



Cycling of Beryllium and Carbon through hillslope soils in Iowa

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Abstract. Isotopes of Be and C were used to reconstruct loess accumulation, hillslope evolution, and agricultural modification in soils of western Iowa. While both elements are derived from additions by the atmosphere (via plants in the case of carbon), the differences in element cycling allow erosional and depositional processes to be separated from biochemical processing. Based on ¹⁰Be, loess accumulation likely occurred simultaneously with hillslope degradation. Rates of loess accumulation declined five-fold between early stages (late Pleistocene and early Holocene) and later stages (late Holocene) of accumulation, but the absolute timing of accumulation requires independent dating methods. Based on ¹⁴C measurements, plant inputs and decomposition are significant near the surface, but below 1–1.5 m carbon inputs are minimal and decomposition is nearly arrested. The amount of carbon below 1.5 m is constant (0.1%) and is composed of soil organic matter that was buried by loess. Agricultural modification results in a dramatic redistribution of ¹⁰Be through soil erosion and deposition. By contrast, the redistribution of soil organic matter is masked by the rapid cycling of C through the topsoil as it continually decomposes and is replaced by plant inputs.

Introduction

Much attention has been paid to processes of C cycling in a variety of terrestrial environments and under a range of land use covers and land uses. The primary mechanism for C uptake onto land is photosynthesis. However, in order for C uptake to be sustained for more than a growing season, C must be protected from plant respiration, decomposition, and fire. In soils, protection from decomposition is evident in (1) lignin-rich materials that are less easily digested by decomposers (Melillo et al. 1982), (2) storage in cold, often deep layers (Trumbore and Harden 1997), or (3) storage in wet, anoxic environments (Gorham 1991). Physical disturbances – by plowing or destruction of cold, deep, or wet conditions – could result in a net release of C to the atmosphere based in part on accelerated decomposition (Johnson 1992; Davidson and Ackerman 1993) or a net uptake of C (Stallard 1998) if C is buried in sediment or water and C is replaced by continued or enhanced production (Harden et al. 1999).

Research on radiocarbon as a tracer to atmosphere-geosphere interactions includes a variety of papers on surface ages and carbon cycling in soils. Soil ages, sometimes referred to as “mean residence times” for soil organic matter, help to resolve the balance between radiodecay, inputs, and decomposition rates Trumbore and Torn (2002); (Wang et al. 1996). Decomposition rates and soil respiration studies have been greatly strengthened by analyzing ^{14}C of soil fractions (Trumbore et al. 1990) and soil gases (Trumbore et al. 1996). In addition, stable isotopes of carbon are also useful for soil processes, especially when shifts in vegetation have occurred naturally (Balesdent et al. 1987) or under agricultural management (Townsend et al. 1995).

The mechanism of replacement of C during erosion and sedimentation contrasts strongly with mechanisms by which ^{10}Be interacts between the atmosphere and geosphere. First and foremost, ^{10}Be , generated in the atmosphere by collision of N and O (Brown et al. 1981; Brown 1987), has an ionic radius and charge that allow a very strong sorption to particulate matter. Whether entering soils as wet fall or dry fall, Be ions likely remain attached to soil particles and are transferred to new surfaces when primary minerals alter and their soluble elements are leached through the soil profile (Barg et al. 1997). Research on ^{10}Be , and more recently ^9Be , has addressed issues on timing of landscape development (Pavich and Vidic 1993; Seidl et al. 1997; Bierman 1994; Monaghan et al. 1983), hillslope creep (McKean et al. 1993), weathering and formation of minerals (Barg et al. 1997; Monaghan et al. 1986), sedimentation (Seidl et al. 1997), and dust accumulation and the genesis of loess/paleosol sequences (Gu et al. 1997). While sources of ^{10}Be in soils include both atmospheric fallout and the radiodecay of *in situ* minerals, our studies are focused on sources from atmospheric additions. Although much of this ^{10}Be is deposited in dust (dry fall) and/or precipitation (wet fall), local reworking of soils as dust or colluvium can be important sources of ^{10}Be as well (Gu et al. 1996).

Given the contrast between the behaviors of Be and C in soils, we attempted to examine the history of dust accumulation, hillslope evolution, and organic matter accumulation by examining loess soils of western Iowa. Questions to be addressed include:

1. How is the cycling of carbon impacted by landform processes such as loess accumulation, hillslope degradation and agricultural management, and how can ^{10}Be cycling be used to understand carbon and landform processes?
2. Did loess accumulation occur episodically and was hillslope evolution a subsequent alteration of the landscape, or did these processes occur simultaneously?
3. What is the origin of organic matter stored in deep loess deposits? Is it ancient detrital or glacial carbon or is it related to more recent soil processes?
4. How has agricultural management impacted rates and processes of soil and landform development?

Methods

We examined the Be and C isotopic compositions of loess soils of western Iowa using a pair of undisturbed prairie and cultivated cropland sites and by sampling soils at stable, erosional, and depositional slope positions. The prairie site, Dinesen Prairie, is a 20-acre hillslope located near the town of Harlan (Figure 1). The cropland site, converted from prairie ~150 years ago, is a 30-acre drainage basin located in the USDA Deep Loess Research Station near the town of Treynor, ~100 km to the south of the prairie site. Soils throughout much of the area are developed in loess deposits (wind-blown silts) originating from glacial to post-glacial outwash along the Missouri River and distal loess sources in Nebraska (Bettis 1990; Muhs and Bettis 2000). Since the early Holocene, valleys have accumulated several generations of alluvium with admixtures of loess, most recently after settlement ~150 years ago (Bettis 1990). Among the first European visitors were Lewis and Clark, who crossed the Missouri River just north of Council Bluffs and settlement began later when the Mormon Trail established a wintering-over point. Permanent agriculture and settlement began in the 1860's, with corn, clover, wheat rotations. In the 1950's modern fertilization was introduced, and corn-clover-wheat rotations gave way to monocrops of corn or soybean (Manies et al. 2001).

The prairie site is surrounded by agricultural croplands and exists as a refuge for birds, tallgrass prairie plants, and gophers. The local slopes vary from about 0 to 3% on the ridgetops, 12% on the steepest midslope, and 0% on the depositional lower slope. The soils are characterized as Mollisols, (Monona series) with very deep, very dark A horizons.

