Global Cooling by Grassland Soils of the Geological Past and Near Future

Gregory J. Retallack

Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403;
email: gregr@uoregon.edu

Keywords

grass, mammal, coevolution, paleosol, paleoclimate, carbon sequestration, albedo

Abstract

Major innovations in the evolution of vegetation such as the Devonian origin of forests created new weathering regimes and soils (Alfisols, Histosols) that increased carbon consumption and sequestration and ushered in the Permian-Carboniferous Ice Age. Similarly, global expansion of grasslands and their newly evolved, carbon-rich soils (Mollisols) over the past 40 million years may have induced global cooling and ushered in Pleistocene glaciation. Grassland evolution has been considered a consequence of mountain uplift and tectonic reorganization of ocean currents, but it can also be viewed as a biological force for global change through coevolution of grasses and grazers. Organisms in such coevolutionary trajectories adapt to each other rather than to their environment, and so can be forces for global change. Some past farming practices have aided greenhouse gas release. However, modern grassland agroecosystems are a potential carbon sink already under intensive human management, and carbon farming techniques may be useful in curbing anthropogenic global warming.

INTRODUCTION

Alfisol: order of forests soils, defined by significant subsurface accumulation of base-rich clays, such as smectite and illite

Histosol: order of swamp soils, defined by surface accumulations of significant thickness of peat

Grazer: an animal that eats mainly grasses

Browser: an animal that eats mainly leaves of plants other than grass

Crumb ped: an ellipsoidal small (4–5 mm) clod of soil, characteristic of Mollisols, and formed as earthworm excrements and in the interstices of fine roots

Soils play a critical role in the carbon cycle by fueling photosynthesis, either directly through roots or indirectly through supply of nutrients in runoff to aquatic communities. Plant creation of reduced organic compounds and export of bicarbonate and nutrient cations in solution are forces for global cooling, because soils, lakes, and oceans store carbon that may otherwise be liberated to the atmosphere as the greenhouse gases CO_2 and CH_4 (Berner & Berner 1996, Schmidt et al. 2011). Some 444 Mya (Late Ordovician), the evolution of early land plants increased depth and intensity of chemical weathering in well-drained soils, and glaciers spread across the current Sahara Desert, then at high latitudes, during the Hirnantian Ice Age (Retallack 2000, Lenton et al. 2012). Again, 390 Mya (Early Devonian), the evolution of trees with their large roots increased the depth and intensity of chemical weathering in well-drained forest soils (Alfisols) (Retallack 1997, Retallack & Huang 2011) and also increased the thickness of wetland peats (Histosols) (Retallack et al. 1996). Atmospheric CO_2 levels then fell (Berner 1997) as glaciers spread across polar regions of the Gondwana supercontinent (Isaacson et al. 2008). These evolutionary leaps in chemical weathering and plant carbon sequestration were undone by evolution of Siluro-Devonian millipedes and insects, followed by Mesozoic termites and sauropod dinosaurs (Retallack 2004). Similarly, coevolution of grassland ecosystems over the past 40 million years may also have been a force for global cooling, recently undone by human exploitation of soils and fossil fuels. This review outlines a role for natural grasslands in climate change (Retallack 2001, 2007b) and the implications for management of agricultural grasslands over the next few decades of expected global warming due to human burning of fossil fuels (Suttle et al. 2005, Sanderman et al. 2010).
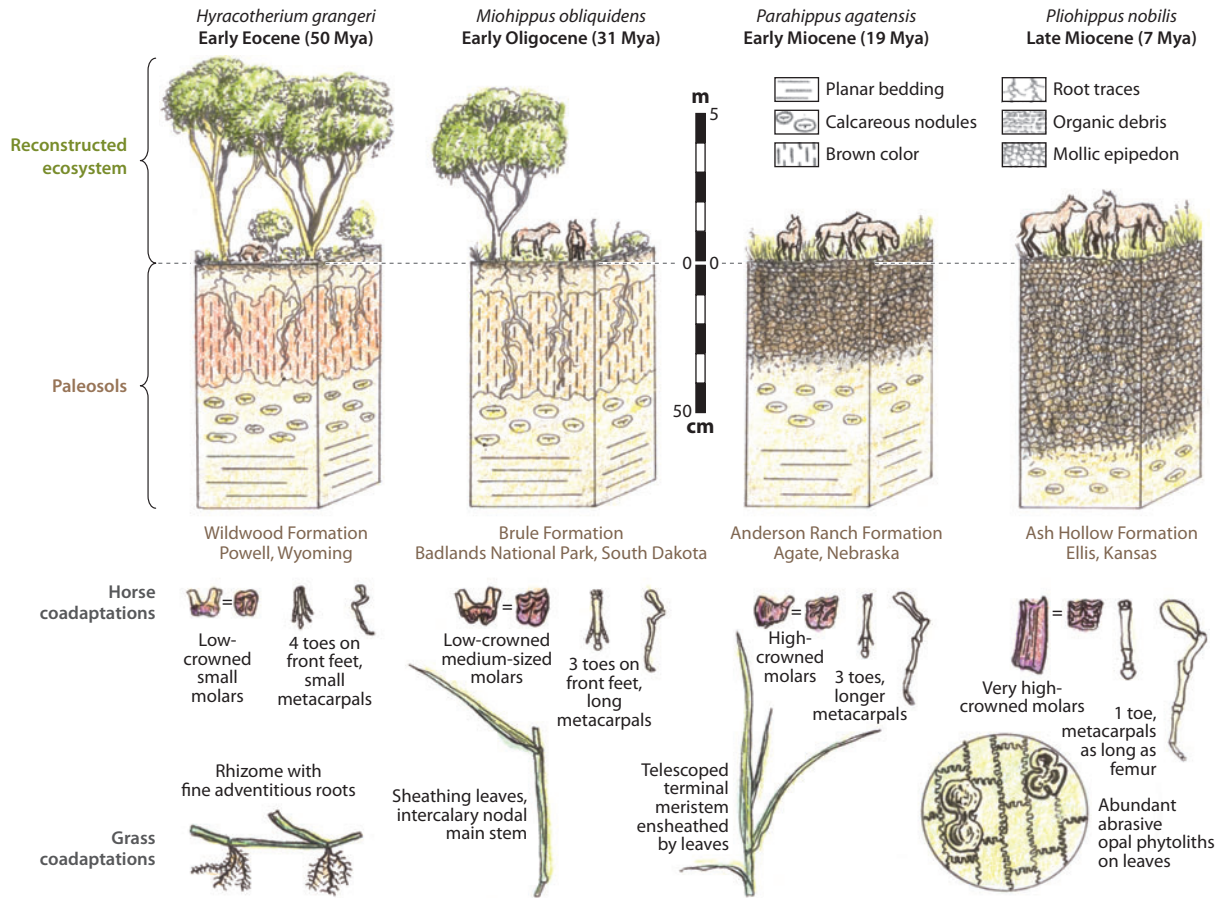
GRASSLAND COEVOLUTION

Coevolution is the coordinated evolution of two different kinds of organisms that are mutually interdependent, such as grasses and grazers (Retallack 2007b). Over the past 40 million years, increasingly high-crowned then continuously growing teeth of ungulates have adapted to abrasive grasses and open dusty plains (Mihlbachler et al. 2011). Slender limbs with hard hooves have evolved for running escape on grassy plains (**Figure 1**). Grasses, in their turn, evolved rhizomes, underground sod, telescoped internodes, intercalary meristems, and abundant opal phytoliths to withstand more effectively than did other plants the onslaught of hard hooves and molars (Edwards et al. 2010). The fossil record of grasses extends back to the Cretaceous (Prasad et al. 2005). Grasses were locally abundant during the early Tertiary (Daghlian 1981), but modern grass clades diversified during the Miocene (Strömberg 2005, Edwards et al. 2010). This matches the schedule of tooth evolution evident from fossil mammals, suggesting mixed grazer-browsers by the latest Eocene and grazers by the middle Miocene (Mihlbachler et al. 2011).

