

SNOWPACK DRIVEN CHANGES IN DECADAL SOIL
EVOLUTION: INSIGHTS FROM A 48-YEAR
SNOW MANIPULATION EXPERIMENT

by

Aaron Michael Feldhaus

A thesis submitted in partial fulfillment
of the requirements for the degree

of

Master of Science

in

Earth Sciences

MONTANA STATE UNIVERSITY
Bozeman, Montana

May 2017

©COPYRIGHT

by

Aaron Michael Feldhaus

2017

All Rights Reserved

DEDICATION

First and foremost, I would like to thank my advisor Dr. Jean Dixon for her guidance, insight, and patience during this project. I thank you for your help in my journey to become a better individual and academic, and it would not have been possible without our numerous thought-provoking discussions and your assistance throughout.

I would also like to thank my graduate committee members, Dr. Dave McWethy and Dr. Jack Brookshire, for their important input to this project and manuscript. I would also like to thank Dr. Tad Weaver for his thoughtful insight during fieldwork.

Many thanks to Dr. Jane Klassen and the staff in the Environmental Analysis Laboratory, as well as Rosie Wallander for their invaluable help with my analyses.

I would also like to thank the present and past members of the Earth Surface lab - Rachel Jensen, Sarah Benjaram, Colin Quinn, Grace Nicholas, and Natali Kragh, for their helpful feedback. Sarah, thanks especially for all the times you helped out in the lab.

I also would like to acknowledge my family and friends for their endless support during my graduate career. None of this would have been possible without your support, and I thank you.

TABLE OF CONTENTS

1. INTRODUCTION	1
2. APPROACH AND METHODS	4
Field Site.....	4
Sampling Methods.....	7
Physical and Chemical Soil Analyses	8
3. RESULTS	12
Particle Size Analysis	12
Carbon and Nitrogen Analysis	16
Fallout Radionuclide Profiles	17
Fallout Radionuclide Modeling.....	20
Fallout Radionuclide Variation among Snowpack Plots.....	23
pH Analysis	25
4. DISCUSSION	26
Snow Driven Changes in Surface Erosion	26
Snow Driven Changes in Bioturbation and Soil Mixing.....	30
Implications for Weathering Feedbacks	31
5. CONCLUSION.....	34
REFERENCES CITED.....	37
APPENDICES	44
APPENDIX A: Sampling Methods.....	45
APPENDIX B: Particle Size Analysis	47
APPENDIX C: Total Carbon and Nitrogen Analysis	54
APPENDIX D: Fallout Radionuclide Measurements and Method	59
APPENDIX E: Method for pH Analysis.....	74

LIST OF TABLES

Table	Page
1. Summative data of physical and chemical soil analyses	14
2. Summative Fallout Radionuclide analysis	22

LIST OF FIGURES

Figure	Page
1. Location and illustration of field area	6
2. Comparative particle size profiles	14
3. Detail particle size distributions	16
4. Total carbon and nitrogen depth profiles	18
5. Fallout radionuclide (FRN) profiles	20
6. Fallout radionuclide modeling.....	23
7. Fallout radionuclide inventories	25
8. Soil pH depth profiles.....	26

ABSTRACT

Soil mantled landscapes are a critical interface that support biological life, weather geologic materials, and develop in response to changes in climate. Climate has long been considered a dynamic control on the evolution of Earth's landscapes. However, we have limited understanding regarding how soils respond to short-term perturbations of key climate variables like precipitation and moisture availability. Furthermore, the timescales over which diverse weathering processes feedback and measurably change soil character are still relatively uncertain, as well as how they respond to swift changes in climate. Here, we explore the role of precipitation in decadal soil evolution by utilizing a 48-year snowpack experiment located in the Greater Yellowstone Ecosystem (GYE) of SW Montana. In this unique field site, we compare soil development across experimental plots with enhanced snowpack, where snow has been doubled (2x) and quadrupled (4x) above ambient conditions for almost five decades. We find that decadal snowpack addition provides multiple pathways for enhancing soil weathering, both physically and chemically. Soils under enhanced snowpack generally contain higher amounts of fine-grained material (clay and silt) and are more acidic (lower soil pH) in nature. Significant (>85%) surface depletions of the fallout radionuclide ^{210}Pb and reduced surface horizon carbon and nitrogen content, along with reduced above and below ground vegetation biomass provide evidence of increased wind erosion of soils that experience enhanced winter snowpack. Modeling of diffusion-like mixing from ^{210}Pb profiles also indicates there is increased bioturbation intensity (soil mixing) under enhanced snowpack. We find that snowpack addition, through associated changes in plant communities and vegetation biomass, along with its effects on physical and chemical weathering processes, produces rapid and measurable changes in the weathered state of soils. Our results indicate that short-term, decadal perturbations in snowpack significantly alter weathering mechanisms in this landscape, which measurably overprint thousands of years of soil development. These findings provide novel insight into the fundamental role of climate on short-term soil evolution and have significant implications for how mountainous or snowpack-dominated systems may respond to perturbations in climate.

INTRODUCTION

Soils are the dynamic interface between the atmosphere, geosphere, biosphere and hydrosphere. They support our habitable world and regulate the release of rock derived nutrients to ecosystems. Quantifying the timescales of and controls on soil evolution are critical for understanding the resilience of the soil system to perturbations (geologic, ecologic, climatic, or anthropogenic). Climate variables such as moisture availability play a critical role in soil chemical weathering and pedogenic processes in diverse systems (Chadwick et al., 1999; Dixon et al., 2016; Dixon et al., 2012; Rasmussen et al., 2011; Vitousek and Chadwick, 2013). Available water is necessary for many mineral weathering reactions that release base cations (Ca, Mg, Na, K) and plant available nutrients, and additionally transports the dissolved and weathered products down-gradient. Moisture availability also affects the rate at which water moves through a soil-mantled landscape. Fluid residence time in soils influence both the rate and intensity of mineral weathering (e.g., Maher, 2010, 2011), and likely the rate of soil production in a landscape. Importantly, it has been found that moisture availability, through precipitation via rainfall and snowpack, might drive non-linear pedogenic thresholds (Chadwick and Chorover, 2001; Vitousek and Chadwick, 2013). These potentially irreversible changes in physical soil character, cation concentrations, and nutrient dynamics have long been associated with distinct transitions in moisture availability.

Important feedbacks among chemical and physical weathering processes are important in soils, and can enhance the intensity of soil weathering in response to external forces such as climate (Dixon et al., 2009a). Links between these processes are

long recognized at both the basin scale (e.g., Millot et al., 2002; Stallard, 1985) and hillslope scale (e.g., Riebe et al., 2001). For example, bioturbation may influence both rates and extent of erosional and chemical weathering processes by the redistribution of soil, increasing soil porosity, and increasing mineral surface area by the physical disturbance of bedrock and saprolite (e.g., Dixon et al., 2009b; Gabet et al., 2003; Wilkinson et al., 2009). This physical disturbance by bioturbating organisms can be significant, and soils that have been extensively reworked and mixed may be more vulnerable to surface erosion, especially if small mounds of material are created through burrowing (Humphreys and Mitchell, 1983; Wilkinson et al., 2009). Furthermore, on hillslopes bioturbation has been noted to be a dominant player in sediment transport, where it can increase the mobility of soil and accelerate downslope movement of soil material (Black and Montgomery, 1991; Gabet et al., 2003).

In accordance with this line of thought, chemical fluxes have been found to increase with higher physical erosion rates globally and at a landscape scale. (e.g., Dixon and von Blanckenburg, 2012; Riebe et al., 2003; Riebe et al., 2001; West et al., 2005). Importantly, the interaction among physical and chemical processes overall work to create a positive feedback in soils that enhance its weathered state. These relationships have largely been quantified in landscapes assumed to be at steady-state, and the timescales over which physical and chemical processes feedback in perturbed systems are not clearly understood. For example, if key climate variables change rapidly, how might physical processes like bioturbation change and feedback with the soil chemical response? While changes in soil evolution are thought to occur over long timescales (i.e.,

millennia), little work has explored the degree to which rapid shifts (decadal) in soil characteristics are measurable and sustained. More work is needed to understand how both physical and chemical processes are connected in soil-mantled landscapes and how this dynamic system responds to changes in climate variables such as moisture availability and precipitation.

In mountainous systems, such as the Northern Rocky Mountains of North America, a key climate variable is snowpack. Winter precipitation dominates these regions and affects the timing of spring peak runoff (e.g., Pederson et al., 2011), the length of the growing season (e.g., Brookshire and Weaver, 2015), and the amount of water flux through soils. Little is known about how physical processes are impacted by changing snow conditions, and even less is understood about how changing this variable may enhance the weathered state of a soil by augmenting weathering feedbacks. To explore the relationship between snowpack and short-term soil evolution we utilize one of the longest continuously running snow manipulation experiments in the world (Weaver and Collins, 1977; Wipf and Rixen, 2010), a 48-year manipulation site located in the Northern Rocky Mountains and Greater Yellowstone Ecosystem (GYE) of southwest Montana.

APPROACH AND METHODS

Field Site

Our field area is located in the Bangtail Mountains of SW Montana (Figure 1), ~25 km from Bozeman, MT and within the Greater Yellowstone Ecosystem (GYE). We utilize a snowpack manipulation experiment initiated in 1968 within a subalpine grassland at 2330 m elevation. The site consists of 3 different snow treatment areas previously described in several papers (Brookshire and Weaver, 2015; Weaver, 1974; Weaver and Collins, 1977; Yano et al., 2015). An unmanipulated area (Control) maintains ambient winter snow conditions of approximately 0.6 m in depth, levels that are generally controlled by strong eastward wind patterns over the meadow (Yano et al., 2015). Two manipulated plots, where snow depths are experimentally doubled (SM2x) and quadrupled (SM4x), are maintained throughout the winter months by snow fence heights of 1.2 and 2.4 m, respectively. Each manipulated plot consists of 5 snow fences, 30 m in length, perpendicular to the dominant wind patterns that create large rectangular sections of enhanced winter snowpack with 2x and 4x ambient (Control) conditions. Continually in operation for 48 years, this is one of the longest running snow manipulation experiments globally (Wipf and Rixen, 2010).

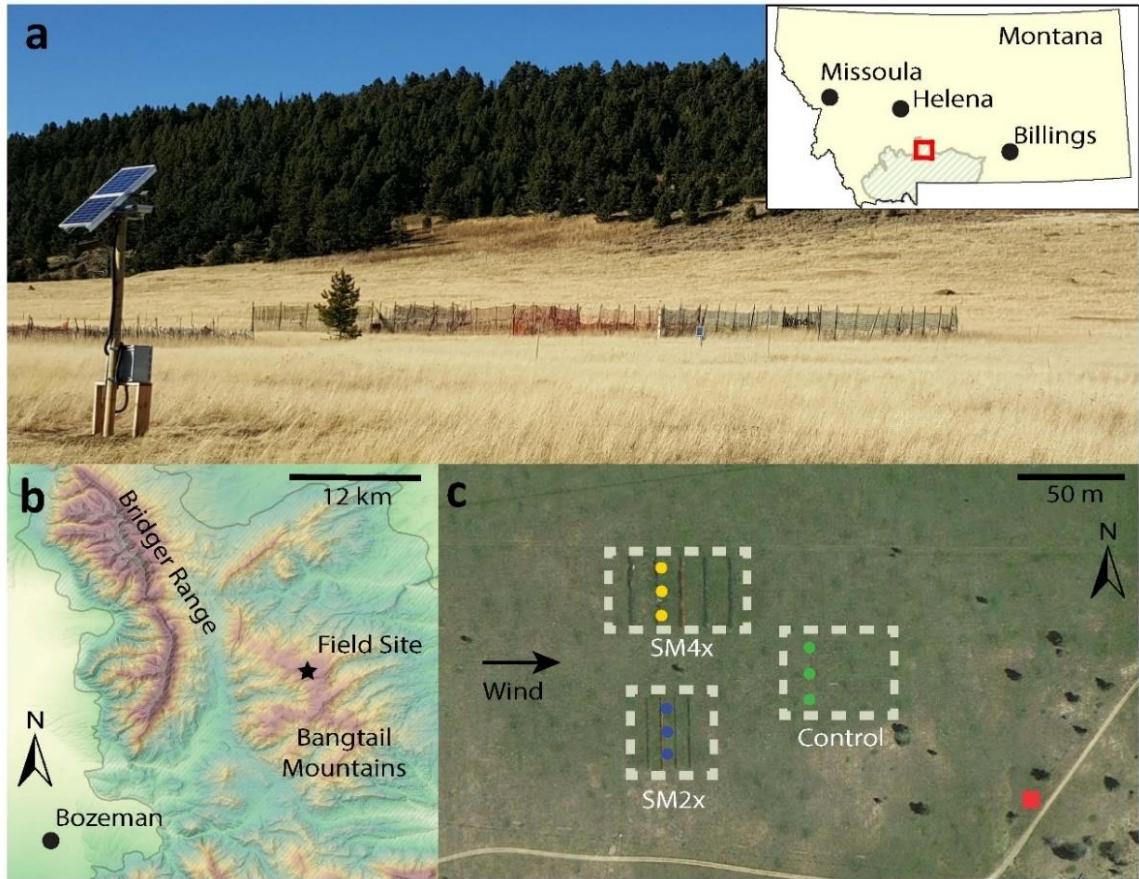


Figure 1. (a) Photo of study site and 48 year running snow manipulation experiment. Inset map shows the location within SW Montana and Greater Yellowstone Ecosystem (GYE; blue shaded region). Red square in the inset map shows the extent of panel (b). Our study site is within the Bangtail Mountains, ~25 km NE of Bozeman, MT at 2330 m elevation. (c) Snow manipulation plots (SM2x, SM4x) are maintained by 30m-long snow fences oriented perpendicular to the prevailing wind pattern over the meadow. Sampling locations, shown here by the colored circles, are color coded for each snow plot (green = control; blue = SM2x; yellow = SM4x). These color designations are consistent throughout this manuscript. The red square in the bottom right of the panel shows the viewpoint for the photograph shown in panel (a) looking NW towards snow fences.