The cropland site near Treynor is one of several experimental watersheds that have been farmed since ~1860. Local slopes are quite similar to the hillslope at the Dineson Prairie, but midslopes are somewhat steeper (15%). The soils are also characterized as the Monona series on uneroded ridgetops, but are generally lighter in color and have thinner A horizons than soils at the Dinesen Prairie. Soils on the sideslopes are mapped as the eroded or severely eroded Graphton series and have topsoils only 20 cm thick that overlie much lighter subsoils. Depositional lower slopes, generally mapped as Napier series, have dark, very thick A horizons.

At both Dinesen Prairie (DP) and the Treynor (TR) cropland sites, soils were sampled at a relatively uneroded interfluvial, or ridgetop (designated R for Ridgetop), an erosional upper midslope (designated U for Upper), and depositional lower (L) slope positions. Using a combination of coring devices or soil excavations, profiles were sampled to depths of 5, 10, 20, and 40 cm, and every 20 cm down to 200 or 300 cm. Bulk density was measured by the core method (described in more detail in Harden et al. (2001)). Triplicate profiles were sampled for the upper 1 m.

Total carbon (TC) in soil samples was determined by measuring the carbon dioxide (CO₂) produced by combusting the sample in a stream of oxygen (O₂). Total carbon measurements were made using a Fisons NA1500 elemental analyzer (EA)/Optima isotope ratio mass spectrometer (IRMS). Inorganic carbon (IC) was determined by measuring the CO₂ generated by heating a sample at 105 °C in acid using a UIC coulometer. Organic carbon (C) was calculated as the difference between

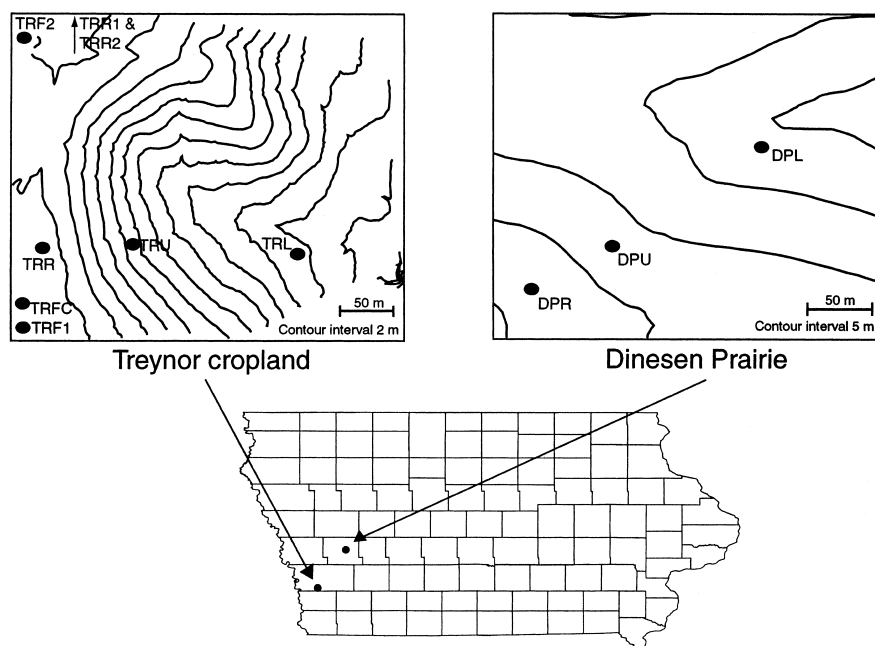


Figure 1. Map of study area showing location of study sites

TC and IC. A Fisons NA1500 EA/Optima IRMS was also used for the determination of total nitrogen and for ^{15}N and ^{13}C measurements. Radiocarbon content was measured on untreated bulk soil by sealing enough homogenized sample to make ~ 1 mg of C with cupric oxide wire and a small piece of silver wire in an evacuated, quartz tube. The CO_2 produced is purified cryogenically and then reduced to graphite using the zinc- or iron-catalyzed reduction method described in Vogel (1984). The ratio of ^{14}C to ^{13}C and ^{12}C atoms is measured directly from the graphite target using a high-energy accelerator as an inlet to a mass spectrometer (AMS). We report ^{14}C data in Fraction Modern or FM units. In calculating the FM, we correct for isotopic fractionation effects that occur as a result of photosynthesis (the origin of soil organic matter) by correcting both sample and standard to a measured ^{13}C value of the same sample. The standard oxalic acid is corrected in the same way, to -19 per mil. Because bulk soil organic matter is a mixture of old and new plant matter, these values do not represent a “date”, but rather represent a history or average age after mixing, decomposition, leaching and burial. Nevertheless, we also report (Tables 1 and 2) corrected radiocarbon ages using (Stuiver and Reimer 1993), in which a bias in the calculated radiodecay ages are mathematically corrected for comparison to tree-ring ages. Inorganic carbon was not present in the bulk fraction of most samples; however some deep layers of DPU and TRU and TRL contained 0.2 to 1% carbonate which is reflected in ^{13}C values $> \approx -12$ (Figure 2d). All values for inorganic, total, and organic carbon data are reported for replicate profiles in Harden et al. (2001)

Table 1. Ages and deposition rates of loess material at Dineson Prairie and Treynor Ridgetops