The evolution of grasslands has been linked to thinning forest cover due to cooling and drying climates of the Oligocene and Miocene (Edwards et al. 2010, Mihlbachler et al. 2011), but evidence from paleosols falsifies the idea that drying and cooling climates were causes of grassland expansion. Paleosols are evidence of grassland vegetation from crumb ped structure and fine root traces (**Figure 2**) and also of paleoclimate from depth to pedogenic carbonate and geochemical proxies (Retallack 2007a). The drier the climate, the shallower the depth to pedogenic carbonate, so that depths known for modern soils are a proxy for mean annual precipitation from paleosols once adjusted for burial compaction (Retallack 2005). Such paleosol data (Retallack 1998, 2001, 2007a,c) are evidence that bunch and then sod grasslands appeared during warm-wet climate spikes of the latest Eocene and early Miocene, respectively, and subsequently displaced older vegetation types in semiarid to subhumid climatic regimes (**Figure 3**). Furthermore, bunch and
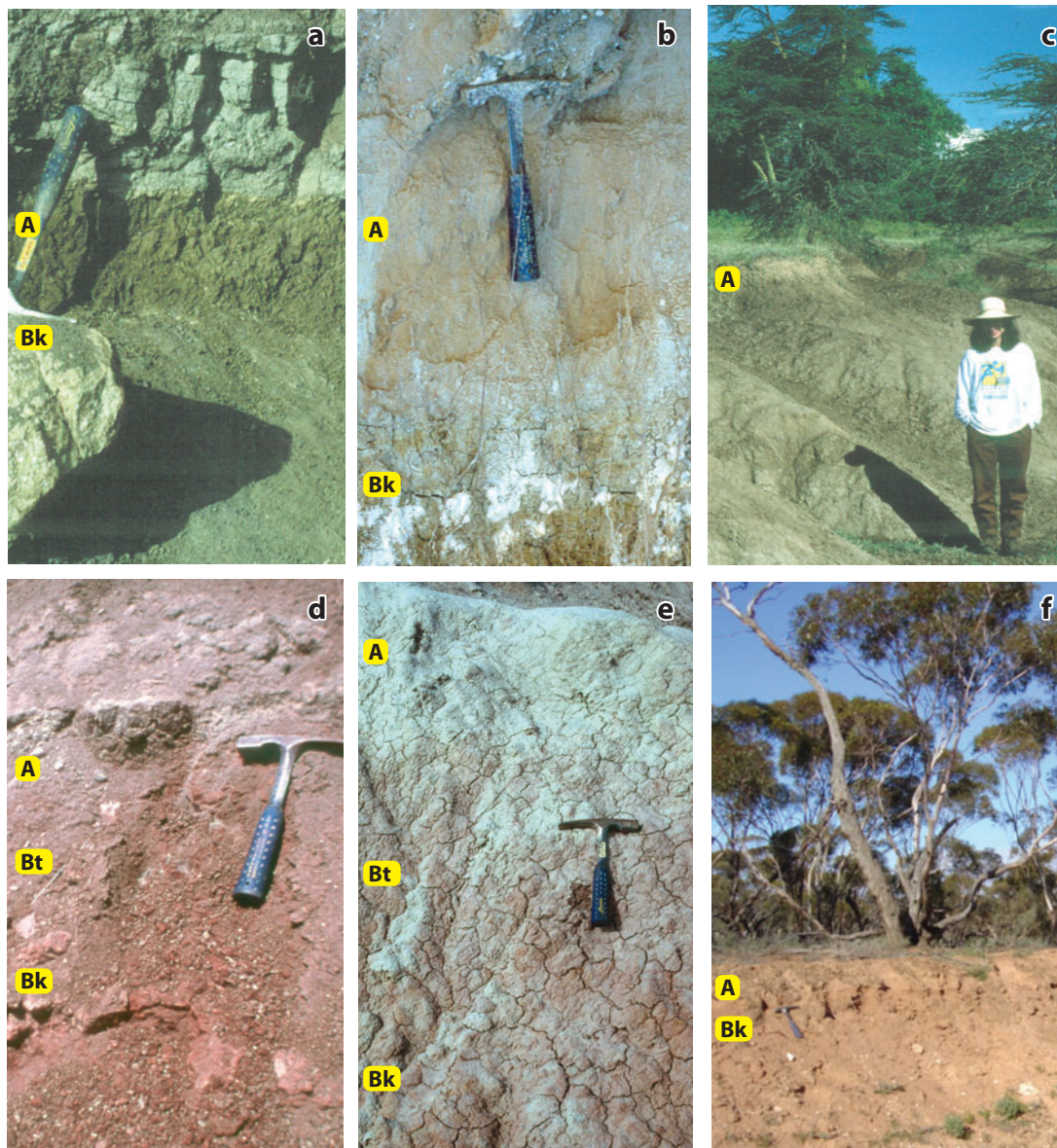
Figure 1

Coevolution of grassland grazers, grasses, and soils (from Retallack 2007b, with permission from Elsevier).

sod grasslands persisted but declined in abundance with subsequent climatic drying and cooling, which encouraged expansion of sagebrush paleosols (**Figure 4**), trace fossils (Sheldon & Hamer 2010), blocky polyhedral phytoliths (Blinnikov et al. 2002), and pollen (Davis & Ellis 2010). These observations are evidence that grasslands were an evolutionary novelty displacing earlier kinds of arid shrubland, then expanding their geographic and climatic range during warm-wet greenhouse transients of the late Eocene, middle Miocene, and late Miocene. Grasslands were a biological force expanding their climatic range and not merely filling in arid regions. Grasslands thus differed from desert shrublands that expanded their range during times of dry climate.

In North America, the first appearance of bunch grasslands, judging from the abundant grass phytoliths, fine root traces, and granular ped structure, coincides with the late Eocene (40 Mya) immigration of the White River Chronofauna (Duchesnean) from Asia (Lucas et al. 2004). The

Figure 2

Grassland Mollisols (*a–c*) as well as Alfisols and Aridisols (*d–f*) of the semiarid woodlands they replaced: (*a*) Onuria paleosol in the middle Miocene (14 Mya) Kericho Phonolite at the Fort Ternan mammal quarry, Kenya (Retallack 1991); (*b*) Psito paleosol in the late Miocene (7 Mya) Ash Hollow Formation north of Ellis, Kansas (Retallack 2001); (*c*) modern Mollisol and wooded grasslands near Lake Nakuru, Kenya (Retallack 2007c); (*d*) Tut paleosol in the early Miocene (20 Mya) Kapurtay Agglomerate near Songhor, Kenya (Retallack 1991); (*e*) Luluta paleosol in the early Eocene (54 Mya) Willwood Formation northeast of Worland, Wyoming (Retallack 1998); (*f*) modern Aridisol and dry woodland (mallee) near Damara Station and Mungo Lakes, New South Wales, Australia. Hammer handles in panels *a*, *b*, *d–f* are 25 cm long. Soil horizon labels describe surface horizons (A) and subsurface horizons enriched in carbonate (Bk) and clay (Bt).
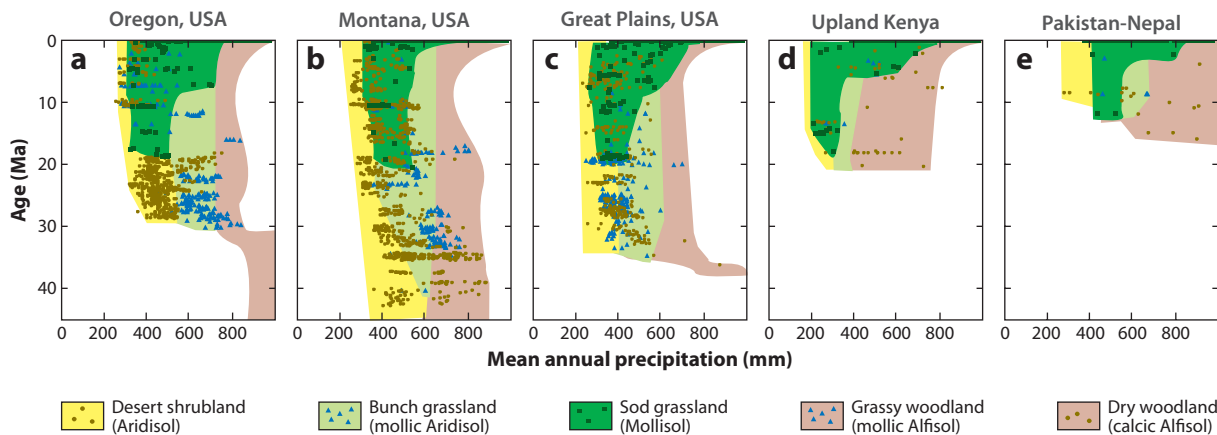
Figure 3

Climatic expansion of bunch grasslands and sod grasslands inferred from different kinds of calcareous paleosols on three continents (data from Retallack 1991, 2001, 2007a,c).

appearance of Mollisols with fine root traces and crumb peds during the early Miocene (19 Mya) in Oregon, Montana, and Nebraska coincides with parahippine horse evolutionary radiation, which spread across the continent from the earliest representatives known in Florida (Mihlbachler et al. 2011). As demonstrated by the prorean gyrus in the brain casts of 19-Ma-old borophagine dogs from Nebraska, these Mollisols appeared at the same time as the earliest known dogs with brains that are comparable with those of modern pack-hunting dogs (van Valkenburgh et al. 2003). A herd of 18 camels (*Stenomylus hitchcocki*) of comparable age (19 Ma) from Nebraska is the oldest known herd of ungulates in North America. Although Oligocene oreodons such as *Merycoidodon culbertsoni* are common and have been considered herd animals, no more than five individuals have been found in any one group of oreodon skeletons (Sundell 2004). These same Nebraska deposits dated to 19 Mya also yield the earliest fossil dung cakes, as opposed to dung pellets common in older rocks (Retallack 1990). Hard hooves focused into herds by pack hunting and fertilization with liquid manure may have been important biological selection pressures in the evolution of sod grasslands and their distinctively crumb-structured soils (Mollisols) (**Figure 2a–c**).