Annual precipitation at this subalpine field site is ~ 960 mm and snow-dominated, with up to 87% falling as snow during September through May (Weaver and Collins, 1977). Annual to inter-annual precipitation is influenced by the intensity and variability of the El Niño–Southern Oscillation (ENSO), as well as multi-decadal variations in of the Pacific Decadal Oscillation (PDO) (McWethy et al., 2010), where warm phase ENSO events generally lead to less winter precipitation than cold phases. PDO phases generally lead to intensify or dampen the effects of ENSO cycles over decadal timescales in the Northern Rockies. These atmospheric circulation patterns along with topographic effects on air masses control the amount of snowfall the field site receives. Regionally the GYE and Rocky Mountains have witnessed unprecedented declines in snowpack over the last century, making rapid changes in snowpack a focus of research and management (Pederson et al., 2011). Despite regional reductions in snowpack, average winter precipitation and SWE has been relatively constant across our study site over the 48-year study period (Brookshire and Weaver, 2015), which contrasts to trends in the greater Northern Rockies and suggests natural variability in atmospheric circulation and topography have been dominant controls throughout our study period.

Soils present at this field site are classified as Typic Cryoborolls (Weaver, 1974), and are formed from an andesitic sandstone from the Upper Cretaceous Fort Union Formation (Skipp et al., 2004). The Bangtail Mountains exhibit a tectonic domain indicative of Laramide-style uplift and compression, and the region occupies the transition zone between the Rocky Mountain and Great Plain physiographic provinces (Locke et al., 1989). Unlike the Yellowstone Plateau and neighboring Bridger Range,

maximum elevations of the range were not sufficient for glaciation during the last known regional glacial cycles, but most likely caused periglacial conditions as recently as 15ka (Pierce, 1982). This makes exact dating of the soil difficult, but here we assume the soil is relatively young, either directly post-glacial (less than 15ka in age), or slightly older. These soils integrate climate and weathering processes over these timescales and experienced multiple changes in climate throughout the Holocene, including periglacial conditions, increased mid-Holocene wetness, and semi-arid conditions in the present day.

Previous studies at this experimental site have primarily focused on ecological changes associated with snowpack manipulation, and have found snowpack variability significantly alters grassland vegetation regimes, plant biomass, and primary productivity (Weaver, 1974; Weaver and Collins, 1977). Enhanced snowpack treatments have also shown to increase leaching of bioavailable nutrients such as nitrate and phosphate (Yano et al., 2015). Observed reductions in natural vegetation, and concurrent increases in summer drying over the study period have led to an large decline in above-ground net primary productivity (ANPP) of the meadow (Brookshire and Weaver, 2015).

Sampling Methods

During October of 2015, nine soil cores were collected from the field site. Three boring locations, approximately 10 m apart, were chosen within each snow depth treatment (Figure 1). Each soil core was collected using a 15cm x 5cm diameter split-core and slide hammer; two successive cores (each) were collected from each sample location to a total depth of 35 cm. This depth corresponds to the thickness of unconsolidated,

bioturbated soil at this site, as saprolite (weathered rock containing relict rock structure) was observed in the bottommost centimeter of several cores). Samples were taken from the field site and brought back to the Department of Earth Sciences at Montana State University, where they were sectioned into 1.5-2.5 cm increments for analysis. Large organics were removed and the incremented samples were dried for 48 hours at 90 °C.

Physical and Chemical Soil Analyses

We measured a suite of physical and chemical parameters to understand how snowpack variability affects soil evolution. Biophysical soil mixing intensities, particle size, total C and N, and soil pH analyses were conducted at Montana State University. We examined biophysical soil mixing using activity profiles of fallout radionuclides (FRNs) ^{137}Cs and ^{210}Pb -excess, following methods highlighted by Dixon et al. (2009b) and Kaste et al. (2006). ^{137}Cs deposition to soils resulted from thermonuclear testing during the Cold War, with a peak deposition input in 1963 (Wright et al., 1999). This nuclide has since shown to be an accurate indicator for short-term soil movement (Kaste et al., 2006; Mabit et al., 2008; Mabit et al., 2013; Walling, 1998). Atmospherically delivered ^{210}Pb , or ^{210}Pb -excess (“unsupported ^{210}Pb ”; $t_{1/2} = 22$ years), is part of the ^{238}U decay series, and the decay product of ^{222}Rn not in chemical equilibrium with the parent ^{226}Ra . It is continually deposited to the land surface via precipitation (Baskaran, 2011; He and Walling, 1996; Wallbrink and Murray, 1996). ^{210}Pb and ^{137}Cs have high affinities for clay particles upon deposition, and are mobilized with physical transport of soil.

Therefore, they are useful tools for understanding soil mixing and sediment transport processes in a landscape (e.g., Mabit et al., 2008).

Nuclide activities were measured by gamma spectroscopy using a broad energy Germanium Detector (BEGe) in the Fallout Radionuclide Laboratory at MSU. Each of 185 samples were sieved to <2mm fraction (nearly all samples contained no coarse grains) and ~25g was packed into small plastic petri dishes (60x15mm) and sealed in paraffin wax. Samples were then left for >15 days to allow secular equilibrium for ^{222}Rn (^{210}Pb parent, with $t_{1/2}=3.8$ days) before loading for nuclide analysis. ^{210}Pb -excess was calculated by subtracting the measured ^{226}Ra in a soil sample, under the assumption that ^{210}Pb formed in soils (^{210}Pb -supported) is in secular equilibrium with ^{226}Ra . We also calculated ^{210}Pb -excess via an alternate method, assuming a single value of “supported” ^{210}Pb is consistent throughout each profile. This threshold value was chosen as the value where no significant changes in total ^{210}Pb occurred with depth, and roughly corresponds with the detection limits of the nuclide. Detailed methods and techniques regarding FRN analysis are provided in Appendix D.

We used FRN depth profiles to provide insight into the intensity of soil mixing by applying a steady-state advection-diffusion model (Eq. 1). This model quantifies relative differences in soil mixing between study sites (DeMaster and Cochran, 1982; Dixon et al., 2009b; Kaste et al., 2007). Here, the ^{210}Pb -excess activity at depth is a function of the surface nuclide activity, advection velocity, and a diffusion coefficient:

$$A(z) = A_0 \left(\frac{V - \sqrt{V^2 + 4\lambda D}}{2D} (z) \right) \quad (\text{Equation 1})$$

Where:

$A(z)$ = nuclide activity at depth

A_0 = surface nuclide activity
 V = advection velocity
 λ = nuclide decay
 D = diffusion coefficient
 z = depth

We choose steady-state conditions for the ease of modeling and comparison across sites. However, steady-state conditions require roughly 3-5 half-lives of consistent conditions to reach, and therefore the manipulated plots have likely not reached these conditions under the experimental changes ~50 years ago. Recognizing these limitations, we use this advection-diffusion model to provide relative differences in soil mixing intensity as affected by snowpack addition.

Following fallout radionuclide analyses, one representative core from each snow plot area was chosen for further physical and chemical analyses and these samples were removed from their wax-sealed casings. Particle Size analysis was conducted using a Malvern Mastersizer 3000 Particle Size Analyzer in the Environmental Analysis Laboratory at Montana State University. Samples were mixed into a slurry of sodium hexametaphosphate to deflocculate and suspend sediments prior to instrument analysis. The particle size spectrum of .01-10,000 μm was then measured by laser diffraction and values were reported in volume percent. Methods are detailed in Appendix B.

The same three soil cores were also used for soil C, N, and pH analyses. Samples were homogenized and pulverized over 12 hours using a steel jar mill, and 50 mg of each was subsampled for C and N analysis. Samples were pressed into tin capsules (3.5x5mm), sealed for analysis, and measured for total C and N at the Environmental Analysis Laboratory (EAL) at MSU using a Costech Instruments Elemental Combustion

System. Soil pH was measured using unpulverized soil material from each sample in a 1:2 soil DI water mixture and an electronic pH probe.

RESULTS

Particle Size Analysis

Data from particle size analysis show quantitative differences in fine-grain soils across our snowpack treatment plots. Soils across all sites contain low clay content (0-3.9 μm), ranging from 0.1-9.0% by volume (Table 1). Clay content of snow treatment plots are slightly higher than the control plot, with average clay fractions of $2.4 \pm 1.1\%$, $3.9 \pm 1.7\%$, and $3.1 \pm 2.0\%$ for the control, SM2x, and SM4x plots respectively ($\mu \pm \text{s.d.}$, $n=24$ for each plot; Table 1). Volume percentages of both clay and silt increase with depth in all profiles, accompanied by decreases in sand-sized fractions. This increase in clay with depth is most apparent in snow-manipulated SM2x and SM4x plots, where clay fractions at 35cm depth are roughly double those in the control plot (Figure 2). Silt fractions show the highest range in the control plot, increasing with depth from 5.4-49.9%, compared to 21.3- 56.6%, and 21.1-61.9% in the SM2x and SM4x plots respectively. Sand-size fractions decrease steadily with depth across all experimental sites, with similar values at the bottom of the profiles.

		% Clay	% Silt	% Sand	% Carbon	% Nitrogen	soil pH
Control	$\mu \pm \text{s.d}$ (range)	2.4 ± 1.1 (1.1-4.8)	33.7 ± 12.1 (5.4-49.9)	63.8 ± 12.8 (45.3-92.9)	3.51 ± 3.94 (1.50-18.87)	0.29 ± 0.25 (0.14-1.27)	5.63 ± 0.07 (5.49-5.70)
SM2x	$\mu \pm \text{s.d}$ (range)	3.9 ± 1.7 (1.2-7.0)	41.1 ± 10.4 (21.3-56.6)	54.9 ± 11.9 (37.7-77.4)	2.49 ± 0.87 (1.26-4.03)	0.23 ± 0.07 (0.12-0.34)	5.52 ± 0.10 (5.36-5.62)
SM4x	$\mu \pm \text{s.d}$ (range)	3.7 ± 2.0 (1.2-9.0)	37.6 ± 11.5 (21.1-61.9)	58.6 ± 13.2 (29.1-77.1)	2.45 ± 0.83 (1.02-4.17)	0.21 ± 0.06 (0.10-0.33)	5.49 ± 0.19 (5.27-5.75)

Table 1. Summary soil particle size, C, N, and pH data. Mean and standard deviations for each soil plot reflect n=24 particle size values (as % volume), n=17 soil C and N values (as % mass), and n=6 soil pH values.

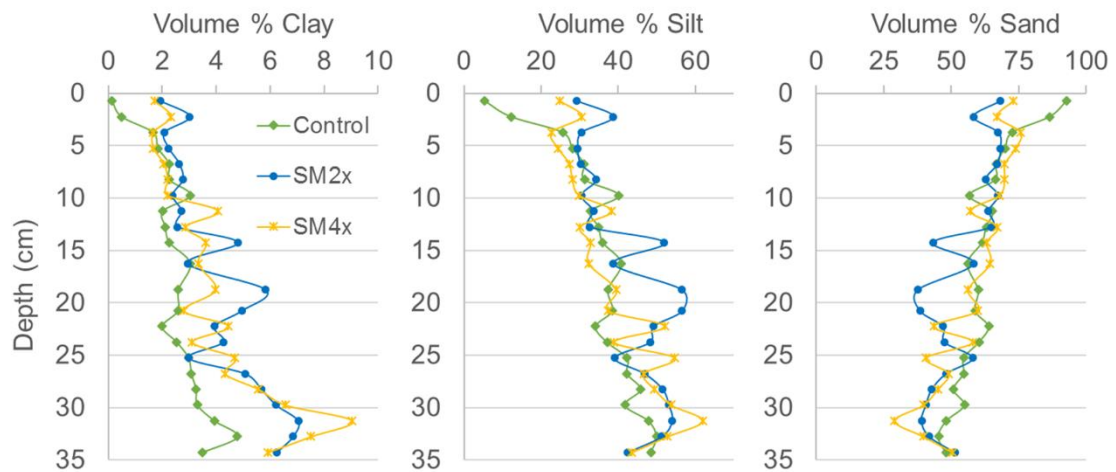


Figure 2. Comparative particle size profiles of clay, silt and sand with depth. Depth trends among snow plots are generally similar for each of the three size categories, though mean clay fractions in the snow manipulation plots are higher than the control profile, especially at depth.