Profile ID	Depth (cm)	Bulk Density (g/cm ³)	Delta ¹³ C (bulk fraction)	Radiocarbon (FM ^a)	Calibrated Age ^b (Yr B.P.)	¹⁰ Be Age 1 (Yr B.P. ^c)	¹⁰ Be Age 2 (Yr B.P. ^d)	¹⁰ Be Age 3 (Yr B.P. ^e)	Deposition Rate 2 ^f (g/cm ² /yr)
Dineson Prairie Ridgetop									
DPPR3	5	0.95	-20.50	1.0514	modern	0	0	0	0.011
DPPR3	10	1.09	-18.74	0.9969	46	868	438	198	0.011
DPPR3	20	1.10	-17.66	0.9741	336	1,892	954	432	0.009
DPPR3	40	0.78	-16.80	0.9433	567	4,402	2,221	1,006	0.009
DPPR3	60	1.03	-16.67	0.8575	1,224	7,733	3,902	1,768	0.013
DPPR4	80	1.05	-16.12	0.7710	2,156	10,942	5,523	2,504	0.024
DPPR4	100	1.19	-17.60	0.6911	3,257	12,739	6,432	2,916	0.036
DPPR4	120	1.31	-19.19	0.6073	4,562	14,056	7,098	3,218	0.045
DPPR4	140	1.29	-21.19	0.4955	6,491	15,239	7,696	3,490	0.050
DPPR4	160	1.30	-22.44	0.4241	7,774	16,275	8,220	3,728	0.051
DPPR4	180	1.39	-23.59	0.3424	9,601	17,310	8,745	3,966	0.064
DPPR4	200	1.31	-23.94	0.2621	12,989	18,189	9,190	4,168	0.060
DPPR4	220	1.31	-24.10	-	-	19,068	9,634	4,370	0.060
DPPR4	240	1.27	-24.66	-	-	19,953	10,083	4,573	0.058
DPPR4	260	1.30	-25.35	-	-	20,837	10,530	4,777	0.059
DPPR4	280	1.33	-24.93	-	-	21,721	10,978	4,980	0.030
DPPR4	300	1.34	-24.93	-	-	23,540	11,438	5,189	-
Treynor Cropland Ridgetop									
TRPRI	5	0.95	-14.26	1.0218	0.0	0	0	0	0.013
TRPRI	10	1.01	-14.14	1.0226	0.0	725	359	165	0.013
TRPRI	20	1.22	-14.26	0.9811	320	1,491	738	338	0.013
TRPRI	40	1.33	-13.74	0.7775	2,041	3,420	1,692	776	0.021
TRPRI	60	1.15	-14.70	0.7319	2,764	5,989	2,963	1,360	0.030
TRPRI	80	1.28	-16.71	0.6524	3,742	7,562	3,743	1,718	0.032
TRPRI	100	1.25	-18.31	0.5717	5,329	9,198	4,553	2,090	0.047
TRPRI	120	1.24	-19.65	0.5304	5,955	10,278	5,089	2,336	0.057
TRPRI	140	1.21	-21.95	0.4999	6,448	11,149	5,520	2,535	0.056
TRPRI	160	1.23	-22.49	0.4389	7,610	12,025	5,955	2,734	0.060
TRPRI	180	1.24	-22.79	0.3463	9,581	12,854	6,366	2,923	0.028
TRPRI	200	1.11	-23.38	0.2853	11,947	14,620	7,242	3,326	-
TRPRI	220	1.24	-23.87	-	-	-	-	-	-
TRPRI	240	1.25	-24.76	-	-	-	-	-	-
TRPRI	260	1.18	-24.05	-	-	-	-	-	-
TRPRI	280	1.21	-23.76	-	-	-	-	-	-
TRPRI	300	1.24	-7.46	-	-	-	-	-	-

^a FM: fraction modern or ¹⁴C/¹²C. See Harden et al. (2001) for +/- and lab numbers.^b CALIB program, University of Washington Quaternary Sciences Laboratory^c Using Equation (1) and atmospheric deposition rate of ¹⁰Be 1.3 * 10⁸ atom/yr^d Using Equation (1) and atmospheric deposition rate of ¹⁰Be 1.3 * 10⁸ atom/cm²/yr plus 1.32 * 10⁶ atom/cm²/yr as intrinsic dust^e Using Equation (1) and atmospheric deposition rate of ¹²Be 1.3*10⁸ atom/cm²/yr plus 4.4 * 10⁶ atom/cm²/yr as intrinsic dust^f Dividing bulk density by Be age 2

Table 2. Examples of Soil Data for Dineson Prairie and Treynor Midslope and Lower Slope Positions

Profile ID	Depth (cm)	Bulk Density (g/cm ³)	Delta ¹³ C (bulk fraction)	Radiocarbon (FM ^a)	Profile ID	Depth (cm)	Bulk Density (g/cm ³)	Delta ¹³ C (bulk fraction)	Radiocarbon (FM ^a)
Selected from replicate profiles reported in Harden et al. (2001)									
Dineson Prairie Midslope									
DPPU2a	5	0.683	-21.58	1.13409	Dineson Prairie Lower Slope	5	0.755	-19.45	1.00075453
DPPU2a	10	1.104	-19.73	1.10093	DPPL2a	10	1.075	-17.79	1.0010745
DPPU2a	20	0.975	-18.07	1.02162	DPPL2a	20	0.900	-15.89	1.00090046
DPPU2a	40	0.682	-17.04	0.97285	DPPL2a	40	0.734	-15.06	1.00073424
DPPU2a	60	0.538	-15.99	0.93822	DPPL2a	60	0.799	-15.52	1.00079862
DPPU2a	80	1.032	-16.33	0.92652	DPPL2a	80	1.420	-13.78	1.00142007
DPPU2a	100	0.765	-17.91	0.87368	DPPL6	100	1.08	-13.91	1.00108132
DPPU6	100	1.166	-19.08	-	DPPL6	120	1.27	-13.57	-
DPPU6	117	1.230	-21.15	-	DPPL6	140	1.28	-13.83	-
DPPU6	140	1.268	-7.97	-	DPPL6	160	1.31	-14.18	1.00126822
DPPU6	160	1.376	-3.03	-	DPPL6	180	1.47	-15.22	-
DPPU6	180	1.211	-2.71	-	DPPL6	200	1.42	-13.57	1.00128207
DPPU6	200	1.524	-2.75	-	DPPL6	218	1.50	-13.65	-
-	-	-	-	-	DPPL6	240	0.85	-14.10	-
-	-	-	-	-	DPPL6	260	-	-13.98	1.00131042
Treynor Cropland Midslope									
TRPU1a	5	1.128	-13.13	1.12058	Treynor Cropland Lower Slope	5	0.021	-13.34	1.0744
TRPU1a	10	1.443	-13.31	1.09454	TRPL1a	10	0.025	-13.13	1.0574
TRPU1a	20	1.443	-14.76	0.95947	TRPL1a	20	0.019	-13.32	1.0383
TRPU1a	40	1.285	-17.66	0.801	TRPL1a	40	0.017	-12.40	0.9896
TRPU1a	60	1.344	-18.87	0.61176	TRPL1a	60	0.014	-13.15	0.8956
TRPU1a	80	1.306	-7.02	-	TRPL1a	80	0.018	-11.17	0.8463
TRPU1a	100	1.138	-3.54	-	TRPL1a	100	0.019	-11.63	0.797
-	-	-	-	-	-	160	1.392	-12.80	0.6302
-	-	-	-	-	-	200	1.333	-14.02	0.5384
-	-	-	-	-	-	260	1.278	-15.37	0.4455

