were associated with deeper calcic horizons and chemical weathering in rangelands (figs. 3, 5) and also with the spread to high paleolatitudes of subtropical to tropical soils (Oxisols, Ultisols, and associated laterites and bauxites; Schwarz 1997). These deeply weathered and highly oxidized paleosols are lean in nutrients (C and P) compared with tundra and grassland soils (Gelisols and Mollisols, respectively) of temperate regions (Retallack 2004c). Extension of Oxisol-Ultisol biomes at the expense of Gelisol-Mollisol biomes would have contributed to atmospheric CO_2 greenhouse spikes with dwindling mineral nutrient supply to the ocean. Conversely, grass-grazer coevolution promoting spread of tundra and grassland ecosystems rich in nutrients (C and P) would have been a biological force for cooling (Retallack 2001a). With lowering atmospheric CO_2 levels, the spread of grasses with the CO_2 -conserving C_4 photosynthetic pathway also reset global carbon isotopic mass bal-

ances (Cerling et al. 1997). Rate and depth of chemical weathering, carbonate precipitation, and carbon inventories of soils have long been important to the global carbon cycle and climate system. Global warmings of the past are now evident on land as well as at sea.

Conclusions

Paleosols are an underexploited and valuable archive of paleoclimatic information for terrestrial environments. Some of these paleosol archives in thick synorogenic sequences (fig. 2) have temporal resolution that includes Milankovitch-scale (100–41 kyr) climatic cycles (fig. 9). Tectonic uplift and volcanic ash eruptions enhance the temporal resolution of paleosol records of global paleoclimatic fluctuations rather than overriding paleoclimatic signals (fig. 8). Paleosol sequences are well exposed and easily accessible (fig. 6). Furthermore, paleosols are colorful and varied, with many paleoclimatically significant features, such as depth to carbonate nodules, visible in the field. Independent chemical and field proxies for mean annual paleotemperature, mean annual precipitation, and seasonality of precipitation (figs. 3, 5) give internal checks for paleosol records of paleoclimate.

Paleosol records of paleoclimate match those of

fossil plants and animals in the same deposits, though fossil sites are not as common as paleosols and so do not present such temporally complete records. In contrast, paleosol records of paleotemperature do not match high-resolution global compilations of oxygen isotopic record of paleotemperature from deep-sea cores, which are regionally biased toward the Southern Ocean and compromised by changing continental ice volumes, marine salinity, long-term crustal recycling, and burial diagenetic alteration (fig. 12A). Paleosol records of paleotemperature do match the carbon isotope record from marine foraminifera (fig. 12B) and the stomatal index record of atmospheric CO₂ (fig. 12C). These and other observations from paleosols that warm climates were also wet climates support the greenhouse hypothesis that atmospheric CO₂ abundance was a major factor controlling Cenozoic paleoclimate.

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