Depth profiles and average abundance of major size fractions highlight differences in soil texture across the experimental snow plots; however, detailed particle size distributions (Figure 3) provide a more comprehensive view and further insight into particle size dynamics and snowpack influence. Particle size analysis results are represented by individual size distribution curve for each sample within a depth profile. Together, these curves highlight particle size differences and depth trends independent of classification system or binning thresholds that distinguish size classes. Soils from the three snow treatments display general trend of decreasing modal (peak) particle size values with depth. Surface (depth=0) distributions are consistently unimodal in character, with more bimodal distributions with increasing depth. The change from more normal to bimodal distributions appears to be augmented by snowpack treatment, as the control plot contains the least bimodal particle size distributions. As snowpack increases, we also see an increasing skew towards finer grained material.

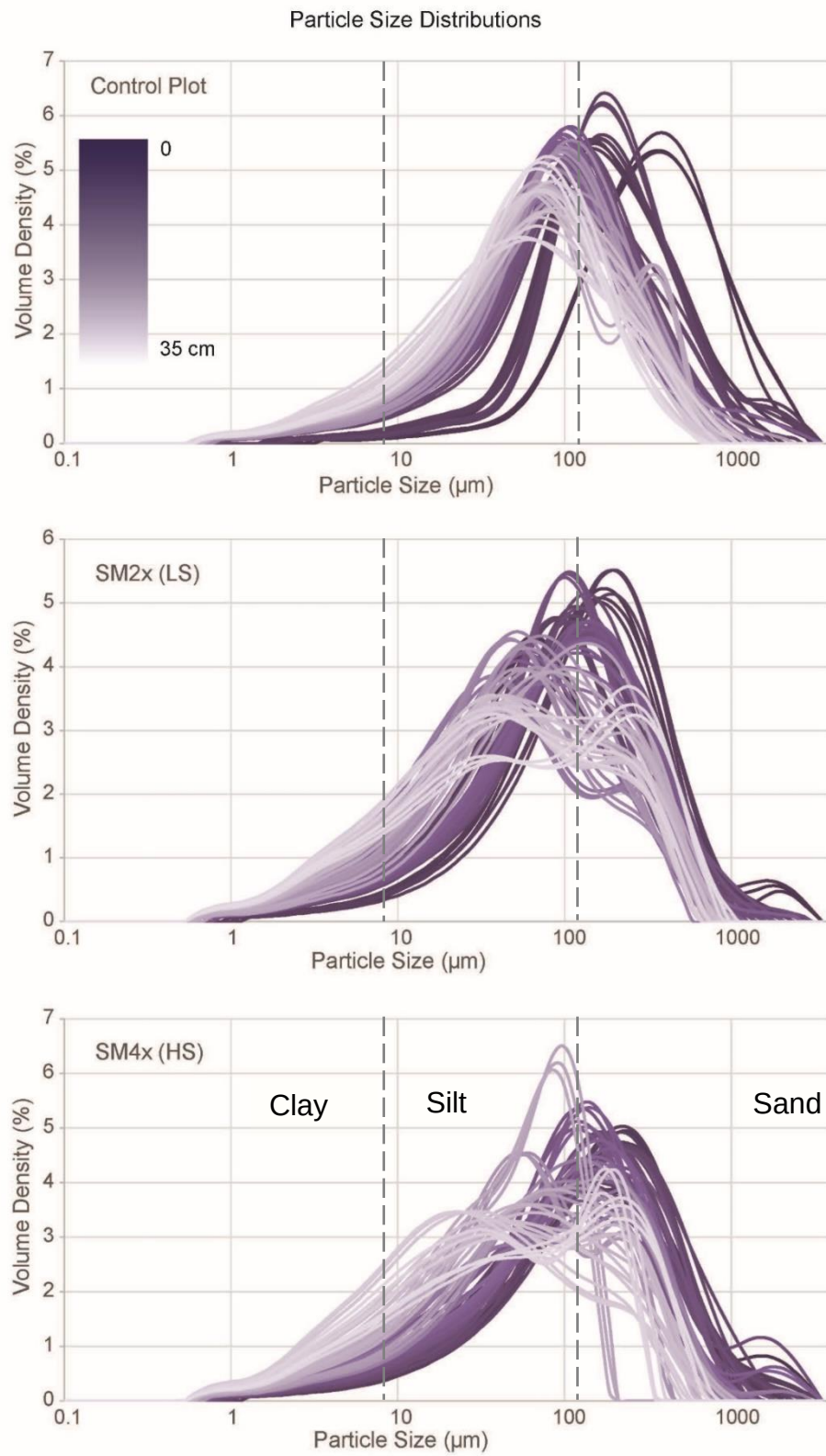


Figure 3. (description on following page)

Figure 3. (previous page) Soil particle size distributions with depth in control, SM2x, and SM4x treatments. Particle size is displayed with a logarithmic axis to highlight changes in fine fractions. Each sample is represented by three replicate distribution curves, and curves are color-graded from dark (surface) to light (maximum depth). We find decreasing particle size with depth, characterized by a shift from unimodal histogram peaks in surface samples to more bimodal distributions at depth that are finer grained in nature. We also note higher volume percentages of fine silts and clays at SM2x and SM4x plots compared to the control. Median size fractions (D_{x50}) are highest in the control (110.9 μm), and lower in SM2x and SM4x plots (D_{x50}= 99.3 and 82.9 μm respectively).

Carbon and Nitrogen Analysis

Analyses of total Carbon and Nitrogen in soils show that both decrease with depth across all sites. This decrease is roughly consistent in magnitude across all plots, with the clear exception of values in uppermost soil horizons (Figure 4). We find a striking 3-fold higher amount of total C in the top 3 cm of the control profile compared to the same depth interval of manipulated plots, and a similar relationship for bulk N is found. This surface difference results in higher average values and greater variability for C and N in the control plots ($3.51 \pm 3.94\%$ and $0.29 \pm 0.25\%$ respectively) compared to snow-manipulated plots (combined snow manipulation plots show $2.47 \pm 0.85\%$ C and $0.22 \pm 0.06\%$ N). SM2x and SM4x plots exhibit similar average of bulk C and N or roughly 70% and 75% the average control C and N content. Total C and N levels in all experimental plots decrease to below 1.50 and 0.14%, respectively, at maximum core depths of 35 cm.

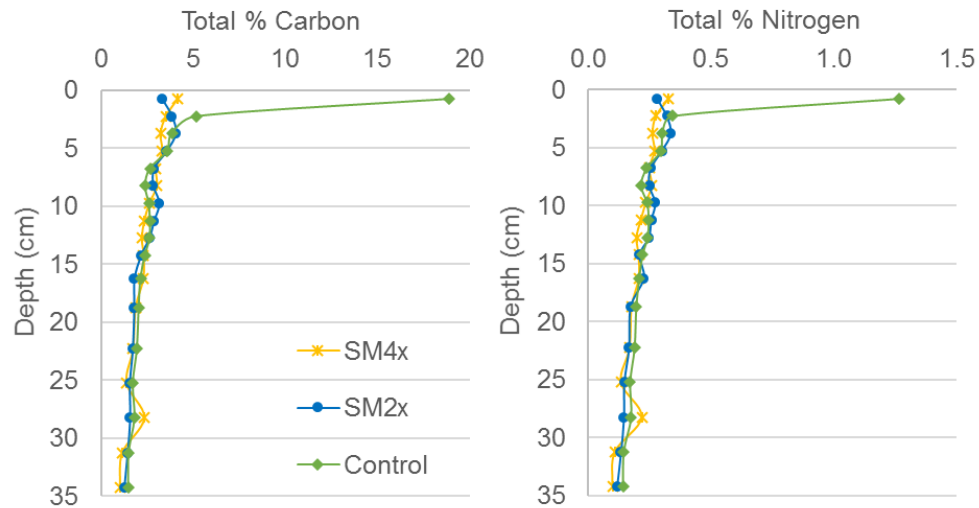


Figure 4. Depth profiles of bulk carbon and nitrogen. Surface soils from the control profile show C and N compared to snow manipulated soils. C and N decrease steadily below 3 cm depth in all snow depth areas.

Fallout Radionuclide Profiles

Distribution of fallout nuclides within the soil profiles show significant differences in profile shape and inventory across experimental sites (Figure 5). We exclude from our analysis one soil core from the control site, C1, which showed evidence of recent human disturbance and surface burial based on ^{210}Pb -excess data. See Appendix E for detailed description. The control plot generally displays profiles with higher activities than experimental sites. Both C2 and C3 have the highest ^{210}Pb -excess activities near the surface as shown in panels (b) and (c) of Figure 5. The control profile C2 is the only core that contains a defined ^{137}Cs peak (panel (a) of Figure 5). Both SM2x and SM4x show profile shapes that are more vertically homogenous and less variable in nature, with slightly higher activities shown at depth. Moreover, the penetration depth of

^{210}Pb -excess increases by roughly 4 cm from ambient to SM4x conditions, with the highest under SM4x conditions (7.25 ± 0.71 cm, Table 2). In general for all core profiles we see a decrease in activity with depth, with no measurable ^{137}Cs and ^{210}Pb -excess beneath a depth of 27 cm.

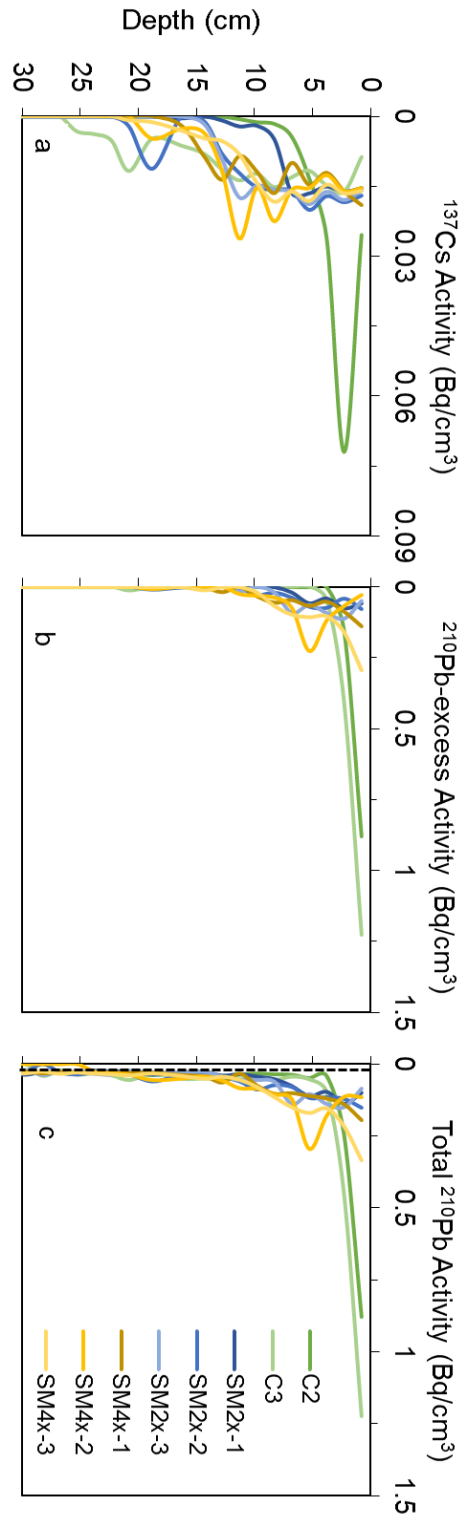


Figure 5. (Description on following page)

Figure 5. (previous page) Fallout radionuclide profiles in soils across the control, SM2x, and SM4x experimental plots. Higher total ^{210}Pb inventories (Bq/cm^2) in the control plot (Figure 7) are represented here in panels (b) and (c) by larger subsurface peaks in activity (Bq/cm^3). Black dashed line represents our threshold value of 0.04 in total ^{210}Pb plot of panel (c), which provides ^{210}Pb -excess (alt) values. Depth of nuclide distribution provides insight into the intensity of soil mixing and diffusion-like processes. Cores SM4x-2 and SM2x-2 have clear ^{137}Cs subsurface peaks at depths of 12 and 18 cm, respectively.

Fallout Radionuclide Modeling

Results from a steady-state advection-diffusion model (Eq. 1) show distinct differences among relative diffusion-style soil mixing rates across the experimental plots. We find a large range in diffusional-mixing values among our snow plots (Table 2), with SM4x plots exhibiting the highest average diffusion ($4.24 \pm 3.43 \text{ cm}^2/\text{yr}$). In general, diffusion coefficients from soils in the snow addition plots are nearly two orders of magnitude higher than those in the control plots, and average diffusion values systemically increase from $0.06 - 4.24 \text{ cm}^2/\text{yr}$ with increased snowpack conditions. We note that high diffusion values are associated with profile shapes that do not have large surface peaks (Figure 6), and calculated values steadily increase as a profile become more vertically homogenous. We also find that small changes in subsurface redistribution can have substantial effects on relative diffusion values, making this model sensitive to changes in bioturbation intensity.