^a FM: fraction modern. See Harden et al. (2001) for +/- and lab numbers.^r Dividing bulk density by Be age 2

Meteoric ^{10}Be adsorbed to soil particles was measured by Accelerator Mass Spectrometry. Aliquots of soil < 2 mm were ground, dried, weighed, and spiked with ^9Be . Soil was decomposed by $\text{HClO}_4 + \text{HF} + \text{HCl}$ and dried down. The dried solid cake was extracted by 3 N HCl and precipitated by NH_4OH to form $\text{Fe}(\text{OH})_3$. The precipitate was dissolved by 10% HF and re-precipitated to $\text{Fe}(\text{OH})_3$ by NH_4OH , while Be complex ion is formed in solution. The solution was dried at 250°C , treated with HCl and NH_4OH to form $\text{Be}(\text{OH})_2$, baked again at 550°C for a BeO target. The BeO was loaded into sample holders and analyzed for $^9\text{Be}/^{10}\text{Be}$ ratios at Lawrence Livermore Laboratory CAMS facilities. This sample preparation method was developed by Fouad Tera of Canegie Institution in Washington D.C.

Results and discussion

Trends in data

The ^{10}Be chemistry of uncultivated sites allows a general assessment of erosional and depositional processes during the history of loess deposition and hillslope degradation. Ridgetop soils are characterized by an accumulation of organic carbon and surface enrichment of radiocarbon and ^{10}Be isotopes (Figure 2a, b, c). Although others have found ^{10}Be to be most concentrated in deeper subsoils (B horizons) as a result of sorption to and translocation of secondary minerals (Pavich et al. 1984; Gu et al. 1997), the near-surface concentration of ^{10}Be in the uncultivated profiles (DP in Figure 2b) is likely due to being a relatively young, unweathered geomorphic surface. Low ^{10}Be concentrations throughout the midslope profiles (“upper” in Figure 2b) indicate that the particles are not exposed long enough to accumulate Be onto their surfaces. By contrast, the high Be concentrations of the depositional profiles (especially at depth; “lower” in Figure 2b) indicate that Be inputs are high and/or the particle surfaces are exposed long enough for Be to accumulate. The differences in ^{10}Be between the slope positions are maintained at depth, which likely indicates a long history of Be transfer from the upper to lower slopes. As a result, hillslope degradation has likely occurred simultaneously with loess accumulation in this upper part of the loess section.

The dynamics of carbon cycling during landscape development can be studied by comparing carbon chemistry to ^{10}Be for soils of different slope positions. Carbon and radiocarbon, although like ^{10}Be are concentrated at the surface, have similar depth profiles among ridgetop, midslope, and lower slope positions (Figure 2a, c), which indicates a very fundamental difference between the ways Be and C are cycled through soil. Whereas both elements ultimately have an atmospheric source, ^{10}Be is refractory (high valence, small size, high sorption potential) and therefore conservative to soil, while carbon is continually replaced by plant additions and decomposition losses.

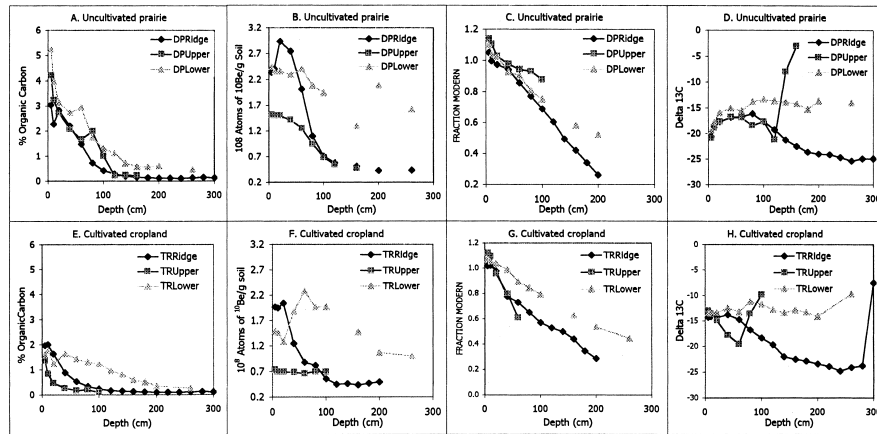


Figure 2. Depth profiles of %C, ^{10}Be , ^{14}C , and ^{13}C for Dinesen Prairie (2a–d) and Treynor Cropland soils (2e–g) for ridge, midslope, and lower slope positions. Data from Harden et al. (1999), for profiles DPPR3, 4; DPPU2a, b, c and DPPL2a, b, c (composite for ^{10}Be ; average for other analyses); TRPR1; TRPU2a, b, c.

Stable carbon isotopes of soil and loess at the uncultivated DP site appear to record a history of vegetation shifts that have been reported in other studies throughout the central US (Muhs et al. 1999). The shift in deep loess deposits that contain carbonate (below 1 m, Figure 2d, -23 to -16 delta ^{13}C values) is similar to the ^{13}C profiles of stalagmite limestone and pollen assemblages from central Iowa (Baker et al. 1996), which are hypothesized to represent a shift from boreal woodland to prairie during the Holocene or late Pleistocene. Stalagmites contain carbonate that has been shown to be in equilibrium with soil atmospheric CO_2 at the time of carbonate precipitation. There is a subsequent shift toward more negative ^{13}C values in the upper 0.5 m of soil organic matter (Figure 2d), which is consistent with a recent shift toward C4 plants. Radiocarbon ages of bulk organics and their stable isotopes are consistent with the regional pattern of climate and vegetation change, which suggests that the carbon in loess is derived from local to regional plant matter. The caveat to this interpretation of course is that soil processes may have altered (by making younger) the radiocarbon age of the sediment.