In the Old World, the middle Miocene (16 Mya) invasion of hypsodont bovids from central Asia introduced sod grasslands into Kenya (Retallack 2007c). The late Miocene (8 Mya) expansion of tall grassland ecosystems into humid regions of North America, Africa, and Asia (Retallack 1991) coincides with a large shift in carbon isotopic composition of pedogenic nodules and mammal teeth and bones, indicating the rise of the C_4 photosynthetic pathway (Cerling et al. 1997, Fox et al. 2012). The late Miocene expansion of C_4 grasses and the decline of associated trees (Strömberg & McInerney 2011) were accompanied by geographic expansion into wetter regions (**Figure 3**), so they were not responses to climatic drying or cooling but instead were due to hypergrazer and megaherbivore pressures (Retallack 2007b). These C_4 grasslands replaced preexisting C_3 sod grasslands that had evolved at least 19 Mya, as indicated by morphology and stable isotopic analyses of both paleosols (Retallack et al. 2004) and phytoliths (McInerney et al. 2011). Also at this time, large monodactyl horses appeared and migrated to most parts of the world, with the exception of Australia and Antarctica (Mihlbachler et al. 2011). The expansion of grasslands into more humid regions coincides with evolutionary innovations enabling them to displace earlier grassy woodlands, not with times of dry and cool paleoclimates (Retallack 2007a).

Mollisol: order of grassland soils defined by significant thickness (>18 cm) of organic, clayey, and crumb-structured surface horizons

Prorean gyrus: a prow-like fold in both hemispheres of the dog forebrain, behind the olfactory bulb, and characteristic of pack-hunting dogs

Hypsodont: high crowned; a form of mammal tooth protruding high above the gum line

C_4: a photosynthetic pathway using the Hatch-Slack cycle, bundle sheath cells, and the enzyme phosphoenolpyruvate carboxylase to produce ^{13}C-enriched four-carbon sugars
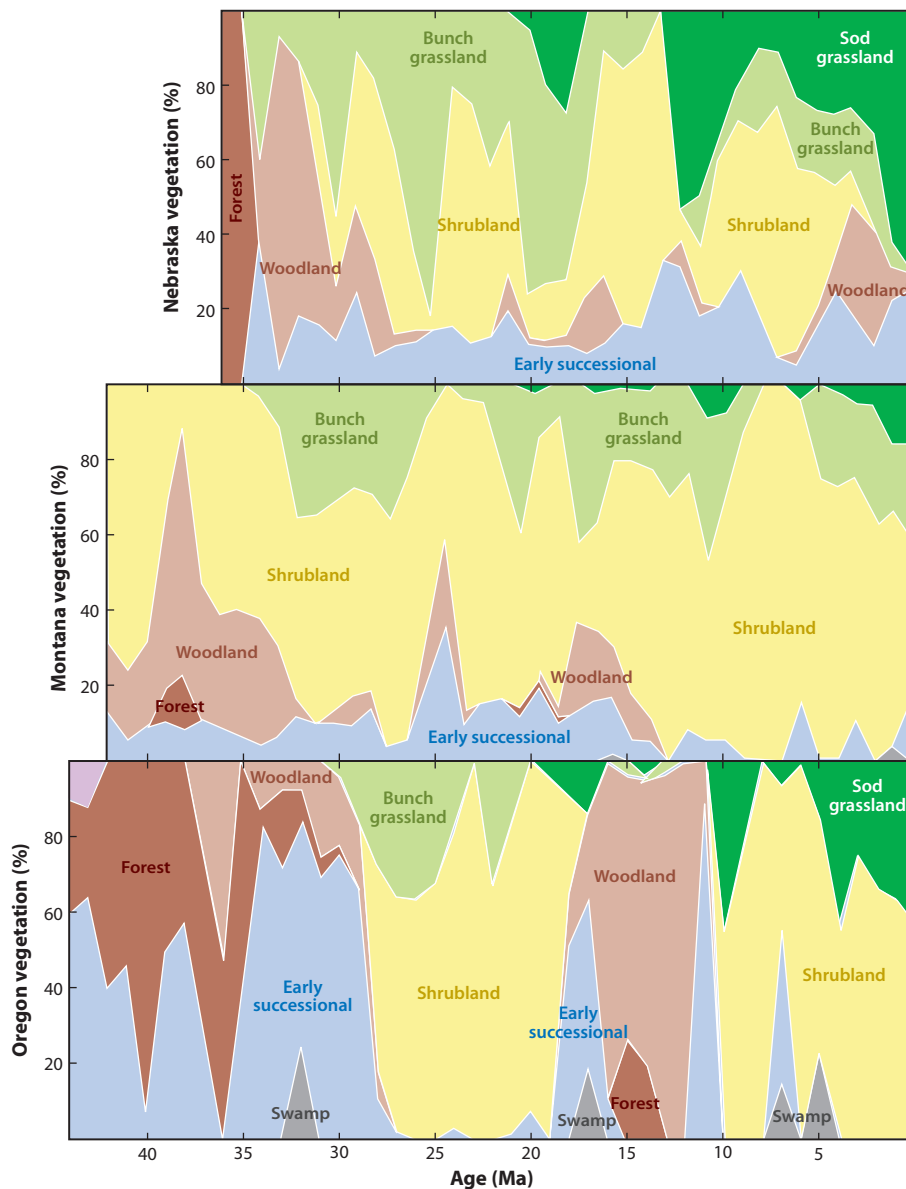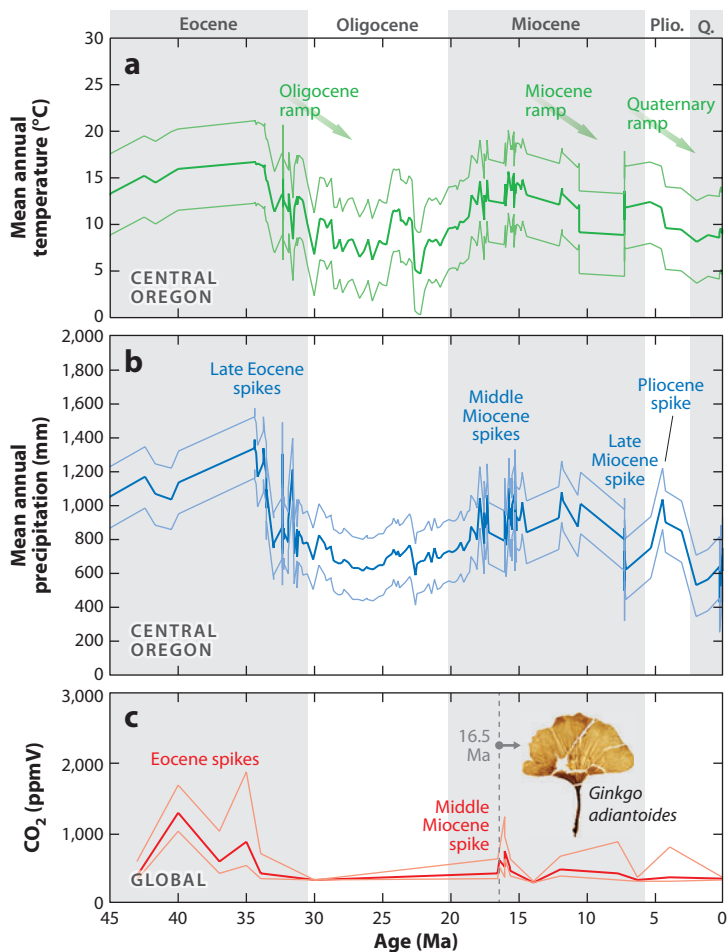
Figure 4

Paleovegetation inferred from relative abundance of paleosols in three American states shows appearance of bunch and sod grasslands at times of warm-wet paleoclimate and expansion of shrubland (now mainly sagebrush) at times of dry climate (data from Retallack 2007a).