FRN Inventories and Modeling					
Sample	¹³⁷ Cs	²¹⁰ Pb-excess		²¹⁰ Pb-excess (alt)	Diffusional mixing (D)
	Inventory (Bq/cm ²)	Inventory (Bq/cm ²)	95% Inv. Depth (cm)	Inventory (Bq/cm ²)	(cm ² /yr)
C2	0.23	1.70	2.25	1.94	0.07
C3	0.26	2.82	2.25	3.25	0.05
SM2x-1	0.14	0.14	5.25	0.67	2.67
SM2x-2	0.26	0.19	6.75	1.07	1.18
SM2x-3	0.23	0.44	6.75	0.94	3.63
SM4x-1	0.21	0.39	8.25	1.12	2.64
SM4x-2	0.26	0.57	6.75	1.58	9.01
SM4x-3	0.22	1.03	6.75	1.82	1.08
Plot Avg.	$\mu \pm \text{s.d}$	$\mu \pm \text{s.d}$	$\mu \pm \text{s.d}$	$\mu \pm \text{s.d}$	$\mu \pm \text{s.d}$
Control	0.24 $\pm$ 0.02	2.26 $\pm$ 0.56	3.75 $\pm$ 1.41	2.60 $\pm$ 0.66	0.06 $\pm$ 0.01
SM2x	0.21 $\pm$ 0.05	0.26 $\pm$ 0.13	6.25 $\pm$ 0.71	0.89 $\pm$ 0.17	2.49 $\pm$ 1.01
SM4x	0.23 $\pm$ 0.02	0.66 $\pm$ 0.27	7.25 $\pm$ 0.71	1.51 $\pm$ 0.29	4.24 $\pm$ 3.43

Table 2. Summative table of FRN analysis. Note nuclide depletion of both ²¹⁰Pb-excess metrics under snowpack addition, however little change is found in ¹³⁷Cs inventories. Penetration depths discussed above are where 95% of nuclide inventory is measured. Diffusion values increase systematically with snowpack addition.

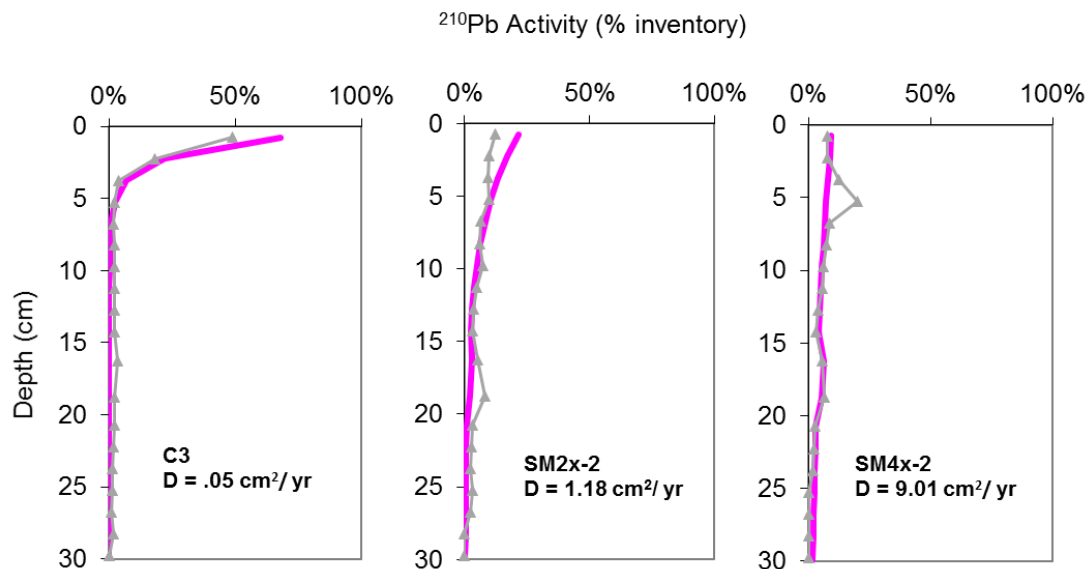


Figure 6. Measured ^{210}Pb -excess inventories and best-fit model data from Equation 1 for the control plot. Since clear subsurface peaks were indiscernible in all but one soil profile, we assumed advection was minimal and set $(V) = 0$ as the model parameter. Best-fit data was found by reducing the sum of residuals, and calculated diffusion (D) values for all profiles were found.

Fallout Radionuclide Variation among Snowpack Plots

Fallout Radionuclide (FRN) data consist of over 185 soil samples analyzed for their ^{137}Cs and ^{210}Pb -excess activities, and represent over 7 months of continuous data collection via gamma ray spectroscopy. Several important observations emerge from our analysis of these data. We find significant differences in ^{210}Pb -excess nuclide inventories between experimental plots (Figure 7). Inventories of ^{210}Pb -excess (calculated by two different methods) are highest in the control plot, with ^{210}Pb -excess and ^{210}Pb -excess (alt) average inventory values of 2.26 ± 0.56 and $2.60 \pm 0.66 \text{ Bq/cm}^2$, respectively. These values are significantly higher ($p > .05$, unpaired t-test) than inventories under SM2x and SM4x conditions (Table 2; Figure 7 b,c). The SM2x plot contains the lowest inventories at our field site for both ^{210}Pb -excess metrics. While we find large differences in ^{210}Pb -excess inventories, ^{137}Cs inventories are similar across all experimental plots (panel (a), Figure 7), with no statistically significant differences found over 48 years of snowpack manipulation.

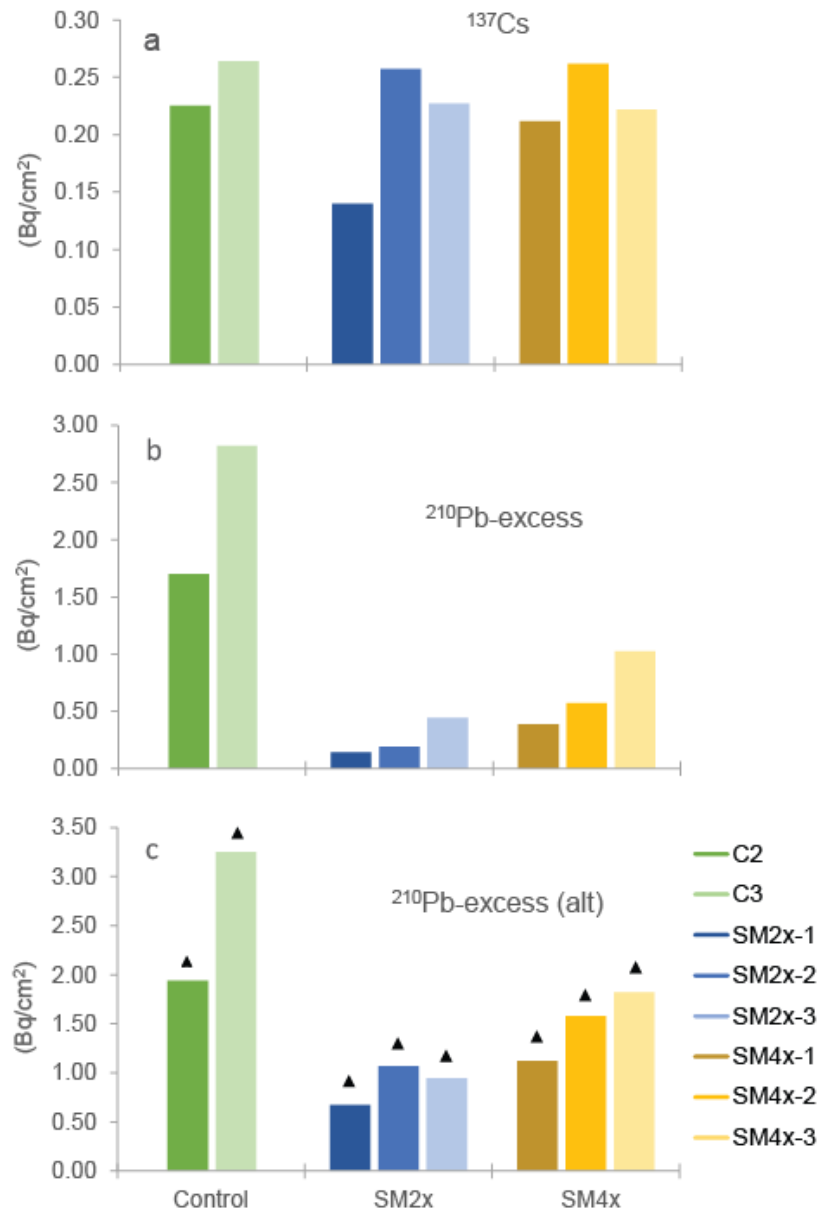


Figure 8. Fallout nuclide inventories in soils. Inventories represent the summed activity across individual soil profiles. Panel (a) Inventories of ¹³⁷Cs show relatively little changes with snowpack addition. Data show higher inventories of both (b) ²¹⁰Pb-excess and (c) ²¹⁰Pb-excess (alt) in control compared to soils under snow manipulation. ²¹⁰Pb-excess inventories are calculated using a per-sample correction for supported ²¹⁰Pb (see methods and appendix E), while ²¹⁰Pb-excess (alt) inventories are calculated assuming a single value for supported ²¹⁰Pb throughout the profile (0.04 Bq/cm²). Black triangles represent inventory values if a threshold value of 0.03 Bq/cm² is implemented. All ²¹⁰Pb-excess data, regardless of assumptions used, are consistent in suggesting nuclide depletion in snow manipulated sites compared to the control.

pH Analysis

Soil pH generally decreases with snowpack addition (Figure 8), and these differences are most apparent in the upper 20 cm of soils. Soil pH in the control plot is roughly uniform with depth (ranging from 5.49-5.70), with an average soil pH of 5.63 ± 0.07 ($\mu \pm \text{sd}$). Average soil pH is both more highly variably and lower on average with snowpack addition (Table 1), with SM4x plot soils averaging pH values of 5.49 ± 0.19 . Although upper soil horizons show more variable pH values across plots, values for soils across all experimental plots converge to roughly 5.6 pH at depths greater than 25 cm.

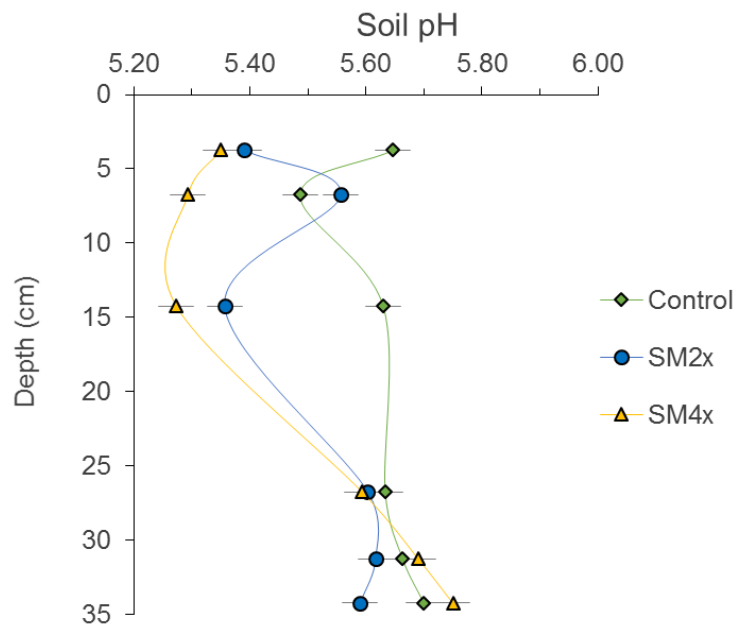


Figure 8. Soil pH-depth profiles show distinct changes in pH values with depth and snowpack addition. The control plot exhibits the highest average soil pH ($5.63 \pm .07$, $n=6$), whereas the SM4x contains the lowest ($5.49 \pm .19$, $n=6$). Error bars represent an uncertainty of 0.03 pH.

DISCUSSION

Snow Driven Changes in Surface Erosion

We find multiple lines of evidence that suggest 48 years of snowpack manipulation are accompanied by distinct changes in soil erosion and weathering processes, both physical and chemical in nature. Importantly, we find snowpack addition results in measurable depletions in ^{210}Pb -excess, without significant changes in ^{137}Cs inventories (Figure 7). This discovery is initially surprising considering ^{137}Cs and ^{210}Pb -excess have similar chemical behaviors in soil.

While fallout nuclide depositional fluxes are strongly correlated with the amount of local precipitation (Baskaran, 2011), we would not expect to see changes in ^{137}Cs with increased experimental snowpack given that it was largely deposited prior to snowpack manipulation. However, we would expect inventories of ^{210}Pb -excess, which is continuously deposited on the surface, to increase proportionally with snowpack addition, as long as there is no significant loss via leaching, desorption from mineral or organic surface area, or surface erosion. Little is known about the ^{210}Pb -excess scavenging efficiencies between rain and snow and few studies have quantified this relationship, but if we conservatively assume between 40-80% of annual precipitation is snow at this field site, then under 2x and 4x ambient snow conditions we would predict an 80-160% and 160-320% increase in ^{210}Pb -excess inventories at SM2x and SM4x sites. However, the observations of ^{210}Pb -excess surface depletion and lower inventories with snowpack addition is not consistent with the above expectation.

What mechanisms could cause the nuclide loss we measure here under enhanced snowpack? The fallout radionuclides (FRN's) used in this study attach strongly to clay minerals and organic matter within the soils. Under normal soil pH and moisture conditions, these two nuclides do not desorb from available surfaces, and to do so would require much more acidic ($\text{pH} < 4$), or highly basic ($\text{pH} > 9$) soil conditions (Rhodes, 1957). Therefore, desorption and subsequent leaching losses are not likely to have occurred during the study period, and measured losses of radionuclides should therefore be accompanied by losses of either clay minerals or organic matter these nuclides are sorbed to. Moreover, if leaching or downward transport out of the soil profile (either dissolved, colloidal, or illuviated clays) were the main driving mechanism for FRN loss, we would expect to see similar losses measured in both ^{137}Cs and ^{210}Pb -excess, yet we only observe changes in ^{210}Pb -excess. This suggests that there is minimal nuclide desorption, and that downward loss of nuclide-sorbed fine-grained material is unlikely.