Radiocarbon chemistry helps to elucidate sources of organic carbon and replacement of soil carbon by plant inputs (turnover). The nearly linear decline with depth in ^{14}C values in uncultivated DP soils (Figure 2c) indicates a changing mixture of modern carbon (FM values around 1) with old carbon (^{14}C of 0.6 or less). Most soil ^{14}C trends of other studies show a dramatic, often exponential decline in ^{14}C with depth (Trumbore 1993; Wang et al. 1999), likely because most of the surface organic matter has much faster turnover and does not contain old forms of organic matter. A comparison of radiocarbon content of soils sampled in 1963 to that of soils sampled in 1997 (Figure 4) is a measure of the replacement of soil carbon over the past 30 years. As a result of above-ground weapons testing that peaked in 1964 (Trumbore et al. 1990), plants and subsequently soil organic matter became

labeled with ^{14}C after 1964. Although very fast cycling forms (“pools”) of soil organic matter would reflect only the previous year(s) of atmospheric ^{14}C , slower cycling forms (slower pools) reflect an accumulation of radiocarbon that decomposes over decades to millennia. The cropland soils are enriched in radiocarbon in the upper 20 cm (comparison of 1963 samples to the cropland samples from the ridgetop, TRR, Figure 4) indicating that plant inputs occurred mainly in the topsoil and were retained in stable pools of organic matter. By contrast, radiocarbon enrichment in the prairie soils indicates significant input and retention to depths of 1 m or more. These inputs are likely in the form of roots, dissolved organics, and soil mixing from the surface.

A comparison of radiocarbon to ^{10}Be data in uncultivated soils demonstrates the contrasts in element cycling and soil development between erosional and depositional landforms. Erosional midslope (DPU) and depositional lower slope (DPL) soils at the prairie have slightly greater carbon and ^{14}C enrichment than the ridgetop profiles (Figure 2c), which is consistent with the hypothesis (Harden et al. 1999) that dynamic replacement of C during erosion results in incorporation of enriched, bomb-spike ^{14}C . In this hypothesis, while uneroded soils maintain both older C and ^{14}C -enriched carbon, eroded soils lose both types to erosion but replace (through plant inputs) only the recent carbon that is ^{14}C enriched. Continual sedimentation at DPL, supported by ^{10}Be enrichment of deep sediments (Figure 2b), reflects the ^{14}C -enriched source from the midslope (DPU). By contrast, ^{10}Be at the eroding DPU site is very low in concentration, reflecting a long history of erosion from the site.

An assessment of both ^{10}Be and C data in cropland soils reflects impacts of erosion and cultivation on soil material and soil chemistry that have occurred over the past 150 years. Cropland soils have some dramatic contrasts to the undisturbed prairie. First, organic carbon content at stable ridgetops is significantly lower at cropland than prairie sites, yet ^{10}Be at the ridgetops is fairly similar in concentration, suggesting that little erosion has occurred at the cropland ridgetop and that the differences in carbon between prairie and cropland are not due to erosion. The lower carbon content at the cropland ridgetop is therefore due to accelerated decomposition and changes in biomass that occurred as a result of farming. At the eroding midslope, however, ^{10}Be is dramatically lower at the cropland than the prairie (Figure 2b, f). At the depositional sites, ^{10}Be concentrations are quite low in the upper 40 to 60 cm of the cropland site, but are similar to prairie-site sediments below about 40 cm. It appears that the eroded profiles have been truncated to depths well below the zone of Be concentration, producing the low depositional signature. Depositional sites may reflect a chronology where eroded sources were originally enriched in Be (at the soil surface) and then later depleted of Be as truncation progressed.

Using ^{10}Be to Model Loess Accumulation and Hillslope Degradation

We constructed simple mathematical models to understand the interplay between loess accumulation, hillslope degradation, agricultural modification, and soil for-

mation. We used ^{10}Be isotopes to estimate the timing and rate of loess accumulation and hillslope degradation. We then used ^{14}C and ^{13}C isotopes to understand inter-relationships of organic matter and soil formation with loess and hillslope processes.

Loess accumulation and hillslope degradation greatly affect ^{10}Be concentrations and rates of ^{10}Be accumulation through the profile. Ages and rates of loess accumulation can be approximated by the inventory of ^{10}Be that accumulates over time, assuming a constant rate of ^{10}Be input:

$$t = -1/\lambda \ln[1 - \lambda N/q] \quad (1)$$

where t , time in years; λ radiodecay of ^{10}Be ; N inventory of ^{10}Be in atom/cm^2 ; q , atmospheric fallout in $\text{atom}/\text{cm}^2/\text{yr}$ corrected for intrinsic or inherited dust. We first calculated ages using a total atmospheric flux (q) of $1.3 \times 10^8 \text{ atom}/\text{g}/\text{yr}$ or $1.3 \times 10^6 \text{ atom}/\text{cm}^2/\text{yr}$ for $100 \text{ cm}/\text{yr}$ rainfall (Monaghan et al. 1986) (Table 1). Based on a ^{10}Be inventory observed for the upper 260 cm at Dinesen Prairie Ridgetop (DPR), the loess at a depth of 300 cm would be $\sim 23,000$ yrs old (Table 1), which is too old based on a radiocarbon date on a snail shell from about 23 m depth (see Forman et al. (1992)). If we introduce a term for ^{10}Be inherited in the dust (“intrinsic” ^{10}Be of Gu et al. (1996)) then the upper loess is as young as 4400 yrs if the intrinsic Be flux is 5X that of the atmospheric deposition (Table 1, scenario 3). The latter example was included because by extending the concentration and bulk density data from our 3 m sample to a depth of 23 m where the loess section was dated at 20,000 yrs (see Forman et al. (1992)) a very high intrinsic dust flux is required. Moreover, an average concentration of $2.2 \times 10^8 \text{ atom}/\text{g}$ with $0.02 \text{ g dust}/\text{cm}^2/\text{yr}$ was used for the minimum age analysis because these values are consistent with paleodeposits modeled by Gu et al. (1996) (see Table 1 caption). However, these young ages also infer loess accumulation rates of $0.15 \text{ g}/\text{cm}^2/\text{yr}$ at 4,400 years to $0.02 \text{ g}/\text{cm}^2/\text{yr}$ historically, which are anomalously high for well dated loess sequences (Gu et al. 1996), but are more similar to rates for Pleistocene loess elsewhere in the Mississippi valley (Forman et al. 1992). The two extreme age scenarios, although greatly disparate, likely bracket the absolute ages of loess deposition for the uppermost 3 meters of sediment. As a compromise between the extremes, we devised an intermediate age for the loess based on intermediate values for intrinsic Be (Age scenario 2, Table 2). All age estimates assume that atmospheric deposition rates have not changed over time, and that inherited or “intrinsic” Be from reworked sources has also not changed over time. Indeed, even the relative ages could be wrong if either of these assumptions is not met. However, the chemistry of these loess sections has been shown to be fairly consistent in the upper 20 m (Muhs and Bettis 2000), suggesting that source areas for the loess have not changed recently. Nevertheless, independent radiometric ages throughout the section would better resolve the depositional history.