CENOZOIC CLIMATE COOLING

The general trajectory of Cenozoic climate is known from chemical paleoclimatic proxies for paleosols in several regions of North America (Retallack 2007a): Eocene (45–35 Mya) spikes of high temperature and precipitation are followed by a ramp of Oligocene (34–19 Mya) cooling and drying, then middle Miocene (19–16 Mya) and late Miocene (7–8 Mya) transient spikes, separating

C₃: a photosynthetic pathway using the Calvin cycle and ribulose 1,5 bisphosphate carboxylase oxygenase (rubisco) to produce ^{13}C-depleted three-carbon sugars
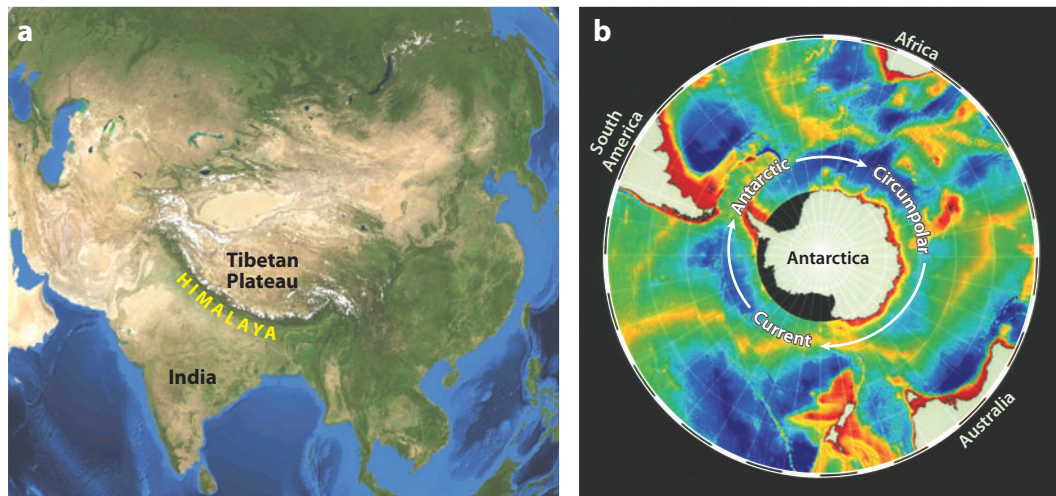
Figure 5

Time series of (*a*) paleotemperature and (*b*) paleoprecipitation from the chemical composition of paleosols in Oregon (Retallack 2007a) and atmospheric CO_2 from the fossil *Ginkgo* stomatal index (Retallack 2009a) both suggest a role for CO_2-greenhouse control of Cenozoic paleoclimate. The fossil leaf (*c*) is middle Miocene (16.5 Mya) *Ginkgo adiantoides* from below the Grande Ronde Basalt, near Wieppe, Idaho (described by Retallack & Rember 2011). Abbreviations: Plio., Pliocene; Q., Quaternary.

ramps into the Pleistocene Ice Age (**Figure 5a,b**). The pedogenic record of paleoclimate from Oregon is more like isotopic records from foraminifera in the South Atlantic (Zachos et al. 2001) than continental paleoclimatic records from other parts of the world (Retallack 2001, 2007c), but there are significant differences. Paleosol records are more like marine carbon isotopic records, which reflect paleoproductivity, than marine oxygen isotope records, which reflect paleotemperature, but are compromised by episodes of continental ice growth (Zachos et al. 2001). Some of these effects were global, because the Eocene (37–38 Mya) glacial expansion and retreat in Greenland (Eldrett et al. 2007) were followed by the Oligocene (32 Mya) expansion of Antarctic glaciers and then the late Miocene (6 Mya) expansion of both polar ice caps (Zachos et al. 2001).

Greenhouse control of Cenozoic paleoclimate is apparent from multiple proxies for atmospheric CO_2 (Beerling & Royer 2011, van de Wal et al. 2011), including pedogenic carbonate

Figure 6

Proposed causes of Cenozoic cooling: (*a*) the Tibetan Plateau (Raymo & Ruddiman 1992) and (*b*) the Antarctic Circumpolar Current (Kennett 1977). Asia topography is from NASA (Blue Marble Next Generation). Southern Ocean seafloor topography is from Smith & Sandwell (2012) and is also public domain (NOAA).

carbon isotopic compositions (Retallack 2009b), foraminiferal boron isotopic analysis (Pearson et al. 2009), and the *Ginkgo* stomatal index (Retallack 2009a, Retallack & Rember 2011). The middle Miocene spike in atmospheric CO_2 is not apparent from the marine alkenone paleobarometer (Pagani 2002), but that proxy may have been compromised by changes in the nutrients in the ocean introduced during intensified middle Miocene weathering on land (Retallack 2009b). Theoretical models predict temperature increases of 1.5 to 6.2°C with the doubling of atmospheric CO_2 (Royer et al. 2007), but the doubling sensitivity from empirical proxy data from the past 300 million years is only 0.8 $\pm$ 1.9°C (Retallack 2009a) and from records of the past 159 years is 1.5 $\pm$ 0.3°C (Gillett et al. 2012). The correspondence of Cenozoic temperature, precipitation, and CO_2 records (**Figure 5a–c**) suggests a role for the carbon cycle in explaining patterns of Cenozoic climate change.

COOLING BY MOUNTAIN UPLIFT?

In their popular hypothesis to explain Cenozoic cooling, Raymo & Ruddiman (1992) postulated that mountain uplift in general and the Himalaya in particular (**Figure 6a**) cooled the planet over the past 35 million years. They considered changes in strontium isotopic ratios in marine carbonates as evidence of enhanced chemical weathering of silicates due to mountain uplift. Hydrolytic weathering consumes carbon as carbonic acid, exporting it as bicarbonate and fueling photosynthetic reduction of carbon, thus cooling the planet by withdrawal and burial of carbon in the ocean. Additional cooling was envisaged from growth of glaciers on the Tibetan Plateau, increasing albedo over a large region and thus exporting solar energy back into space.

One problem with this view is conflation of physical weathering (promoted by mountain uplift) with chemical weathering (needed for CO_2 consumption). Studies of the geochemistry of Himalayan streams (Jacobson et al. 2002) and sediments (Singh 2010) are evidence that 80% of the strontium isotopic signal comes from simple dissolution of Himalayan carbonate, not from CO_2-consuming hydrolytic weathering of silicates. Soils and sediments of the Ganges Plain are

Stomatal index:
percentage of stomates over stomates-plus-epidermal cells, measurable on fossil plant cuticles as a proxy for paleoatmospheric CO_2 levels

Alkenone: ketones formed by addition of a carbonyl group to a straight-chain hydrocarbon (alkane), common in cell walls of marine phytoplankton

Albedo: the ratio of light reflected from a surface over the amount of incident light, expressed as a percentage or fraction

six times more chemically weathered than those of the High Himalayas (West et al. 2002). Thus High Himalayan silicate chemical weathering is much less than that of preexisting lowlands. Nevertheless, chemical weathering did increase globally over the past 70 Ma, as revealed by marine foraminifera with 8–9‰ overall enrichment in δ^7Li, which is a more sensitive indicator of increased chemical weathering than are rising marine ^{87}Sr/^{86}Sr ratios (Misra & Froelich 2012). Surprisingly, constant global physical weathering is also revealed by marine ^{10}Be/^{9}Be ratios over the past 14 Ma (Willenbring & von Blanckenburg 2010).

Even though a change from lowland jungle to snow and ice would have been a significant increase in albedo for the Himalayan peaks and Tibetan Plateau, this is a small region compared with the Saharan and Asiatic deserts that also expanded in area (**Figure 6a**) (Retallack 1991). Not all these deserts are in the rain shadow of the Himalayas, and they indicate even wider regions of reduced weathering and carbon sequestration. This spread of desert regions, and their lower soil CO_2 and thus lower chemical weathering, may have kept atmospheric CO_2 from dipping below 180 ppm. This important atmospheric minimum has been modeled by Pagani et al. (2009) and Beerling et al. (2012), who argue that the spread of grasslands was responsible, but their assumption of less weathering under grassland than under shrubland is false. Soil monitoring of grassland-shrubland pairs in Azerbaijan, in the same temperate arid climate, revealed 4,500 ppm late growing season soil CO_2 under grassland but only 2,200 ppm for desert shrubland (Retallack 2009a). It is soil CO_2, not the much lower atmospheric CO_2, that should be used to model chemical weathering.