Surface erosion may instead provide a mechanism for removal of ^{210}Pb -excess. Erosion by overland flow has previously been invoked to explain nuclide depletion in soil profiles (e.g., Dixon et al., 2009b; Kaste et al., 2007; Kaste et al., 2006; O'farrell et al., 2007). Though possible here, this mechanism is unlikely at this field site given its negligible (near zero) slope gradient and absence of topographic indicators of overland flow, including rill marks from late spring snowmelt or summer rainfall events. Furthermore, prior field observations at this site indicate that meltwaters largely percolate downward through the soil column without causing significant surface flow or erosion from the snow plots (Yano et al., 2015).

Wind erosion of fine soils and organic matter may be a more likely explanation for inventory losses of ^{210}Pb -excess. Previous work at this site documented an over 30% reduction in aboveground biomass driven by snowpack addition (Holsinger, 2013). Here, observed reductions in above and belowground biomass and increased bare ground exposure were attributed to the shift from tall, lignacious C3 bunch grasses (Idaho Fescue) to smaller annual forb species (Holsinger, 2013). Aboveground biomass reduction is linked to decreased soil cover in other grassland settings (e.g., Naudin et al., 2012). The observed vegetation regime shifts and subsequent exposure of surface soils experiencing enhanced snow conditions increases the likelihood of surface wind erosion. As vegetation cover is reduced, topsoil cohesion is reduced and surface soil particles containing ^{210}Pb -excess can be lost. More broadly, wind erosion is widely recognized to increase with decreased vegetation cover due to short-term aridity changes (Fryrear, 1985; Gill, 1996; Neal and Motts, 1967; Okin et al., 2006). For example, Munson et al. (2011) found increases in aeolian dust emissions from decreasing perennial vegetation cover on the Colorado Plateau. Elmore et al. (2008) documented measurable changes in nuclide inventories due to decreased vegetation cover in Owens Valley, California, and Mabit et al. (2008) note the importance of wind erosion on nuclide inventory studies. Our findings suggest wind erosion that has been caused by reduced vegetative soil cover is a primary mechanism for ^{210}Pb -excess inventory depletion at this field site.

Similar behavior but different delivery mechanisms can help to explain why ^{210}Pb -excess inventories are likely more vulnerable to wind erosion than ^{137}Cs . ^{210}Pb -excess is constantly supplied to the ground surface via meteoric deposition, and highest

inventories are often found in the top few centimeters of a soil profile (Walling, 1998). Therefore, ^{210}Pb -excess peaks at the surface of soils, where it is most sensitive to surface removal by wind erosion. ^{137}Cs , on the other hand, was a single input prior to snowpack manipulation and subsequent vegetation changes at this subalpine grassland. It is possible then that advection and diffusional processes prior to loss of soil cover may have protected ^{137}Cs from depletion. This same phenomenon has been documented in soil cores near Exeter, UK, where ^{210}Pb -excess inventories were preferentially depleted over ^{137}Cs by erosional redistribution (Walling, 1998). Furthermore, ^{210}Pb -excess is more sensitive to changes in organic matter due to a slightly higher adsorption preference than ^{137}Cs (Dörr and Münnich, 1989). We suggest that the reduction in vegetation biomass and the increased likelihood for wind-driven soil and litterfall loss is responsible for ^{210}Pb -excess loss in enhanced snow plots. The reduction in vegetation biomass and increased chance for surface litterfall may also contribute to the measured reductions in C and N in this study, and previous work conducted at this field site has found similar changes in both C and N under snow manipulation (Yano et al., 2015). Surface removal and reduced organic matter alone may not entirely describe the significant losses of these elements under enhanced snowpack measured in both studies, and suggests more than one pathway is responsible for this change.

Despite significant reductions in ^{210}Pb -excess inventories under enhanced snow, we find more clay-sized particles as well. Given that this nuclide sorbs strongly to clay minerals and organic matter in the subsurface, we might suggest there could be higher inventories under enhanced snowpack. There have been measured increases in clay sized

particles under SM2x and SM4x conditions, however this change is mostly apparent at depths (Figure 2) deeper than measured ^{210}Pb -excess penetration. Though there may be translocation of fine-grained material here, if illuviation of this fine-grained material was occurring from near the surface and transporting to depth, then we would expect that higher amounts of fine-grained material would be accompanied by higher nuclide activities. Instead, these results suggest the higher levels of silt and clay under SM2x and SM4x conditions are likely produced at depth by either chemical or physical weathering and therefore not affecting ^{210}Pb -excess and ^{137}Cs dynamics.

Snow Driven Changes in Bioturbation and Soil Mixing

While surface removal can explain disparate patterns in nuclide inventories across our study site, physical weathering may also be enhanced by increased snowpack. Our steady-state model, although limited in its assumptions, provides relative differences in physical soil mixing and previous work has found that increased diffusive soil mixing in soils correlates with increased bioturbation intensity (Dixon et al., 2009b). Weaver and Collins (1977) noted that gopher bioturbation intensity increased in soils with enhanced snow and this higher burrowing activity has remained throughout the experimental period (Weaver, personal communication). These observations are consistent with our modeled diffusion rates, and indicate that soil mixing, via bioturbation, increases with snowpack manipulation. This increase may be correlated with increased thermal buffering of habitat during the coldest winter months (Thorn, 1978; Yano et al., 2015). The soils present under ambient snowpack depths freeze periodically during winter, whereas under

enhanced snowpack it has been documented that soils remain roughly isothermal and protected from sub-freezing temperatures. This insulation may provide a more suitable winter habitat for the dominant bioturbator in this system: the American pocket gopher (Hansen and Morris, 1968).

At the same time wind erosion at the surface depletes ^{210}Pb -excess, bioturbation processes redistribute remaining nuclides to depth, and we find the highest average penetration of ^{210}Pb -excess under increased snowpack (Table 2). We also find few undisturbed subsurface peaks in ^{137}Cs , especially under SM2x and SM4x conditions, indicating the diffusional soil mixing by bioturbation is pervasive and may dominate over advection processes in these soils.

Implications for Weathering Feedbacks

The enhanced bioturbation and mixing under deeper snowpack indicated by this study has several important implications for soil weathering in this system, both physical and chemical in nature. Snow addition appears to provide multiple pathways that contribute to more weathered soils in this mountainous grassland. Bioturbation can increase chemical weathering by creating fresh mineral surfaces, and increasing both porosity and water flux in soils (Gabet et al., 2003; Kelly et al., 1998; Wilkinson et al., 2009). Moreover, increased moisture availability (i.e. enhanced snowpack here) has been found to amplify chemical weathering in soil-mantled landscapes (Chadwick and Chorover, 2001; Dixon et al., 2016; Dixon et al., 2009a; Maher, 2010; Rasmussen et al., 2011; Vitousek and Chadwick, 2013; White et al., 1999). The physical changes in grain

size measured in this study, and higher amounts of fine-grained material under manipulated plots, might be a product of this greater soil mixing and disturbance, and indicate more available surfaces for chemical weathering processes.

The same thermal insulation that appears to promote higher levels of bioturbation might also promote more intense chemical weathering of these soils. Soil pH changes and cation leaching were noted within the first decade of snow manipulation by Weaver and Collins (1977), and Yano et al. (2015) measured similar soil pH changes under deeper snow conditions. The measured acidification with increased snow may reflect higher base cation leaching, but has previously been linked to higher microbial activity resulting from thermal insulation, whereby soil pH reflects increased microbial carbonic acid production and mineral oxidation (Sharp et al., 1999; Yano et al., 2015). Regardless of the mechanism of pH decrease, these changes are likely feeding back to enhance weathering at the site, and previous unpublished work noted higher losses of Ca, Mg, and K in soil leachate under 2x and 4x snow conditions (J. Brookshire, personal communication). This enhanced weathering environment, with increased soil respiration due to higher levels of microbial activity, is most likely another contributing factor to the measured reductions in C and N, where soil organic matter is decreased under deeper snow conditions.

Globally, soil pH is noted to vary nonlinearly with water balance (Slessarev et al., 2016) and these threshold style decreases in pH and available base cations have been noted with increased available water along rainfall gradients in the basaltic soils of Hawaii (Vitousek and Chadwick, 2013) and young, post-glacial soils of the South Island of New Zealand (Dixon et al., 2016). Modeling of associated thresholds by Vitousek et

al. (2016) suggested that these thresholds may be strongly modulated by short-term variable moisture availability, though the specific timescale of soil change was not constrained. Here, our results indicate increased acidity and base cation leaching. The loss of these macronutrients appears to be occurring much more rapidly (decadal timescales) than the rate that mineral weathering can rejuvenate the soil (likely millennial timescales) as long as these moisture and hydrologic conditions are maintained. The changes in snowpack observed in this study could therefore lead to effectively non-reversible acidification of these mountainous soils under relatively short temporal scales.

CONCLUSION

Utilizing one of the longest running snow manipulation sites in the world, we find that 48 years of enhanced snowpack drives significant changes in soils of a subalpine grassland. These changes include decreases in soil pH and particle size, and increases in bioturbation and surface erosion. We find both physical and chemical changes in these soils, highlighting the potential importance of weathering-erosion feedbacks in soil responses to perturbations. Increased bioturbation and soil mixing under enhanced snowpack may increase the ability to weather more mineral surfaces and provide more surface area for exchangeable cations and macronutrient pools. Thermal insulation from this deeper snow additionally produces microbial conditions that enhance mineral weathering and reduce the amount of organic matter in the subsurface, while increased water availability subsequently works to augment this chemical weathering by providing the necessary moisture and increased leaching of base cations. Our results indicate that increased snow overall contributes to a positive feedback creating more weathered soils in this subalpine grassland.

Our data also provide insight into the importance of vegetation change for surface weathering and erosion intensity, as changes in plant community composition at this site from tall bunch grasses to sparse annual forb species likely led to increased wind erosion from significantly reduced aboveground biomass. A concurrent 50% decline in grassland productivity mediated by increased late summer dryness at this field site measured by *Brookshire and Weaver, 2015*, may also lead to heightened wind erosion of these soils. More work is needed to constrain the exact mechanisms

influencing these vegetation shifts and subsequent biomass loss, however our results suggest that enhanced-snowpack driven vegetation changes may cause significant alteration in erosional mechanisms and may increase surface erosion.

Subsequently, how do the weathering processes we measure in this study inform Earth scientists interested in future climate impacts on soil development and fertility? Current “downscaled” global climate models (GCM’s) suggest the region will undergo significant warming in the coming decades, with significantly drier summers and changes in seasonal rainfall patterns (Saunders et al., 2011). However, the change in annual precipitation is relatively unclear, and climate models have not reliably indicated how snowpack will change across the Greater Yellowstone Ecosystem (McWethy et al., 2010). Somewhat counterintuitively, if this region sees declining snowpack (and snow manipulation is removed) it could lead to slow soil recovery if leaching intensity and wind erosion decrease, however the rate of recovery would be slow-on the scale of millennia- and dependent on existing cation abundance still left in unaltered mineral grains and the rates of mineral supply of bedrock to soil (averaging ~50 mm/ky for upland soils) (e.g., Dixon and von Blanckenburg, 2012). This work shows that short-term changes in climate variables have the capability of significantly altering weathering feedbacks and processes in soils, and potentially can drive a soil system towards irreversible thresholds in various soil characteristics.

In summary, this study found that rapid shifts in snowpack may have significant impacts on short term soil evolution by affecting a number of weathering processes. Importantly, these decadal perturbations in moisture availability through snowpack

variation in a snow dominated system can measurably overprint signals developed over thousands of years of soil development. Our work sheds light on the timescales of soil formation, and lends evidence to the idea that soil weathered state is a product of punctuated and rapid responses to short-term changes in climate variables, rather than a reflection of mean conditions over its formation.