Dividing the soil mass by the age provides a rate of loess accumulation for the post-glacial history (Table 1). For both age scenarios, loess accumulation rates are higher by a factor of 5 in the earlier, basal sediments than the later, surficial sedi-

ments. The slowing accumulation rate toward the present is consistent with increasing ^{10}Be concentrations toward the surface (Figure 2b, Table 1), as atmospheric additions of ^{10}Be are allowed to accumulate in a stable soil surface.

Hillslope degradation at the uncultivated Dinesen Prairie was examined quantitatively by using the basic equations of Brown (1987) to examine the timing of hillslope degradation and its relationship to loess accumulation:

$$N = (q - \epsilon\eta)[1 - e^{(-\lambda t)}]/\lambda \quad (2)$$

Where N is ^{10}Be inventory (which we measure); q atmospheric fallout rate of ^{10}Be in atoms/cm²/yr; ϵ erosion rate in g/cm²/yr; η is concentration of ^{10}Be in eroding material; λ is the radiodecay rate constant for ^{10}Be ; and t is time elapsed. Using the ridgetop profile at the prairie site, we set $\epsilon = 0$ for a no-erosion control, which reduces Equation (2) to be the same as Equation (1). Therefore, loess ages and loess accumulation rates are defined by atmospheric fallout rates at the ridgetop (with or without intrinsic, inherited Be), and redistribution of loess along the hillslope is defined by ϵ , which is estimated from model-data comparisons. The concentration of eroding material η is assumed to be equal to uppermost layers of the midslope DPU at each time increment. Using the ridgetop rates and model ages (Table 1 model age 2) for loess accumulation, we then calculate the adjusted thickness of the DPU and DPL sites so that we can compare depth plots of data vs model results.

$$Z_{\text{soil}_{t_2}} = Z_{\text{soil}_{t_1}} + (\Delta Loess_{t_2} - \epsilon)/BD_{\text{soil}}*(t_2 - t_1) \quad (3)$$

where Z is depth in cm, Δ is accumulation rate and ϵ is erosion rate both in g sediment/cm²/yr, BD_{soil} is bulk density of initial loess material of 1.3 g/cm³ (Table 1, bulk density of deep loess), t is time. For each time increment ($t_2 - t_1$ from model age 2 in Table 1), the sediment thickness ($Z_{\text{soil}_{t_2}}$ in cm) is adjusted by loess accumulation and erosion rate. For plotting purposes, the basal age for the mid and lower slopes is adjusted so that $t = 0$ at $Z = 280$ cm; rates of loess accumulation after $t = 0$ are defined by loess accumulation rates at the ridgetop as adjusted for hillslope erosion and deposition (Figure 3 caption).

The curve shapes of the ^{10}Be model (Figure 3a) generally match the ^{10}Be data (Figure 3b) in that ^{10}Be accumulation rate is highest in the lower slope, lowest in the upper midslope, and intermediate in the ridgetop position. Using the intermediate age scenario (Table 1), a constant erosion rate of ~ 0.002 g/cm²/yr provides the best fit to ^{10}Be data at the eroded midslope DPU site (Figure 3a vs Figure 3b). By contrast, the minimum age scenario requires erosion rates of .009 g/cm²/yr for a good model-data fit, therefore the sensitivity of the erosion rates to initial ages and Be fallout rates is great. Regardless of the age model and regardless of the source for Be (whether atmospheric or intrinsic), however, these results demonstrate that the loess (dust) is blowing onto the slope and being redistributed by hillslope erosion throughout the model years. The dissimilarities between model and data curves are also noteworthy, particularly the separation of data curves (Figure 3b)

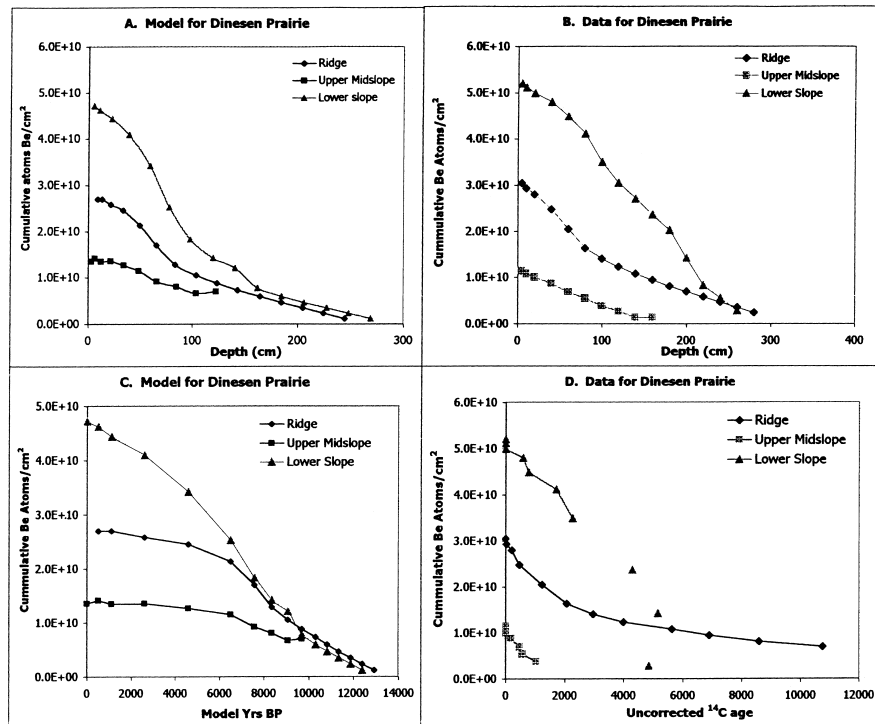


Figure 3. Model results and data for depth plots (a, b) and age plots (c, d) for Dinesen Prairie slope positions. Models (a, b) are based on Equations (2) and (3) starting at the basal depth and using η of 0.44×10^8 atom/g, q of 1.3×10^8 atom/g/yr, and ϵ of zero for ridgetop, 0.001 g/cm²/yr for eroding, and -0.003 g/cm²/yr for depositional sites. The ¹⁰Be concentration η for each time increment t was set to equal ¹⁰Be inventory of $t-1$ previous time increment, divided by bulk density of 1.3 and 20 cm depth increment. Model ¹⁰Be ages (c) are calculated from Equation (1) for ridgetop positions, initializing ¹⁰Be accumulation at 17,000 years at the base according to Model age 2 in Table 1. Data for ¹⁰Be inventories (b, d) and radiocarbon ages from Table 1, using uncalibrated age = $-8033(\ln FM)$, where $FM = ^{14}C/^{12}C$.

for the different slopes as compared to model curves (Figure 3a). It is possible that early erosion of the hillslopes involved other sources of Be (reworked dust) or dust accumulation rates that were not constant with time. Overall, however, the difference in Be accumulation rates among ridge, midslope, and lower slope positions (Figure 3a, b) indicate that hillslope degradation likely occurred simultaneously with loess accumulation for the past several millennia.