Tethyan subduction associated with uplift of the Tibetan Plateau generated enormous amounts of CO_2 during the late Eocene (Rowley & Currie 2006, Kent & Muttoni 2008). Metamorphic decarbonation reactions with tectonic uplift of the Himalayas continue to be a source of greenhouse gases in widespread hot springs: For example, the 32,000-km^2 Narayani Basin of the central Himalayas is degassing more than 1.3×10^{10} mol per year of CO_2, which is four times the consumption of CO_2 by chemical weathering in that basin (Evans et al. 2008). In general, mountain uplift, volcanic eruptions, and extraterrestrial impacts are forces for global warming, because they release carbon to the atmosphere through frictional, intrusive, and impact heating of carbonates and organic matter (Retallack 2009a, Kidder & Worsley 2012). The warm-wet CO_2 spike of the latest Eocene (35 Mya) coincides with Ethiopian plateau basalts (Pik et al. 2008) and Chesapeake impact structure (Gohn et al. 2009), and the warm-wet spike of the middle Miocene (16 Mya) coincides with Columbia River Basalts (Barry et al. 2010, Kidder & Worsley 2012) and Steinheim-Ries impact craters (Mihalyi et al. 2009).

COOLING BY ANTARCTIC CIRCUMPOLAR CURRENT?

According to an alternative hypothesis proposed by Kennett (1977), Cenozoic cooling may have been caused by the completion of the Antarctic Circumpolar Current (Katz et al. 2011) as Australia and South America drifted away from Antarctica as a result of seafloor spreading (**Figure 6b**) (Smith & Sandwell 2012). This had the effect of thermally isolating Antarctica from warm currents that formerly flowed south along the east coasts of Australia and South America. Cold nutrient-rich Antarctic waters became biologically productive, burying carbon in marine clays, and the growth of ice in Antarctica raised planetary albedo and locked away much atmospheric water vapor.

There have been problems with the timing of Antarctic isolation by seafloor spreading in the Southern Ocean. Both Australian separation dated at 35 Mya and South American separation at 41 Mya (Scher & Martin 2006) significantly predate cooling in paleoclimatic records (**Figure 5**). Paleosols of Antarctica are evidence of dramatically lowered chemical weathering and vanquished beech forests over the past 15 million years (Retallack et al. 2001). There is also evidence from
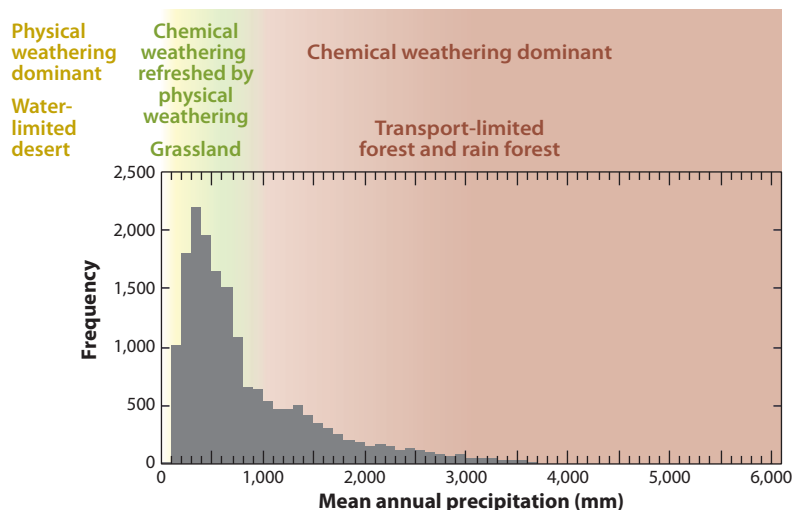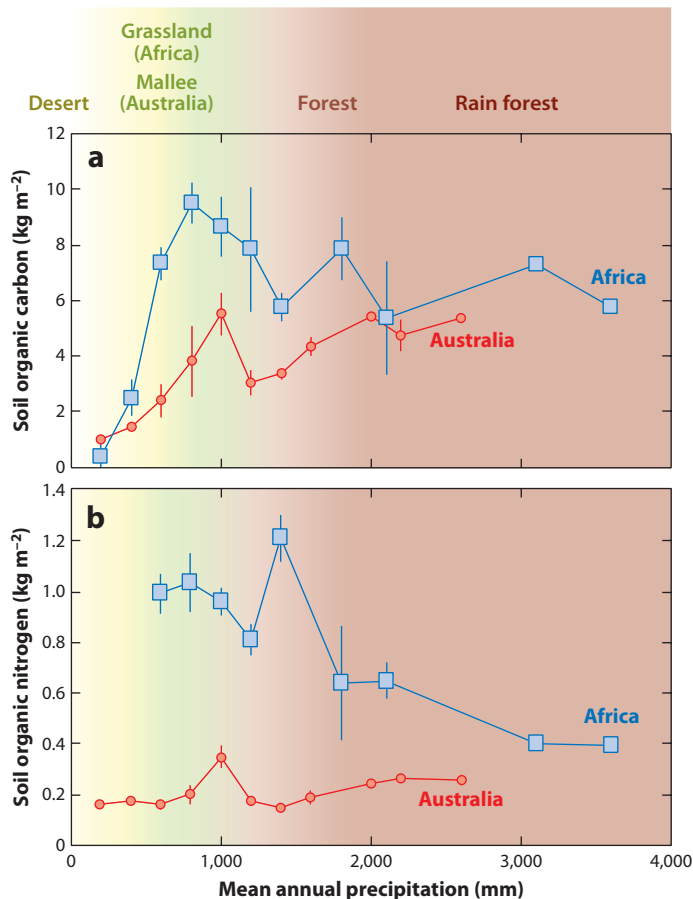
Figure 7

Frequency distribution of mean annual precipitation on land, showing dominance of grassland climates (400–1000-mm mean annual precipitation). These data are from a global network of 17,689 terrestrial precipitation gauges (Grieser & Rudolf 2005).

fossil marine invertebrates for lowered productivity and thus lower carbon sequestration of shallow marine ecosystems over the past 45 million years. Modern Antarctic benthic marine communities with brachiopods and crinoids resemble Paleozoic marine communities more than they do the mollusc- and crab-dominated communities present during the Eocene (Aronson 2009). Computer modeling indicates that the increase of albedo in Antarctica was not sufficient to cool the world without a contribution of declining CO_2 in the atmosphere, to which unvegetated Antarctica did not contribute. By contrast, carbon burial in the Southern Ocean upwelling zones in the models did contribute to carbon sequestration in the ocean, but largely in midlatitudes, which gained a greater proportion of Earth's heat budget (DeConto & Pollard 2003).

COOLING BY GRASSLAND COEVOLUTION?

My explanation (Retallack 2001) for Cenozoic cooling is grassland coevolution (**Figure 1**). Organisms within coevolutionary trajectories are concerned less with their physical environment than with their biological environment; thus, coevolution has the potential to change the physical environment (Retallack 2007b). Grassland ecosystems are well placed as a force for global climate change because they have colonized the most productive soils between the 400- and 1,000-mm isohyet (Suttle et al. 2005, Staver et al. 2011), capturing the mode in mean annual precipitation for all continental regions (**Figure 7**), and mean annual continental precipitation of 762 mm (Grieser & Rudolf 2005). This climatic zone is not only the most widespread, but also the most fertile region of our planet. Soils receiving less than 400 mm are water limited, and soils receiving more than 1,000 mm become increasingly nutrient limited (Retallack 2005). Annual and deciduous angiosperms create higher chemical weathering rates than do perennials and conifers (Volk 1989). Unlike sagebrush and cactus, grasslands in this critical habitat do not appear to have adapted to cooler and drier climates (**Figure 4**). Instead, they seem to have created newly evolved sod (mollic epipedon) and soils (Mollisols) (**Figure 2a–c**) that cooled the atmosphere by means of increased soil carbon sequestration, higher surface albedo, and reduced evapotranspiration (Retallack 2001).

Figure 8

Contrasting climatic gradients of organic (*a*) carbon and (*b*) nitrogen sequestration in Australian soils without native grass-ungulate ecosystems and African soils with native grass-ungulate ecosystems (data from Zinke et al. 1984, Wynn et al. 2005).