REFERENCES CITED

- Baskaran, M., 2011, Po-210 and Pb-210 as atmospheric tracers and global atmospheric Pb-210 fallout: a review: *Journal of Environmental Radioactivity*, v. 102, no. 5, p. 500-513.
- Black, T. A., and Montgomery, D. R., 1991, Sediment transport by burrowing mammals, Marin County, California: *Earth Surface Processes and Landforms*, v. 16, no. 2, p. 163-172.
- Brookshire, E. N. J., and Weaver, T., 2015, Long-term decline in grassland productivity driven by increasing dryness: *Nature Communications*, v. 6, p. 7148.
- Burke, I. C., Yonker, C. M., Parton, W. J., Cole, C. V., Schimel, D. S., and Flach, K., 1989, Texture, Climate, and Cultivation Effects on Soil Organic Matter Content in U.S. Grassland Soils: *Soil Science Society of America Journal*, v. 53, no. 3, p. 800-805.
- Chadwick, O. A., and Chorover, J., 2001, The chemistry of pedogenic thresholds: *Geoderma*, v. 100, no. 3, p. 321-353.
- Chadwick, O. A., Derry, L. A., Vitousek, P. M., Huebert, B. J., and Hedin, L. O., 1999, Changing sources of nutrients during four million years of ecosystem development: *Nature*, v. 397, no. 6719, p. 491-497.
- DeMaster, D. J., and Cochran, J. K., 1982, Particle mixing rates in deep-sea sediments determined from excess ²¹⁰Pb and ³²Si profiles: *Earth and Planetary Science Letters*, v. 61, no. 2, p. 257-271.
- Dixon, J. L., Chadwick, O. A., and Vitousek, P. M., 2016, Climate-driven thresholds for chemical weathering in postglacial soils of New Zealand: *Journal of Geophysical Research: Earth Surface*, v. 121, no. 9, p. 1619-1634.
- Dixon, J. L., Hartshorn, A. S., Heimsath, A. M., DiBiase, R. A., and Whipple, K. X., 2012, Chemical weathering response to tectonic forcing: A soils perspective from the San Gabriel Mountains, California: *Earth and Planetary Science Letters*, v. 323, p. 40-49.
- Dixon, J. L., Heimsath, A. M., and Amundson, R., 2009a, The critical role of climate and saprolite weathering in landscape evolution: *Earth Surface Processes and Landforms*, v. 34, no. 11, p. 1507-1521.
- Dixon, J. L., Heimsath, A. M., Kaste, J., and Amundson, R., 2009b, Climate-driven processes of hillslope weathering: *Geology*, v. 37, no. 11, p. 975-978.
- Dixon, J. L., and von Blanckenburg, F., 2012, Soils as pacemakers and limiters of global silicate weathering: *Comptes Rendus Geoscience*, v. 344, no. 11, p. 597-609.

- Dörr, H., and Münnich, K., 1989, Downward movement of soil organic matter and its influence on trace-element transport (210 Pb, 137 Cs) in the soil: *Radiocarbon*, v. 31, no. 03, p. 655-663.
- Elmore, A., Kaste, J., Okin, G., and Fantle, M., 2008, Groundwater influences on atmospheric dust generation in deserts: *Journal of Arid Environments*, v. 72, no. 10, p. 1753-1765.
- Epstein, H., Lauenroth, W., and Burke, I., 1997, Effects of temperature and soil texture on ANPP in the US Great Plains: *Ecology*, v. 78, no. 8, p. 2628-2631.
- Fryrear, D. W., 1985, Soil cover and wind erosion: *Trans. ASAE*, v. 28, no. 3, p. 781-784.
- Gabet, E. J., Reichman, O., and Seabloom, E. W., 2003, The effects of bioturbation on soil processes and sediment transport: *Annual Review of Earth and Planetary Sciences*, v. 31, no. 1, p. 249-273.
- Gill, T. E., 1996, Eolian sediments generated by anthropogenic disturbance of playas: human impacts on the geomorphic system and geomorphic impacts on the human system: *Geomorphology*, v. 17, no. 1-3, p. 207-228.
- Hansen, R. M., and Morris, M., 1968, Movement of rocks by northern pocket gophers: *Journal of Mammalogy*, v. 49, no. 3, p. 391-399.
- He, Q., and Walling, D., 1996, USE OF FALLOUT Pb-210 MEASUREMENTS TO INVESTIGATE LONGER-TERM RATES AND PATTERNS OF OVERBANK SEDIMENT DEPOSITION ON THE FLOODPLAINS OF LOWLAND RIVERS: *Earth Surface Processes and Landforms*, v. 21, no. 2, p. 141-154.
- Holsinger, J. P., 2013, Biogeochemical and plant functional group response to long-term snow manipulation in a subalpine grassland: Montana State University-Bozeman, College of Agriculture.
- Humphreys, G., and Mitchell, P., 1983, A preliminary assessment of the role of bioturbation and rainwash on sandstone hillslopes in the Sydney Basin: *Aspects of Australian sandstone landscapes*, p. 66-80.
- Kaste, J. M., Heimsath, A. M., and Bostick, B. C., 2007, Short-term soil mixing quantified with fallout radionuclides: *Geology*, v. 35, no. 3, p. 243-246.
- Kaste, J. M., Heimsath, A. M., and Hohmann, M., 2006, Quantifying sediment transport across an undisturbed prairie landscape using cesium-137 and high resolution topography: *Geomorphology*, v. 76, no. 3, p. 430-440.

- Kelly, E. F., Chadwick, O. A., and Hilinski, T. E., 1998, The effect of plants on mineral weathering, *Plant-induced soil changes: Processes and feedbacks*, Springer, p. 21-53.
- Locke, W. W., Lageson, D. R., and Bozeman, M., TRIP 4 ROAD LOG GEOLOGY AND GEOMORPHOLOGY OF THE ROCKY MOUNTAINS/GREAT PLAINS TRANSITION, *in* Proceedings Geologic Resources of Montana: 1989 Field Conference Guidebook 1989, Volume 2, Montana Geological Society, p. 463.
- Mabit, L., Benmansour, M., and Walling, D., 2008, Comparative advantages and limitations of the fallout radionuclides ^{137}Cs , ^{210}Pb ex and ^7Be for assessing soil erosion and sedimentation: *Journal of environmental radioactivity*, v. 99, no. 12, p. 1799-1807.
- Mabit, L., Meusburger, K., Fulajtar, E., and Alewell, C., 2013, The usefulness of ^{137}Cs as a tracer for soil erosion assessment: a critical reply to Parsons and Foster (2011): *Earth-science reviews*, v. 127, p. 300-307.
- Maher, K., 2010, The dependence of chemical weathering rates on fluid residence time: *Earth and Planetary Science Letters*, v. 294, no. 1, p. 101-110.
- Maher, K., 2011, The role of fluid residence time and topographic scales in determining chemical fluxes from landscapes: *Earth and Planetary Science Letters*, v. 312, no. 1, p. 48-58.
- McWethy, D., Gray, S., Higuera, P., Littell, J., Pederson, G., Ray, A., and Whitlock, C., 2010, Climate and terrestrial ecosystem change in the US Rocky Mountains and Upper Columbia Basin: historical and future perspectives for natural resource management: Natural Resource Report NPS/GRYN/NRR—2010/260. National Park Service, Fort Collins, Colorado.
- Millot, R., Gaillardet, J., Dupré, B., and Allègre, C. J., 2002, The global control of silicate weathering rates and the coupling with physical erosion: new insights from rivers of the Canadian Shield: *Earth and Planetary Science Letters*, v. 196, no. 1, p. 83-98.
- Munson, S. M., Belnap, J., and Okin, G. S., 2011, Responses of wind erosion to climate-induced vegetation changes on the Colorado Plateau: *Proceedings of the National Academy of Sciences*, v. 108, no. 10, p. 3854-3859.
- Naudin, K., Scopel, E., Andriamandroso, A., Rakotosolofo, M., Ratsimbazafy, N. A., Rakotozandriny, J. d. N., Salgado, P., and Giller, K., 2012, Trade-offs between biomass use and soil cover. The case of rice-based cropping systems in the Lake Alaotra region of Madagascar: *Experimental Agriculture*, v. 48, no. 02, p. 194-209.

- Neal, J. T., and Motts, W. S., 1967, Recent geomorphic changes in playas of western United States: *The Journal of Geology*, v. 75, no. 5, p. 511-525.
- O'farrell, C. R., Heimsath, A. M., and Kaste, J. M., 2007, Quantifying hillslope erosion rates and processes for a coastal California landscape over varying timescales: *Earth Surface Processes and Landforms*, v. 32, no. 4, p. 544-560.
- Okin, G., Gillette, D., and Herrick, J., 2006, Multi-scale controls on and consequences of aeolian processes in landscape change in arid and semi-arid environments: *Journal of Arid Environments*, v. 65, no. 2, p. 253-275.
- Pederson, G. T., Gray, S. T., Woodhouse, C. A., Betancourt, J. L., Fagre, D. B., Littell, J. S., Watson, E., Luckman, B. H., and Graumlich, L. J., 2011, The unusual nature of recent snowpack declines in the North American Cordillera: *Science*, v. 333, no. 6040, p. 332-335.
- Pierce, K. L., 1982, History and dynamics of glaciation in the northern Yellowstone National Park area.
- Rasmussen, C., Brantley, S., Blum, A., Dixon, J., and White, A., 2011, Strong climate and tectonic control on plagioclase weathering in granitic terrain: *Earth and Planetary Science Letters*, v. 3, no. 301, p. 521-530.
- Rhodes, D., 1957, The effect of pH on the uptake of radioactive isotopes from solution by a soil: *Soil Science Society of America Journal*, v. 21, no. 4, p. 389-392.
- Riebe, C. S., Kirchner, J. W., and Finkel, R. C., 2003, Long-term rates of chemical weathering and physical erosion from cosmogenic nuclides and geochemical mass balance: *Geochimica et Cosmochimica Acta*, v. 67, no. 22, p. 4411-4427.
- Riebe, C. S., Kirchner, J. W., Granger, D. E., and Finkel, R. C., 2001, Strong tectonic and weak climatic control of long-term chemical weathering rates: *Geology*, v. 29, no. 6, p. 511-514.
- Saunders, S., Findlay, D., Easley, T., and Christensen, S., 2011, Greater Yellowstone in Peril: The Threats of Climate Disruption.
- Saxton, K. E., and Rawls, W. J., 2006, Soil Water Characteristic Estimates by Texture and Organic Matter for Hydrologic Solutions: *Soil Science Society of America Journal*, v. 70, no. 5, p. 1569-1578.
- Sharp, M., Parkes, J., Cragg, B., Fairchild, I. J., Lamb, H., and Tranter, M., 1999, Widespread bacterial populations at glacier beds and their relationship to rock weathering and carbon cycling: *Geology*, v. 27, no. 2, p. 107-110.

- Skipp, B., Lageson, D., and McMannis, W., 2004, US DEPARTMENT OF THE INTERIOR US GEOLOGICAL SURVEY GEOLOGIC INVESTIGATIONS SERIES I-2634 Version 2.1 GEOLOGIC MAP OF THE SEDAN QUADRANGLE, GALLATIN AND PARK.
- Slessarev, E., Lin, Y., Bingham, N., Johnson, J., Dai, Y., Schimel, J., and Chadwick, O., 2016, Water balance creates a threshold in soil pH at the global scale: *Nature*, v. 540, no. 7634, p. 567-569.
- Stallard, R. F., 1985, River chemistry, geology, geomorphology, and soils in the Amazon and Orinoco basins, *The chemistry of weathering*, Springer, p. 293-316.
- Thorn, C. E., 1978, A preliminary assessment of the geomorphic role of pocket gophers in the alpine zone of the Colorado Front Range: *Geografiska Annaler. Series A. Physical Geography*, p. 181-187.
- Tyler, A. N., Carter, S., Davidson, D. A., Long, D. J., and Tipping, R., 2001, The extent and significance of bioturbation on ¹³⁷Cs distributions in upland soils: *Catena*, v. 43, no. 2, p. 81-99.
- Vitousek, P., Dixon, J. L., and Chadwick, O. A., 2016, Parent material and pedogenic thresholds: observations and a simple model: *Biogeochemistry*, v. 130, no. 1-2, p. 147-157.
- Vitousek, P. M., and Chadwick, O. A., 2013, Pedogenic thresholds and soil process domains in basalt-derived soils: *Ecosystems*, v. 16, no. 8, p. 1379-1395.
- Wallbrink, P., and Murray, A., 1996, Determining soil loss using the inventory ratio of excess lead-210 to cesium-137: *Soil Science Society of America Journal*, v. 60, no. 4, p. 1201-1208.
- Walling, D., 1998, Use of ¹³⁷Cs and other fallout radionuclides in soil erosion investigations: progress, problems and prospects: *Use of*, v. 137, p. 39-62.
- Weaver, T., 1974, Ecological effects of weather modification: effect of late snowmelt on *Festuca idahoensis* Elmer meadows: *American midland naturalist*, p. 346-356.
- Weaver, T., and Collins, D., 1977, Possible effects of weather modification (increased snowpack) on *Festuca idahoensis* meadows: *Journal of Range Management*, p. 451-456.
- West, A. J., Galy, A., and Bickle, M., 2005, Tectonic and climatic controls on silicate weathering: *Earth and Planetary Science Letters*, v. 235, no. 1, p. 211-228.
- White, A. F., Blum, A. E., Bullen, T. D., Vivit, D. V., Schulz, M., and Fitzpatrick, J., 1999, The effect of temperature on experimental and natural chemical weathering

rates of granitoid rocks: *Geochimica et Cosmochimica Acta*, v. 63, no. 19, p. 3277-3291.

Wilkinson, M. T., Richards, P. J., and Humphreys, G. S., 2009, Breaking ground: pedological, geological, and ecological implications of soil bioturbation: *Earth-Science Reviews*, v. 97, no. 1, p. 257-272.

Wipf, S., and Rixen, C., 2010, A review of snow manipulation experiments in Arctic and alpine tundra ecosystems: *Polar Research*, v. 29, no. 1, p. 95-109.

Wright, S., Howard, B., Strand, P., Nylén, T., and Sickel, M., 1999, Prediction of ¹³⁷Cs deposition from atmospheric nuclear weapons tests within the Arctic: *Environmental Pollution*, v. 104, no. 1, p. 131-143.