Organic carbon and ¹⁴C

Organic matter accumulation occurs on young (late Quaternary) landforms (Harden et al. 2001; Birkeland (1984, 1999)). Therefore, radiocarbon should be a sensitive measure of landscape change because soil carbon is replaced by plant inputs (turn-

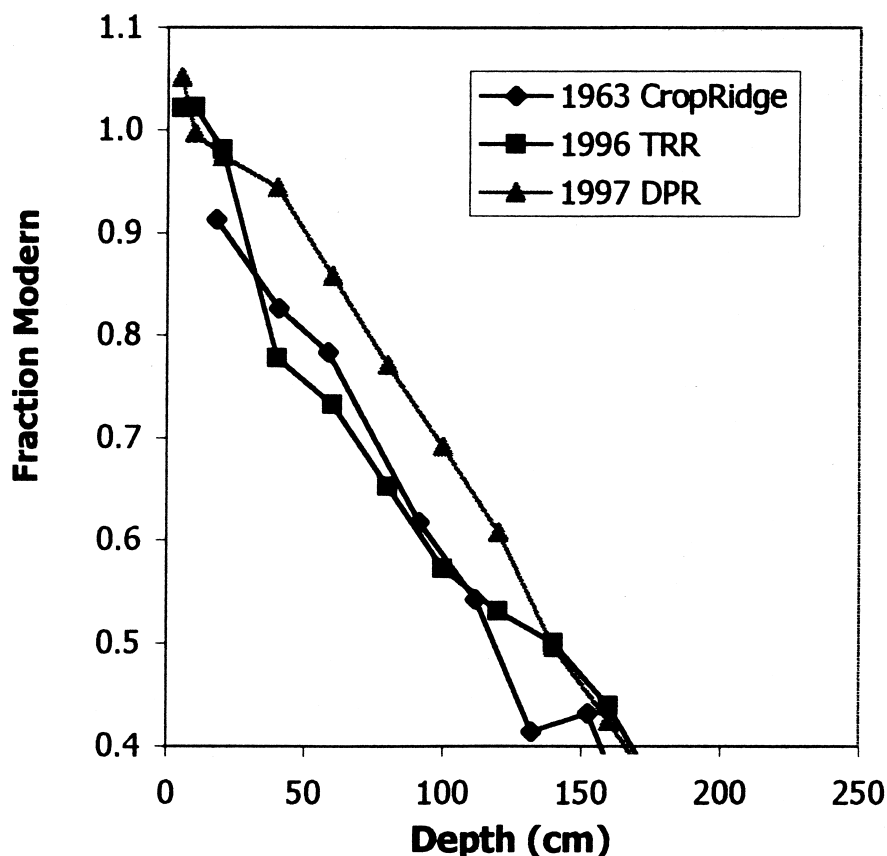


Figure 4. Radiocarbon content of 1963 and 1997 samples of ridgetop soils.

over) on timescales of years to millennia (Trumbore et al. 1996) and because the radiodecay half-life of ^{14}C is about 5700 yrs.

Bomb-spike radiocarbon is present in the upper 1.5 m of soil, whereas the organics below 1.5 m are not affected by bomb-spike radiocarbon (Figure 4) and therefore are not likely part of the modern soil and plant system. Bomb-radiocarbon may influence the curvilinear trend in the $\text{Be}-^{14}\text{C}$ curve of the depositional site (Figure 3d), since the model does not contain bomb-spike additions. Since even the slowest pools of organic matter (estimated at 600–1500 yrs by Manies et al. (2001)) would have decomposed within 500–1500 yrs since deposition, we can discern that decomposition rates are dramatically reduced below 1.5 m, which is consistent with the findings of Trumbore et al. (1996) and equations of Rosenbloom et al. (2001).

Within the upper meter, ^{14}C ages diverge from ^{10}Be ages (Table 1) according to assumptions about intrinsic (inherited) Be. Two possibilities exist for explaining the radiocarbon age of deep loess: (1) newly accumulating soil organic matter is generated at the surface and then buried by accumulating loess, or (2) new soil organic

material is mixed with old or even “radiocarbon dead” organic material during loess deposition. In model 3, $\sim 80\%$ of the Be would have to be “intrinsic” in order to satisfy the 20,000 yr age at 23 m (see above). However, based on ^{14}C content of the soils, no more than 15% of the carbon could be derived from ancient (radiocarbon dead) deposits because modern carbon has a maximum FM value of 1.8. Therefore, soil organic matter incorporated into loess likely accounts for the majority of buried carbon in the loess.

Agricultural modification at the Treynor Cropland Site

Since the advent of agriculture in Iowa ~ 150 years ago, the amount of ^{10}Be added by atmospheric fallout has been exceedingly small (1.9×10^8) relative to the amount accumulated in the profile over several millennia ($1\text{--}3 \times 10^{10}$). Although some soil environments could potentially alter the sorption capacity of ^{10}Be , the soils of this study are nearly pH neutral, well drained, and are not likely to alter the ^{10}Be sorption or concentration in only 150 years. Models for ^{10}Be for the past 150 years, therefore, do not need to account for radiodecay and therefore ^{10}Be can be used as a tracer that is conservative to the soil as it mixes and moves with agricultural management practices.