Australian perennial sclerophyllous communities lack coevolved native ungulates and sod grasses and have few Mollisols (Milton Moore 1970, Retallack 2012). In Australia, organic carbon and nitrogen found in the soil, as well as net primary productivity and biomass, increase with mean annual precipitation (Zinke et al. 1984, Wynn et al. 2005). The same general trend is true for Africa, but African grassland soils sequester much more organic carbon and nitrogen than do Australian woodlands receiving comparable precipitation (**Figure 8**). Mollisols of tall grassland have as much as 10 wt% organic carbon to depths of more than 1 m, whereas comparable amounts of soil carbon under woodland seldom are deeper than 10 cm (Retallack 2001). The markedly increased carbon storage in soils of tall C_4 grasses, compared with short C_3 grasses, is partly due to their deep roots (Fisher et al. 1994). The organic crumb-textured surface horizon of sod grasslands was, and is, a significant carbon sink as long as grasses persist (Schmidt et al. 2011). Fire is less harmful to grasses with their underground rhizomes, than it is for woody plants, and allows grasslands to spread at the expense of woodlands (Hirota et al. 2011, Staver et al. 2011). Grassland fires leave a residue of grass char, which is a refractory form of organic carbon (Lehmann et al. 2008). Expansion of such organic soils from 0 to 40% of the world's land area

(Suttle et al. 2005) since 19 Mya (**Figure 1**) may explain the long-term downward trajectories in temperature, precipitation, and CO_2 seen in paleoclimatic records (**Figure 5**). If grassland soils sequester approximately 1 kg per m^{-2} C more than do preexisting woodland soils (Retallack 2001), 40% of the current world's land surface of 148,940,000 km^2 could have sequestered an additional 596 Pg C, comparable with 750 Pg C as CO_2 currently found in the atmosphere (Sanderman et al. 2010) and compatible with the observed halving of atmospheric CO_2 over the past 19 million years (**Figure 5**). Estimation of carbon sequestration from marine (Zachos et al. 2001) or pedogenic carbon isotopic analyses (Retallack et al. 2004) is compromised by changing ratios of C_3 and C_4 grasses during the Cenozoic (Cerling et al. 1997, Fox et al. 2012).

Grasslands are light colored (albedo 17–19%), especially when covered by snow (albedo >50%), whereas woodlands are dark (albedo 9–14%) and seldom completely covered with snow (Myhre et al. 2005). In nonsnowy climatic regions, grasses die back and bleach during a dry season, thereby increasing albedo. Thus, Cenozoic grassland expansion (**Figure 3**) has increased planetary losses of solar energy by reflection (Retallack 2001).

Grasslands, and especially sod grasslands, have moist soil but dry air because of low transpiration. In contrast, woodlands and forests create dry soil and moist air. In the data compilation of Alton et al. (2009), temperate C_3 grasslands have transpiration/precipitation ratios of 0.26–0.28, compared with 0.21–0.44 for tropical C_4 grasslands and 0.22–0.33 for shrublands and woodlands. *Eucalyptus*, now planted worldwide, is an infamous soil desiccant (Calder et al. 1997). Woodlands and forests have been locally cooled by evapotranspiration, which creates regionally moist air. Fully coupled ocean-atmosphere global circulation models applied to the Amazon Basin, for example, show significant regional declines in available moisture owing to spread of grassland and shrubland expected with global warming (Cowling et al. 2008). Because water vapor is a powerful greenhouse gas, water would be a greater problem for global warming than is CO_2, were it not so easily rained out, which is not an option for CO_2 (Lacis et al. 2010). Nevertheless, grassland soil sequestration of water vapor may have played a role in desiccating the atmosphere over geological time (Retallack 2001).

FUTURE COOLING EFFECT OF AGRICULTURAL GRASSLANDS?

If the spread of organic-rich grassland soils (Mollisols) were an engine for global cooling over the past 30 million years, could this engine be used to combat coming global warming? Soils are the largest reservoirs of carbon under active management (Sanderman et al. 2010): 1,500 Pg of reduced carbon and 900–1,700 Pg of inorganic carbon in soils globally exchange 60 Pg C per year with an atmosphere of 750 Pg C as CO_2. Christine Jones and Carbon Farmers of Australia have promoted carbon farming to combat global CO_2-greenhouse warming (Kiely 2010). Their aim is to determine the organic carbon content of particular fields as an initial baseline and then compensate farmers for future carbon sequestration achieved by a variety of management practices. Using only a few established carbon-conserving methods, it may be possible for the 50 Mha of intensively managed Australian agricultural land to sequester at least 100 Tg per year, which is roughly 15% of current Australian greenhouse gas emissions (Sanderman et al. 2010). Australia is particularly well suited for experiments in carbon farming, because natural sod grassland communities colonizing the rest of the world never made it past the ocean and forest barriers of Wallace's Line in Indonesia (Milton Moore 1970). Australian vegetation and soils are thus ecologically more like the Eocene dry woodland communities that predated sod grassland evolution on other continents (Metzger & Retallack 2010, Retallack 2012). Storage of both nitrogen and carbon in native Australian soils is notoriously low (**Figure 8**), and sequestration of both elements is greatly increased by agroecosystems that are not native to Australia (Wynn et al. 2005). In other parts of the world,

Figure 9

Cell grazing with cattle penned by an electric fence at Stowe Park Station, Calliope, Queensland, Australia. Photo courtesy of Dan Carney.

carbon farming can restore historically squandered soil carbon (Suttle et al. 2005), but it is unlikely to exceed preagricultural sequestration.

The natural grassland communities of Africa give clues regarding the agricultural practices most effective in building soil carbon. In Australia, the first step in the process was taken in 1788 when sheep and cattle were imported from Eurasia. Subsequent pasture improvements as a result of importing grasses, subterranean clover, earthworms, dung beetles, and livestock created true Mollisols in some agricultural districts of Australia (Milton Moore 1970, Baker et al. 2006).

Other techniques for carbon sequestration of grassland soils include increasing stocking density to match the effect of herds maintained by pack hunters in African grasslands (Savory & Butterfield 1999). This technique is called planned grazing or cell grazing, because small pens defined by electric or other fences limit the cattle to as much feed as needed for a single day or week of grazing. This is followed by rotation to the next cell as the previous cell recovers (**Figure 9**). As a result, weeds as well as grasses are eaten and covered in manure, so that only the most vigorous and palatable grasses regrow from the sod, obviating the need for chemical weed control. If cattle are allowed to spread out over large pastures, they eat only the most palatable grasses, and fields are quickly overrun with noxious weeds. As South African farmers have long known, "you have to hammer the veld to make it sweet" (Savory & Butterfield 1999).

Another carbon-building technique is the successive use of pasture by several different animals to mimic the grazing succession found in the African savannas in which various animals utilize pasture during different seasons of inundation and to differing extents (Vesey-Fitzgerald 1973), thereby stimulating both plant growth and beef production (Odadi et al. 2011). Such techniques of natural sequence farming can also be tough on weeds, reduce loads of parasites specific to particular kinds of livestock, and create multiple farming enterprises on a single site (Kiely 2010).

Cell grazing: planned grazing in which livestock are limited to small pastures and rotated through successive small pastures as grass recovers from grazing

Firing of pastures during the dry season builds refractory grass char in the soil and also limits pathogens (Lehmann et al. 2008). Sowing charcoal into soil can achieve a similar result while improving soil stability and fertility (Kiely 2010). This agricultural technique was used for 7,000 years by Native Americans in the black earth (*terra preta*) fields of Amazonia (Glaser 2007).

Plowing opens soil organic matter to oxidation and erosion. This can be prevented by pasture cropping, in which seed is drilled through the grassy sward reduced to stubble and fertilized by grazing (Kiely 2010). Maintaining a living carpet of sod is critical for maintaining moist, cool soil (Breshears et al. 1997), which preserves soil carbon. Building soil carbon also builds soil fertility, structure, and moisture, and many locally appropriate techniques may be used to achieve that goal (Kiely 2010).

CONCLUSIONS

Humans are not the first creatures to alter Earth's climate. Evolution of termites and dinosaurs also contributed to global warming following a Permian-Carboniferous Ice Age induced by the evolution of trees and forest soils (Alfisols, Histosols) (Retallack 2004). Human agroecosystems and fossil fuel burning are undoing millions of years of grass-grazer coevolution, which created moist organic grassland soils (Mollisols) and sod of high albedo and low evapotranspiration (Retallack 2001). This long-term biological force for global cooling has been interrupted by orogenic, volcanic, and impact crises over the past 45 million years (Retallack 2009a) but culminated in the Pleistocene Ice Age. Future global warming may be mitigated by commercial carbon farming using such techniques as new grass varieties, cell grazing, natural sequence farming, stubble burning, and pasture cropping (Kiely 2010).

DISCLOSURE STATEMENT

ACKNOWLEDGMENTS

LITERATURE CITED

Alton P, Fisher R, Los S, Williams M. 2009. Simulations of global evapotranspiration using semiempirical and mechanistic schemes of plant hydrology. *Glob. Geochem. Cycles* 23:GB4023

Aronson RB. 2009. Metaphor, inference, and prediction in paleoecology: climate change and the Antarctic bottom fauna. *Paleontol. Soc. Pap.* 15:177–94

Baker GH, Brown G, Butt K, Curry JP, Scullion J. 2006. Introduced earthworms in agricultural and reclaimed land: their ecology and influences on soil properties, plant production and other soil biota. *Biol. Invasions* 8:1301–16

Barry TL, Self S, Kelley SP, Reidel S, Hooper P, Widdowson M. 2010. New ^{40}Ar/^{39}Ar dating of the Grande Ronde lavas, Columbia River basalts, USA; implications for duration of flood basalt eruption episodes. *Lithos* 118:213–22

Beerling DJ, Royer DL. 2011. Convergent Cenozoic CO_2 history. *Nat. Geosci.* 4:418–10

Beerling DJ, Taylor LL, Bradshaw CDC, Lunt DJ, Valdes PJ, et al. 2012. Ecosystem CO_2 starvation and terrestrial silicate weathering: mechanisms and global-scale quantification during the late Miocene. *J. Ecol.* 100:31–41

Berner EK, Berner RA. 1996. *Global Environment: Water, Air and Geochemical Cycles.* Upper Saddle River, NJ: Prentice-Hall. 376 pp.