Yano, Y., Brookshire, E. J., Holsinger, J., and Weaver, T., 2015, Long-term snowpack manipulation promotes large loss of bioavailable nitrogen and phosphorus in a subalpine grassland: *Biogeochemistry*, v. 124, no. 1-3, p. 319-333.

APPENDICES

APPENDIX A

SAMPLING METHODS

During the Fall Semester of 2015, nine soil cores were collected from the Bangtail field site approximately 24 km NW of Bozeman, Montana. Three boring locations were chosen for each snow depth area for replication purposes. Soil cores were extracted using a vertical hammer coring device with plastic cylinder tubes within the coring device. The core cylinders were removed from the device, capped, and labeled for correct orientation and depth increment. Cores were collected to a depth of 35 cm. We chose this depth due to the common depth trends for Fallout Radionuclides and their distribution in soil profiles, and the amount of time required to gain FRN data for this project (over 7 months of constant data collection). Collected cores were taken from the Bangtail Field Sight and brought back to the Department of Earth Sciences. Here, the cores were sectioned into 1.5 cm increments for high resolution analysis of FRN's, particle size, Carbon and Nitrogen analysis, pH, and elemental chemistry analysis. Due to the coring device design, depth sections 0-15 cm and 20-35 cm were consistent 1.5 cm increments, however the 15-20 cm sections could only be separated into two, 2.5 cm sections. Large organics were removed and the incremented samples were dried for 48 hours at 90 °C before further packaging. One representative core from each snow plot area was then chosen for further analysis of particle size, total C and N, pH, and elemental chemistry.

APPENDIX B

PARTICLE SIZE ANALYSIS

This study area provides a perfect laboratory to see how snowpack addition over decadal scales may affect soil texture. In general, soil texture plays a critical role in various soil characteristics such as hydraulic conductivity, porosity, soil carbon amounts, primary productivity, etc. (Burke et al., 1989; Epstein et al., 1997; Saxton and Rawls, 2006). Soil texture also plays a key role in the behavior of Fallout Radionuclides upon deposition due to their high affinity for fine grained particles (Kaste et al., 2007; Tyler et al., 2001).

One soil core from each snow plot area was chosen for detailed particle size analysis. Each core consists of samples 1.5 cm in thickness to approximately 35 cm in depth, with two 2.5 cm increment samples. These samples were removed from their wax-sealed casings following fallout radionuclide analysis and prepared for particle size measurement with a Malvern Mastersizer 3000 Particle Size Analyzer in the Environmental Analysis Laboratory of the Land Resources and Environmental Science Department (LRES) of MSU. Samples were subsampled into small cryotubes (approximately 1 gram of soil material) containers to achieve a reasonable amount of soil for measurement in the Malvern instrument. Each sample was then mixed with an appropriate amount of sodium hexametaphosphate to promote particle suspension and deflocculation (20 mL of NA-Hex was used per sample, concentration of 2.5 g of solid per liter of DI water). This slurry was then put on a stir pad until all particles were in complete suspension. A pipette was used to withdraw the most representative subsample while the mixture was stirring on the stir pad. The sample was then added into the instrument for measurement. The Malvern Mastersizer 3000 is a laser diffraction

instrument that sends the slurry mixture through a chamber for diffraction analysis to obtain particle size. Therefore, a correct laser obscuration range was necessary for the instrument to achieve correct results. Samples were added to the instrument until the obscuration range was within 5-7%, as dictated by the software. The Malvern Mastersizer 3000 then runs three consecutive measurements each sample, then was cleaned thoroughly for each new sample. The result is a full range of particle sizes by percent volume. This type of result layout is the most effective with this type of instrument, as unnecessary error is added to the results if they are transformed to a particle size count number layout. These three measurements were then processed for average particle size percent volumes per profile increment, using bin sizes of clay, silt, and sand for data interpretation and presentation. Particle size could then be plotted with depth and correlated to other soil weathering metrics.

Table A1. Summative Particle Size data for Control, SM2x, and SM4x plots

Control Sample	Clay % Vol.	Silt % Vol.	Sand % Vol.	SM4x Sample	Clay % Vol.	Silt % Vol.	Sand % Vol.	Depth
Ca1	0.1	5.4	92.9	Ha1x	2.2	30.1	67.0	0.75
Ca2x	0.5	13.7	85.6	Ha1y	1.7	23.3	74.7	0.75
Ca2y	0.7	13.2	85.0	Ha1z	1.2	21.1	77.1	0.75
Ca2z	0.3	9.6	89.0	Ha2	2.3	30.5	67.0	2.25
Ca3	1.7	25.6	72.7	Ha3	1.7	22.7	75.6	3.75
Ca4	1.8	28.1	70.0	Ha4	1.7	24.4	73.9	5.25
Ca5	2.3	31.1	66.6	Ha5	2.1	27.3	69.8	6.75
Ca6	2.3	31.3	66.4	Ha6	2.2	28.1	69.7	8.25
Ca7	3.0	40.2	56.8	Ha7	2.2	29.9	67.8	9.75
Ca8	2.0	32.8	65.2	Ha8	4.1	38.4	57.0	11.25
Ca9	2.1	34.9	63.0	Ha9	2.9	30.1	67.1	12.75
Ca10	2.3	36.0	61.5	Ha10	3.6	32.7	63.2	14.25
Ca11	3.0	40.8	56.1	Ha11	3.3	32.3	64.4	16.25
Ca12	2.6	37.4	60.0	Ha12	4.0	39.4	56.3	18.75
Cb1	2.6	38.5	59.0	Hb1	2.8	37.4	59.8	20.75
Cb2	2.0	34.2	63.9	Hb2	4.5	52.1	43.5	22.25
Cb3	2.5	37.1	60.3	Hb3	3.1	38.9	57.9	23.75
Cb4	3.0	42.3	54.7	Hb4	4.7	54.6	40.8	25.25
Cb5	3.1	42.3	54.6	Hb5	4.3	46.6	49.0	26.75
Cb6	3.3	45.8	51.0	Hb6	5.6	49.4	45.0	28.25
Cb7	3.3	41.7	54.9	Hb7	6.6	53.7	39.7	29.75
Cb8	3.9	47.9	48.2	Hb8	9.0	61.9	29.1	31.25
Cb9	4.8	49.9	45.3	Hb9	7.5	52.8	39.7	32.75
Cb10	3.5	48.5	48.0	Hb10	5.9	43.6	50.5	34.25
SM2x Sample	Clay % Vol.	Silt % Vol.	Sand % Vol.					
La1x	3.2	41.6	53.7					
La1y	1.4	24.9	73.7					
La1z	1.2	21.3	77.4					
La2	3.0	38.7	58.2					
La3	2.1	30.5	67.4					
La4	2.2	29.5	68.2					
La5	2.6	30.3	67.1					
La6	2.8	34.3	62.8					
La7	2.4	30.4	67.2					
La8	2.7	33.6	63.7					
La9	2.6	32.5	64.9					
La10	4.8	52.0	43.2					
La11	2.9	38.7	58.3					
La12	5.8	56.5	37.7					
Lb1	4.9	56.6	38.5					
Lb2	3.9	49.1	47.0					
Lb3	4.3	48.2	47.5					
Lb4	3.0	39.1	58.0					
Lb5	5.1	46.8	48.1					
Lb6	5.7	51.5	42.8					
Lb7	6.2	53.1	40.7					
Lb8	7.0	54.0	39.0					
Lb9	6.9	51.2	42.0					
Lb10	6.2	42.4	51.3					

Table A2. Standard deviations of individual sample particle sizes (n=3).

Control s.d			SM2x s.d			SM4x s.d		
Clay	Silt	Sand	Clay	Silt	Sand	Clay	Silt	Sand
0.07	0.32	0.27	0.10	0.65	0.69	0.15	1.02	0.41
0.07	0.82	0.56	0.15	1.30	1.31	0.11	0.66	0.82
0.07	0.79	0.24	0.14	1.29	1.30	0.16	0.97	1.01
0.05	0.77	0.73	0.19	1.55	1.65	0.18	1.49	1.59
0.13	1.18	1.34	0.11	0.85	0.94	0.25	1.84	2.07
0.14	1.33	1.48	0.16	1.31	1.47	0.20	1.59	1.71
0.22	2.04	2.28	0.15	1.26	1.43	0.23	2.06	1.13
0.21	1.97	2.19	0.15	1.40	1.39	0.20	1.37	1.55
0.18	1.42	1.61	0.18	1.46	1.64	0.16	1.10	1.28
0.17	1.73	1.91	0.19	1.63	1.81	0.11	0.42	0.95
0.18	1.68	1.86	0.14	1.21	1.36	0.18	1.26	1.44
0.11	1.29	1.57	0.17	0.87	1.03	0.27	1.76	1.83
0.30	2.76	3.08	0.43	3.53	3.87	0.24	1.29	1.55
0.23	2.30	2.54	0.31	1.41	1.70	0.26	1.52	1.73
0.19	1.86	2.07	0.23	0.90	1.15	0.24	1.55	1.81
0.21	2.14	2.35	0.25	1.64	1.88	0.28	1.96	2.25
0.18	1.74	1.93	0.30	2.10	2.42	0.27	1.63	1.90
0.31	2.85	3.16	0.38	2.58	2.95	0.38	2.63	3.03
0.23	1.80	2.02	0.39	1.74	2.14	0.31	1.63	1.96
0.29	2.44	2.74	0.42	1.84	2.26	0.38	1.46	1.83
0.26	2.04	2.31	0.43	1.86	2.27	0.42	1.43	1.87
0.31	2.32	2.64	0.39	1.16	1.53	0.41	0.36	0.76
0.38	1.83	2.19	0.51	2.02	2.53	0.54	2.02	2.56
0.29	2.08	2.36	0.36	1.27	1.62	0.36	0.94	1.30

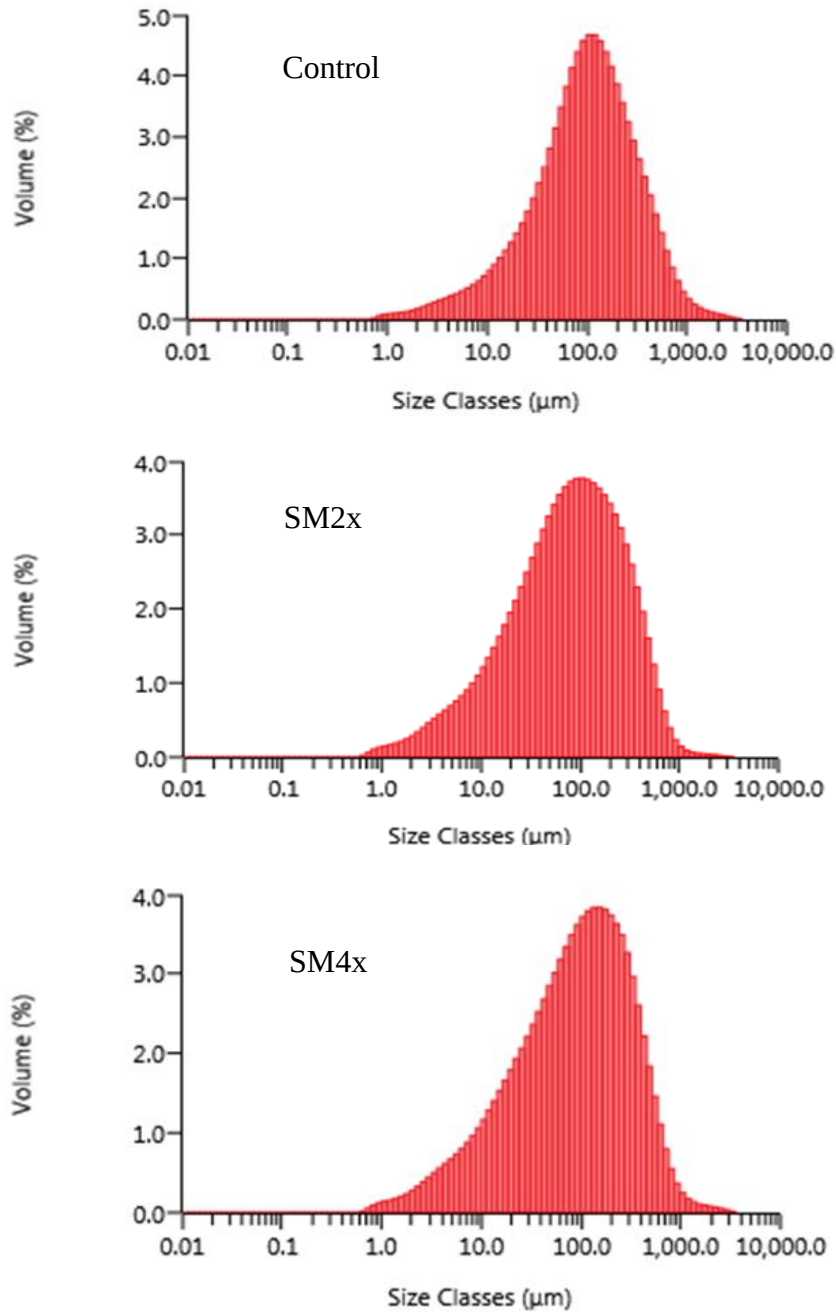


Figure A1. Average particle size histograms from the control and snow addition sites.