Using the Dinesen Prairie midslope as a control site, we estimated that the eroding midslope TRU soil has likely lost $\sim 42 \text{ g/cm}^2$ of sediment over the past 150 years. Whereas the depth profile of ^{10}Be at the cropland ridgetop is fairly similar to that at the prairie ridgetop (Figure 2b, f), the upper profile at the eroded midslope position at the cropland (Figure 2f, TRU) has concentrations typical of depths >100 cm at the prairie site (Figure 2b, DPU). The simplest way to interpret this difference is to assume that the cropland site lost about 60 cm of soil to erosion. The abrupt change in ^{13}C (Figure 2h, TRU) and the steep decline in C and ^{14}C in this profile supports an erosion into deep, old layers that contain carbonate. 60 cm represents about 42 g/cm^2 of soil that eroded over a period of 150 years, which translates to a rate of $0.27 \text{ g/cm}^2/\text{yr}$. By comparison, Manies et al. (2001) estimated erosion rates of about 0.2 to $0.6 \text{ g/cm}^2/\text{yr}$ from the cropped Treynor watershed (not just the midslope) based on cropping and management history and on recent ^{137}Cs data (Spomer et al. 1985).

While there is evidence for recent deposition at the cropland’s lower slope, the depositional history is more complicated than the erosional history at the midslope. Below about 80 cm, ^{10}Be profiles are similar between the cropland and prairie sites (Figure 2b, 2f), but above 80 cm, cropland soils have relatively low ^{10}Be concentrations at the surface and intermediate concentrations at 60–80 cm. We think that the uppermost layers (to 20 cm) are derived from a mixture of deep subsoils (typically $0.7 \times 10^8 \text{ }^{10}\text{Be}$ atoms/g; Figure 2f) and topsoils from eroding slopes. In the field, we noted a subtle soil boundary at 40 cm that appeared to be a buried surface at the depositional cropland (TRL), which is consistent with the peak in ^{10}Be concentration at that site. Treynor’s original surface could be at 40 cm depth, or it is also possible that ~ 40 cm of soil was first eroded from the site and then replaced with more sediment.

Carbon and radiocarbon at the cropland (Treynor) sites show a fundamentally different response to agricultural modification as compared to ^{10}Be . Whereas the ^{10}Be depth profile is dramatically truncated by erosion at the eroding midslope (Figure 2f, TRU), the radiocarbon content of the surface layers is maintained at a value similar to the prairie surface, and the radiocarbon profile is steepened rather than truncated (Figure 2g). At the depositional sites, radiocarbon appears enriched throughout the upper meter of soil, rather than being similar to deep layers at the eroded site. It is likely that the lower slope at the agricultural site is receiving sediment from the eroding topsoil (which is consistent with the ^{10}Be interpretation), and that the topsoil is continually replacing carbon (enriched in radiocarbon) to the uppermost, eroding layers. As a result of this dynamic replacement of C, radiocarbon in the depositional layers is enriched relative to the ridgetop profile. If true, then the buried soil detected at 40 cm is not likely the original surface because the radiocarbon enrichment occurs below 40 cm and continues to a depth of 1 m. Alternatively, rooting or leaching processes could be enriching the carbon below that boundary.

The amount of C transferred through an eroding site by the processes of dynamic loss and replacement can be compared to the amount of Be being transferred by erosion and deposition. Manies et al. (2001) used local Iowa data for climate and cropping history to parameterize the Century model at the Treynor sites and produced estimates for storage and turnover of fast, slow, and very slow cycling carbon for the top 20 cm: Based on her study, the midslope position at the Treynor cropland likely started with $\sim 1.36 \text{ g/cm}^2$ of carbon in the upper 120 cm. The cropland site gained $\sim 5 \text{ g/cm}^2$ of carbon by net primary production (NPP) over 150 years and lost $\sim 2.5 \text{ g/cm}^2$ to harvest, $\sim 3.7 \text{ g/cm}^2$ to decomposition, and $\sim 0.5 \text{ g/cm}^2$ to erosion. The eroding site probably acted as a net source of CO_2 in its early history, and upon fertilizer inputs may have acted as a net sink for C (Harden et al. 1999), depending on whether the eroded material was protected from decomposition. Over the past 150 years therefore, $\sim 60\%$ of carbon exports were replaced (soil inputs 2.5 to soil losses of 4), whereas $\sim 3\%$ of ^{10}Be exports (1.9×10^8 in to 5.96×10^9 out) were replaced, representing a twenty-fold difference in the rate of replacement or cycling of these elements through the soil. Based on the redistribution of ^{10}Be , $\sim 12\%$ of carbon exports were lost to erosion in the past 30 years on the cropped midslope, the remainder being lost to decomposition. Moreover, that 12% has likely been exchanged through input and decomposition and mixed with bomb-spike radiocarbon. As a result, agricultural modification directly affects the areal distribution of both ^{10}Be and ^{14}C , but the dynamic exchange of C through soil results in a ubiquitous distribution of bomb-spike radiocarbon that masks the local and regional effects of erosion and deposition.

	0–20 cm turnover time
SOM1D	2–3 yr
SOM2C	20 yr
SOM3C	570 y

Conclusions

Many geologic studies have presented geomorphic and soil processes as separate events in time. For example, the term 'soil forming intervals' was applied by Morrison (1978) to sequences of deposits that were 'separated' by buried soils in semi-arid regions where shifts in climate were thought to be alternately conducive or hostile toward soil development. Subsequently, several workers (Birkeland and Shroba 1974; Marchand and Allwardt 1981; Chadwick and Davis 1990) clarified that rates of alluvial, eolian, and colluvial deposition varied by orders of magnitude over recent millennia, and that sedimentation rates, not pulses of warm or wet conditions, were key to the formation and preservation of soils. This study illustrates that soil-forming processes – in this case carbon turnover – proceed simultaneously with landform evolution – in this case erosion and deposition- and that geologic studies should incorporate both geomorphic and soil processes in descriptive or quantitative models. In answer to questions posed in the introduction:

1. Carbon cycling is impacted by landform processes that (a) bury and protect carbon from decomposition, and (b) expose carbon that was previously protected from decomposition. Be isotopes act as tracers to soil materials that were eroded in recent centuries during which time Be isotopes have not had time for significant radiodecay.
2. Loess accumulation rates were most rapid during the late Pleistocene and declined (3–5 fold) toward the present. The loess hills were likely forming during loess accumulation.
3. Organic matter in deep loess material likely originated from surface soil processes during loess accumulation. There is no strong evidence that the deep organic matter significantly pre-dates the loess deposition.
4. Agricultural modification greatly increased erosion rates in this region, resulting in dramatically altered soils, in some cases changing the soil order from Mollisol to Entisol.

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