Berner RA. 1997. The rise of plants and their effect on weathering and atmospheric CO_2. *Science* 276:543–46

Blinnikov M, Busacca A, Whitlock C. 2002. Reconstruction of the late Pleistocene grassland of the Columbia Basin, Washington, USA, based on phytolith records in loess. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 177:77–101

Calder IR, Rosier PTW, Prasanna KT, Parameswarappa S. 1997. *Eucalyptus* water use greater than rainfall input: possible explanation from southern India. *Hydrol. Earth Syst. Sci.* 1:249–56

Cerling TE, Harris JM, MacFadden BJ, Leakey MG, Quade J, et al. 1997. Global vegetation change through the Miocene/Pliocene boundary. *Nature* 389:153–58

Cowling SA, Shin Y, Pinto E, Jones CD. 2008. Water recycling by Amazonian vegetation. *Philos. Trans. R. Soc. Lond. Ser. B* 363:1865–71

Daghlian CP. 1981. A review of the fossil record of monocotyledons. *Bot. Rev.* 47:517–55

Davis OK, Ellis B. 2010. Early occurrence of sagebrush steppe, Miocene (12 Ma) on the Snake River Plain. *Rev. Palaeobot. Palynol.* 160:172–80

DeConto RM, Pollard D. 2003. A coupled climate-ice sheet modeling approach to the early Cenozoic history of the Antarctic ice sheet. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 198:39–52

Edwards EJ, Osborne CP, Strömberg CAE, Smith SA, C$_4$ Grasses Consortium, et al. 2010. The origins of C$_4$ grasslands: integrating evolutionary and ecosystem science. *Science* 328:587–91

Eldrett JS, Harding IC, Wilson PA, Butler E, Roberts AP. 2007. Continental ice in Greenland during the Eocene and Oligocene. *Nature* 446:176–79

Evans MJ, Derry LA, France-Lanord C. 2008. Degassing of metamorphic carbon dioxide from the Nepal Himalaya. *Geochem. Geophys. Geosyst.* 9:Q04021

Fisher MJ, Rao IM, Ayarza MA, Lascano CE, Sanz JI, et al. 1994. Carbon storage by introduced deep-rooted grasses in the South American savannas. *Nature* 371:236–38

Fox DL, Honey JG, Martin RA, Peláez-Campomanes P. 2012. Pedogenic carbonate stable isotope record of environmental change during the Neogene in the southern Great Plains, southwest Kansas, USA: oxygen isotopes and paleoclimate during the evolution of C$_4$-dominated grasslands. *Geol. Soc. Am. Bull.* 124:431–43

Gillett NP, Arora VK, Flato GM, Scinocca JF, von Salzen K. 2012. Improved constraints on 21st-century warming derived using 160 years of temperature observations. *Geophys. Res. Lett.* 39:L01704

Glaser B. 2007. Prehistorically modified soils of central Amazonia: a model for sustainable agriculture in the twenty-first century. *Philos. Trans. R. Soc. Lond. Ser. B* 362:187–96

Gohn GS, Koeberl C, Miller KG, Reimold WU. 2009. Deep drilling in the Chesapeake Bay impact structure: an overview. *Geol. Soc. Am. Spec. Pap.* 458:1–20

Grieser CJ, Rudolf B. 2005. A new monthly precipitation climatology for the global land areas for the period 1951 to 2000. *Ger. Weather Serv. Offenbach Clim. Status Rep.* 2004:181–90

Hirota M, Holmgren M, van Nes EG, Scheffer M. 2011. Global resilience of tropical forest and savanna to critical transitions. *Science* 334:232–35

Isaacson PE, Díaz-Martínez E, Grader GW, Kalvoda J, Babek O, Devuyst FX. 2008. Late Devonian-earliest Mississippian glaciation in Gondwanaland and its biogeographic consequences. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 268:126–42

Jacobson AD, Blum JD, Walter LM. 2002. Reconciling the elemental and Sr isotope composition of Himalayan weathering fluxes: insights from the carbonate geochemistry of stream waters. *Geochim. Cosmochim. Acta* 66:3417–29

(Berner & Berner 1996) Provides a textbook treatment of the carbon cycle and other environmentally significant geochemical reactions.

(Berner 1997) Proposes Paleozoic cooling because of early plant evolution.

(Edwards et al. 2010) Reviews the Cenozoic evolution of grasses.

Katz ME, Cramer BS, Toggweiler JR, Esmay G, Liu C, et al. 2011. Impact of the Antarctic Circumpolar Current development on late Paleogene ocean structure. *Science* 332:1076–79

Kennett JP. 1977. Cenozoic evolution of Antarctic glaciation: the circum-Antarctic oceans and their impact on global paleoceanography. *J. Geophys. Res.* 82:3843–59

Kent DV, Muttoni G. 2008. Equatorial convergence of India and early Cenozoic climate trends. *Proc. Natl. Acad. Sci. USA* 105:16065–70

Kidder DL, Worsley TR. 2012. A human-induced hothouse climate? *GSA Today* 22(2):1–11

Kiely M. 2010. *Carbon Farming Handbook*. Goolma, NSW: Carbon Farmers Aust. 189 pp.

Lacis AA, Schmidt GA, Rind D, Ruedy RA. 2010. Atmospheric CO_2: principal control knob governing Earth's temperature. *Science* 330:356–59

Lehmann J, Skjemstad J, Sohi S, Carter J, Barson M, et al. 2008. Australian climate-carbon cycle feedback reduced by soil black carbon. *Nat. Geosci.* 1:832–35

Lenton TM, Crouch M, Johnson M, Pires N, Dolan L. 2012. First plants cooled the Ordovician. *Nat. Geosci.* 5:86–89

Lucas SG, Foss SE, Mihlbacher MC. 2004. *Achaenodon* (Mammalia, Artiodactyla) from the Eocene Clarno Formation, Oregon, and the age of the Hancock Quarry local fauna. In *Paleogene Mammals*, ed. SG Lucas, KE Zeigler, PE Kondrashov. *N. M. Mus. Nat. Hist. Sci. Bull.* 26:89–95

McInerney FA, Strömberg CAE, White JC. 2011. The Neogene transition from C_3 to C_4 grasslands in North America: stable isotope ratios of fossil phytoliths. *Paleobiology* 37:23–49

Metzger CA, Retallack GJ. 2010. Middle Miocene climate change in the Australian outback. *Aust. J. Earth Sci.* 57:871–85

Mihalyi K, Gucsik A, Szabo J, Berczi S. 2009. Facts, theories and further questions around the Ries-Steinheim simultaneous impact event: a review. *Lunar Planet. Sci. Conf. Abstr.* 40:1542

Mihlbachler MC, Rivals F, Solounias N, Semprebon GM. 2011. Dietary change and evolution of horses in North America. *Science* 331:1178–81

Milton Moore R, ed. 1970. *Australian Grasslands*. Canberra: Aust. Natl. Univ. Press. 455 pp.

Misra S, Froelich PN. 2012. Lithium isotope history of Cenozoic seawater: changes in silicate weathering and reverse weathering. *Science* 335:818–23

Myhre G, Kvalevåg MM, Schaaf CB. 2005. Radiative forcing due to anthropogenic vegetation change based on MODIS surface albedo data. *Geophys. Res. Lett.* 32:L2410

Odadi WO, Karachi MK, Abduirazak SA, Young TP. 2011. African wild ungulates complete with or facilitate cattle depending on season. *Science* 333:1753–55

Pagani M. 2002. The alkenone-CO_2 proxy and ancient atmospheric carbon dioxide. *Philos. Trans. R. Soc. Lond. Ser. A* 360:609–32

Pagani M, Caldeira K, Berner R, Beerling DJ. 2009. The role of terrestrial plants in limiting atmospheric CO_2 decline over the past 24 million years. *Nature* 460:85–89

Pearson PN, Foster GL, Wade BS. 2009. Atmospheric carbon dioxide through the Eocene-Oligocene transition. *Nature* 461:1110–13

Pik R, Marty B, Carignan J, Yirgu G, Ayalew T. 2008. Timing of East African Rift development in southern Ethiopia: implication for mantle plume activity and evolution of topography. *Geology* 36:167–70

Prasad V, Strömberg CAE, Alimohammadian H, Sahni A. 2005. Dinosaur coprolites and the early evolution of grasses and grazers. *Science* 310:1177–80

Raymo ME, Ruddiman WF. 1992. Tectonic forcing of late Cenozoic climate. *Nature* 359:117–22

Retallack GJ. 1990. The work of dung beetles and its fossil record. In *Evolutionary Paleobiology of Behavior and Coevolution*, ed. AJ Boucot, pp. 214–26. Amsterdam: Elsevier

Retallack GJ. 1991. *Miocene Paleosols and Ape Habitats from Pakistan and Kenya*. New York: Oxford Univ. Press. 346 pp.

Retallack GJ. 1997. Early forest soils and their role in Devonian global change. *Science* 276:583–85

Retallack GJ. 1998. Fossil soils and completeness of the rock and fossil record. In *The Adequacy of the Fossil Record*, ed. SK Donovan, CRC Paul, pp. 131–62. Chichester, UK: Wiley

Retallack GJ. 2000. Ordovician life on land and early Paleozoic global change. *Paleontol. Soc. Pap.* 6:21–45

Retallack GJ. 2001. Cenozoic expansion of grasslands and global cooling. *J. Geol.* 109:407–26

(Kennett 1977) Outlines the case that thermal isolation of Antarctica caused Cenozoic global cooling.

(Kiely 2010) Outlines a variety of farming techniques for building soil carbon.

(Mihlbachler et al. 2011) Summarizes the evolution of horses and other grassland ungulates.

(Milton Moore 1970) Contrasts native Australian grasslands with introduced Eurasian agroecosystems.

(Raymo & Ruddiman 1992) Attributes Cenozoic climatic cooling to mountain uplift, especially of the Himalaya.

(Retallack 2001) Proposes grassland evolution as a biological force for global cooling.

Retallack GJ. 2004. Soils and global change in the carbon cycle over geological time. In *Treatise on Geochemistry*, ed. HD Holland, KK Turekian, 5:581–605. Oxford: Pergamon

Retallack GJ. 2005. Pedogenic carbonate proxies for amount and seasonality of precipitation in paleosols. *Geology* 33:333–36

Retallack GJ. 2007a. Cenozoic paleoclimate on land in North America. *J. Geol.* 115:271–94

Retallack GJ. 2007b. Coevolution of life and earth. In *Earth Evolution*, ed. D Stevenson, pp. 214–320. Amsterdam: Elsevier

Retallack GJ. 2007c. Palaeosols. In *Handbook of Paleoanthropology*, Vol. 1: *Principles, Methods and Approaches*, ed. W Henke, I Tattersall, pp. 383–408. Berlin: Springer

Retallack GJ. 2009a. Greenhouse crises of the past 300 million years. *Geol. Soc. Am. Bull.* 121:1441–55

Retallack GJ. 2009b. Refining a pedogenic CO_2 paleobarometer for quantifying the middle Miocene greenhouse spike. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 281:57–65

Retallack GJ. 2012. Mallee model for Mesozoic and early Cenozoic mammalian communities. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 342–43:111–29

Retallack GJ, Huang C-M. 2011. Ecology and evolution of Devonian trees in New York, USA. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 299:110–28

Retallack GJ, Krull ES, Bockheim JG. 2001. New grounds for reassessing palaeoclimate of the Sirius Group, Antarctica. *J. Geol. Soc. Lond.* 158:925–35

Retallack GJ, Rember WC. 2011. New estimates of the middle Miocene atmospheric carbon dioxide maximum from fossil *Ginkgo* leaves of Idaho. *Or. Acad. Sci. Proc.* 70:31

Retallack GJ, Veevers JJ, Morante R. 1996. Global early Triassic coal gap between late Permian extinction and middle Triassic recovery of peat-forming plants. *Bull. Geol. Soc. Am.* 108:195–207

Retallack GJ, Wynn JG, Fremd TJ. 2004. Glacial-interglacial-scale paleoclimatic changes without large ice sheets in the Oligocene of central Oregon. *Geology* 32:297–300

Rowley DB, Currie BS. 2006. Palaeo-altimetry of the late Eocene to Miocene Lunpola Basin, central Tibet. *Nature* 439:677–81

Royer DL, Berner RA, Park J. 2007. Climate sensitivity constrained by CO_2 concentrations over the past 420 million years. *Nature* 446:530–32

Sanderman J, Farquharson R, Baldock J. 2010. *Soil Carbon Sequestration Potential: A Review for Australian Agriculture*. Melbourne: CSIRO. 81 pp.

Savory A, Butterfield J. 1999. *Holistic Management: A New Framework for Decision Making*. Washington, DC: Island. 616 pp.

Scher HD, Martin EM. 2006. Timing and climatic consequences of the opening of Drake Passage. *Science* 312:428–30

Schmidt MWI, Torn MS, Abiven S, Dittmar T, Guggenberger G, et al. 2011. Persistence of soil organic matter as an ecosystem property. *Nature* 478:49–56

Sheldon ND, Hamer JMM. 2010. Evidence for an early sagebrush ecosystem in the latest Eocene of Montana. *J. Geol.* 118:435–45

Singh P. 2010. Geochemistry and provenance of stream sediments of the Ganga River and its major tributaries in the Himalayan region, India. *Chem. Geol.* 269:220–36

Smith WHF, Sandwell DT. 2012. *Predicted seafloor topography*. NGDC data announcement 94-MGG-04, NOAA Natl. Geophys. Data Cent., Boulder, CO. **http://www.ngdc.noaa.gov/mgg/fliers/94mgg04.html**

Staver AC, Archibald S, Levin SA. 2011. The global extent and determinants of savanna and forest as alternative biome states. *Science* 334:230–32

Strömberg CAE. 2005. Decoupled taxonomic radiation and ecological expansion of open-habitat grasses in the Cenozoic of North America. *Proc. Natl. Acad. Sci. USA* 102:11980–84

Strömberg CAE, McInerney FA. 2011. The Neogene transition from C_3 to C_4 grasslands in North America: assemblage analysis of fossil phytoliths. *Paleobiology* 37:50–71

Sundell KA. 2004. Paleoecology of oreodonts (Merycoidodontidea) within the White River Formation: prairie dogs of the Oligocene. *Geol. Soc. Am. Abstr.* 36(5):380

Suttle JM, Reynolds SG, Batello C, eds. 2005. *Grasslands of the World*. Rome: FAO. 22 pp.

(Retallack 2007a) Provides detailed North American paleosol records of Cenozoic paleoclimate.

(Suttle et al. 2005) Summarizes distribution, physical constraints, biota, and management of the major grasslands regions of the world.

van de Wal RSW, de Boer B, Lourens LJ, Köhler P, Bintanja R. 2011. Reconstruction of a continuous high resolution CO_2 record over the past 20 million years. *Clim. Past* 7:1459–69

van Valkenburgh B, Sacco T, Wang X. 2003. Pack hunting in Miocene borophagine dogs: evidence from craniodental morphology and body size. *Am. Mus. Nat. Hist. Bull.* 279:147–62

Vesey-Fitzgerald D. 1973. Grazing succession in East African game animals. *J. Mammal.* 41:161–72

Volk T. 1989. Rise of angiosperms as a factor in long-term climatic cooling. *Geology* 17:107–10

West AJ, Bickle MJ, Collins R, Brasington J. 2002. Small-catchment perspective on Himalayan weathering fluxes. *Geology* 30:355–58

Willenbring JK, von Blanckenburg F. 2010. Long-term stability of global erosion rates and weathering during late-Cenozoic cooling. *Nature* 465:211–14

Wynn JG, Bird MI, Vellen L, Grand-Clement E, Carter J, Berry SL. 2005. Continental-scale measurement of the soil organic pool with climatic, edaphic and biotic controls. *Glob. Biogeochem. Cycles* 20:GB1007

Zachos J, Pagani M, Sloan L, Thomas E, Billups K. 2001. Trends, rhythms and aberrations in global climate. *Science* 292:686–93

Zinke PJ, Stangenberger AG, Post WM, Emanuel WR, Olson JS. 1984. Worldwide organic soil and nitrogen data. *Oak Ridge Natl. Lab. Environ. Sci. Div. Publ.* 2212:1–141

RELATED RESOURCES

Carbon Farmers of Australia: **http://www.carbonfarmersofaustralia.com.au/CarbonFarmers/index.html**

Department of Geological Sciences, University of Oregon: **http://pages.uoregon.edu/dogsci/doku.php?id=directory/faculty/greg/about**

**Annual Review
of Earth and
Planetary Sciences**

Volume 41, 2013

Contents