References Cited

- Burke, I. C., Yonker, C. M., Parton, W. J., Cole, C. V., Schimel, D. S., and Flach, K., 1989, Texture, Climate, and Cultivation Effects on Soil Organic Matter Content in U.S. Grassland Soils: Soil Science Society of America Journal, v. 53, no. 3, p. 800-805.
- DeMaster, D. J., and Cochran, J. K., 1982, Particle mixing rates in deep-sea sediments determined from excess ^{210}Pb and ^{32}Si profiles: Earth and Planetary Science Letters, v. 61, no. 2, p. 257-271.
- Dixon, J. L., Heimsath, A. M., and Amundson, R., 2009, The critical role of climate and saprolite weathering in landscape evolution: Earth Surface Processes and Landforms, v. 34, no. 11, p. 1507-1521.
- Epstein, H., Lauenroth, W., and Burke, I., 1997, Effects of temperature and soil texture on ANPP in the US Great Plains: Ecology, v. 78, no. 8, p. 2628-2631.
- He, Q., and Walling, D., 1996, USE OF FALLOUT Pb-210 MEASUREMENTS TO INVESTIGATE LONGER-TERM RATES AND PATTERNS OF OVERBANK SEDIMENT DEPOSITION ON THE FLOODPLAINS OF LOWLAND RIVERS: Earth Surface Processes and Landforms, v. 21, no. 2, p. 141-154.
- Kaste, J. M., Heimsath, A. M., and Bostick, B. C., 2007, Short-term soil mixing quantified with fallout radionuclides: Geology, v. 35, no. 3, p. 243-246.
- Ligero, R., Barrera, M., Casas-Ruiz, M., Sales, D., and López-Aguayo, F., 2002, Dating of marine sediments and time evolution of heavy metal concentrations in the Bay of Cádiz, Spain: Environmental Pollution, v. 118, no. 1, p. 97-108.
- Saxton, K. E., and Rawls, W. J., 2006, Soil Water Characteristic Estimates by Texture and Organic Matter for Hydrologic Solutions: Soil Science Society of America Journal, v. 70, no. 5, p. 1569-1578.
- Tyler, A. N., Carter, S., Davidson, D. A., Long, D. J., and Tipping, R., 2001, The extent and significance of bioturbation on ^{137}Cs distributions in upland soils: Catena, v. 43, no. 2, p. 81-99.

APPENDIX C

TOTAL CARBON AND NITROGEN ANALYSIS

The same representative cores used for particle size analysis were used to find total Carbon and Nitrogen amounts among snowpack depths. Upon removal from their wax-sealed containers used for FRN analysis via gamma spectroscopy, 10 grams of soil material were subsampled from each profile increment to a depth of 35 cm, totaling in 66 samples for this analysis. These subsamples were then homogenized and pulverized over 12 hours using a steel jar mill located in the LRES department of MSU. Of this pulverized material, 50 mg of each sample were taken and placed into Costech pressed tin capsules (3.5x5mm) and sealed for analysis. Once prepared, samples were ran for total C and N by the staff at the Environmental Analysis Laboratory (EAL) at MSU using a Costech Instruments Elemental Combustion System. The data was obtained from the EAL staff for further processing. This instrument provides data in the form of percent total mass of Carbon and Nitrogen, as well as C:N ratios of the soil material. The soil samples needed to be pulverized for adequate combustion and representative results. These values could then be plotted with depth to understand C and N trends in the soil profile. Statistical tests were performed for significant correlations in averages, variance, and total amounts.

Table A3. Summative Control plot Carbon and Nitrogen data.

Sample ID	run date	%TN	%TN uncertainty	%TN flag	%TC	%TC uncertainty	%TC flag	C/N	error
Ca1	161214	1.26	0.06	ok	18.97	0.95	bql	15.01	0.05
Ca1	161214	1.27	0.06	ok	18.76	0.94	bql	14.76	0.05
Ca1#	161214	2.25	0.11	ok	30.38	1.52	ok	13.53	0.05
Ca1#	161212	1.13	0.06	aql	23.83	1.19	aql	21.05	0.05
Ca2	161214	0.35	0.02	ok	5.16	0.26	ok	14.95	0.05
Ca2#	161212	0.61	0.03	aql	10.29	0.51	FALSE	16.82	0.05
Ca3	161212	0.30	0.02	ok	3.85	0.19	ok	12.67	0.05
Ca4	161212	0.30	0.01	ok	3.60	0.18	ok	12.13	0.05
Ca5	161212	0.25	0.01	ok	2.73	0.14	ok	11.12	0.05
Ca5y	161214	0.24	0.01	ok	2.66	0.13	ok	11.13	0.05
Ca5z	161214	0.24	0.01	ok	2.62	0.13	ok	11.14	0.05
Ca6	161212	0.22	0.01	ok	2.42	0.12	ok	11.19	0.05
Ca7	161212	0.24	0.01	ok	2.61	0.13	ok	10.82	0.05
Ca8	161212	0.25	0.01	ok	2.67	0.13	ok	10.81	0.05
Ca9	161212	0.24	0.01	ok	2.66	0.13	ok	10.93	0.05
Ca10	161212	0.22	0.01	ok	2.40	0.12	ok	10.79	0.05
Ca11	161212	0.21	0.01	ok	2.16	0.11	ok	10.33	0.05
Ca12	161212	0.20	0.01	ok	2.07	0.10	ok	10.49	0.05
Cb2	161212	0.19	0.01	ok	1.97	0.10	ok	10.36	0.05
Cb4	161212	0.17	0.01	ok	1.75	0.09	ok	10.17	0.05
Cb6	161212	0.18	0.01	ok	1.84	0.09	ok	10.43	0.05
Cb8	161212	0.15	0.01	ok	1.52	0.08	ok	10.43	0.05
Cb10	161212	0.14	0.01	ok	1.50	0.08	ok	10.49	0.05

This symbolizes replicate values that are outside the range of normal values and therefore we use the initial value, or averaged replicates that are within the appropriate error. Error and uncertainty (5%) is based off of instrument sensitivity dictated by Costech software. These data represent total mass percent of Carbon and Nitrogen in Control plot samples. Duplicate samples were ran for replication purposes.

Table A4. Summative SM2x plot Carbon and Nitrogen data.

Sample ID	run date	%TN	%TN uncertainty	%TN flag	%TC	%TC uncertainty	%TC flag	C/N	error
La1	161212	0.28	0.01	ok	3.29	0.16	ok	11.68	0.05
La2	161212	0.32	0.02	ok	3.88	0.19	aql	11.98	0.05
La2	161214	0.32	0.02	ok	3.79	0.19	ok	11.87	0.05
La3	161212	0.34	0.02	ok	4.03	0.20	aql	11.92	0.05
La4	161212	0.30	0.02	ok	3.59	0.18	ok	11.81	0.05
La4	161214	0.30	0.01	ok	3.51	0.18	ok	11.73	0.05
La5	161212	0.26	0.01	ok	2.92	0.15	ok	11.27	0.05
La5y	161214	0.25	0.01	ok	2.78	0.14	ok	11.22	0.05
La5z	161214	0.26	0.01	ok	2.87	0.14	ok	11.03	0.05
La6	161212	0.25	0.01	ok	2.87	0.14	ok	11.39	0.05
La6	161214	0.25	0.01	ok	2.78	0.14	ok	11.26	0.05
La7	161212	0.27	0.01	ok	3.14	0.16	ok	11.47	0.05
La8	161214	0.26	0.01	ok	2.87	0.14	ok	11.11	0.05
La9	161212	0.25	0.01	ok	2.61	0.13	ok	10.52	0.05
La10	161212	0.21	0.01	ok	2.17	0.11	ok	10.31	0.05
La11	161212	0.23	0.01	ok	1.80	0.09	ok	7.95	0.05
La12#	161212	0.27	0.01	ok	1.81	0.09	ok	6.82	0.05
La12	161214	0.17	0.01	ok	1.79	0.09	ok	10.35	0.05
Lb2	161212	0.17	0.01	ok	1.72	0.09	ok	10.28	0.05
Lb4	161212	0.15	0.01	ok	1.58	0.08	ok	10.52	0.05
Lb4	161214	0.15	0.01	ok	1.52	0.08	ok	10.51	0.05
Lb6	161212	0.15	0.01	ok	1.57	0.08	ok	10.77	0.05
Lb8	161212	0.13	0.01	ok	1.44	0.07	ok	10.71	0.05
Lb10	161212	0.12	0.01	ok	1.26	0.06	bql	10.65	0.05

Table A5. Summative SM4x plot Carbon and Nitrogen data.

sample ID	run date	%TN	%TN uncertainty	%TN flag	%TC	%TC uncertainty	%TC flag	C/N	error
Ha1	161212	0.33	0.02	ok	4.17	0.21	aql	12.67	0.05
Ha2	161212	0.28	0.01	ok	3.53	0.18	ok	12.68	0.05
Ha3	161212	0.26	0.01	ok	3.23	0.16	ok	12.22	0.05
Ha4	161212	0.27	0.01	ok	3.32	0.17	ok	12.22	0.05
Ha5	161212	0.25	0.01	ok	3.02	0.15	ok	12.08	0.05
Ha5y	161214	0.25	0.01	ok	2.91	0.15	ok	11.73	0.05
Ha5z	161214	0.26	0.01	ok	3.02	0.15	ok	11.85	0.05
Ha6	161212	0.26	0.01	ok	3.03	0.15	ok	11.62	0.05
Ha7	161212	0.24	0.01	ok	2.63	0.13	ok	11.13	0.05
Ha8	161212	0.22	0.01	ok	2.35	0.12	ok	10.88	0.05
Ha9	161212	0.20	0.01	ok	2.22	0.11	ok	11.14	0.05
Ha10	161212	0.21	0.01	ok	2.32	0.12	ok	11.19	0.05
Ha11	161212	0.21	0.01	ok	2.30	0.11	ok	11.05	0.05
Ha12	161212	0.18	0.01	ok	1.88	0.09	ok	10.50	0.05
Hb2	161212	0.17	0.01	ok	1.74	0.09	ok	10.19	0.05
Hb4	161212	0.14	0.01	ok	1.39	0.07	ok	10.22	0.05
Hb6	161212	0.22	0.01	ok	2.35	0.12	ok	10.67	0.05
Hb8	161212	0.11	0.01	ok	1.15	0.06	bql	10.35	0.05
Hb10	161212	0.10	0.01	ok	1.02	0.05	bql	10.14	0.05

APPENDIX D

FALLOUT RADIONUCLIDE MEASUREMENTS AND METHOD

FRN samples were created by packing the dried soil material into a container of known geometry and volume. These sample containers were then taped closed with standard electrical tape, and subsequently sealed using melted paraffin wax. This sealing is necessary to ensure that our measurements of Pb-210 are correct. Since Pb-210 comes in two varieties in soils, supported (in-situ) and unsupported (meteoric), the sample must be sealed to achieve secular equilibrium of the in-situ Pb-210 with its parent isotope Ra-226. This process is considered to take roughly 5 half-lives of Rn-222 (half-life of 3.8 days), or approximately 19 days (He and Walling, 1996; Ligerio et al., 2002). This process is necessary to differentiate the supported from the unsupported, which we call Pb-210 excess. After this time period, samples were ran between 24-48 hours in a Canberra Industries high purity, broad energy Germanium Detector (BEGe) in the Fallout Radionuclide Laboratory at MSU. Using the GENIE 2K software package for FRN detections, Pb-210 is measured at an energy peak of 46 kiloelectronvolts (keV), while Cs-137 is measured at an energy of 661 keV. Counts/sec/kg are converted to FRN activities (abundances) as describe below.

When measuring any FRN activities, it is imperative to recognize the efficiency level of the analysis. Due to the sensitive nature of gamma spectroscopy with germanium detectors, we must perform system calibrations to maintain representative results of FRN activities. These energy and efficiency calibrations help us determine both the efficiency of the detector at recording gamma particle interactions and the efficiency of the sample geometry within the detector. These are performed by running a gamma standards Co-60, Cs-137, and Ba-133 in the detectors for 1-3 minutes, then the measured energy peaks for

these standards are compared to the calculated values by the GENIE 2K software. If these values fit the same curve for both the energy and efficiency calibrations, then the detectors are running at optimal conditions for measuring activities. This process is important because the detector efficiency for measuring certain energy levels is already very low. For example, the detector efficiency for measuring Cs-137 gamma emission from a source material is approximately 4%, which makes calibration a very important process to achieve quality results. Due to background radiation in the laboratory, calibrations were performed multiple times over the 7 month data collection period.

In order to obtain FRN activities, we take the counts/sec/kg based on count time and sample mass at a known keV (661 keV for Cs-137, 46 keV for Pb-210). We then take the counts/sec/kg divided by the emission probability of the FRN and efficiency of the detector to obtain an activity in Becquerels (Bq)/kg. This value is multiplied by an accepted soil density for the soil type of our field site to gain activity values in units of Bq/cm³. To gain depth corrected values, we multiply the activity per cm³ by the sample depth increment to gain activity levels in units of Bq/cm². As discussed above, we derive Pb-210 excess (unsupported) from subtracting Ra-226 activity from total Pb-210 activity. This process accounts for the relationship of secular equilibrium between Pb-210 in-situ (supported) and its parent Ra-226.

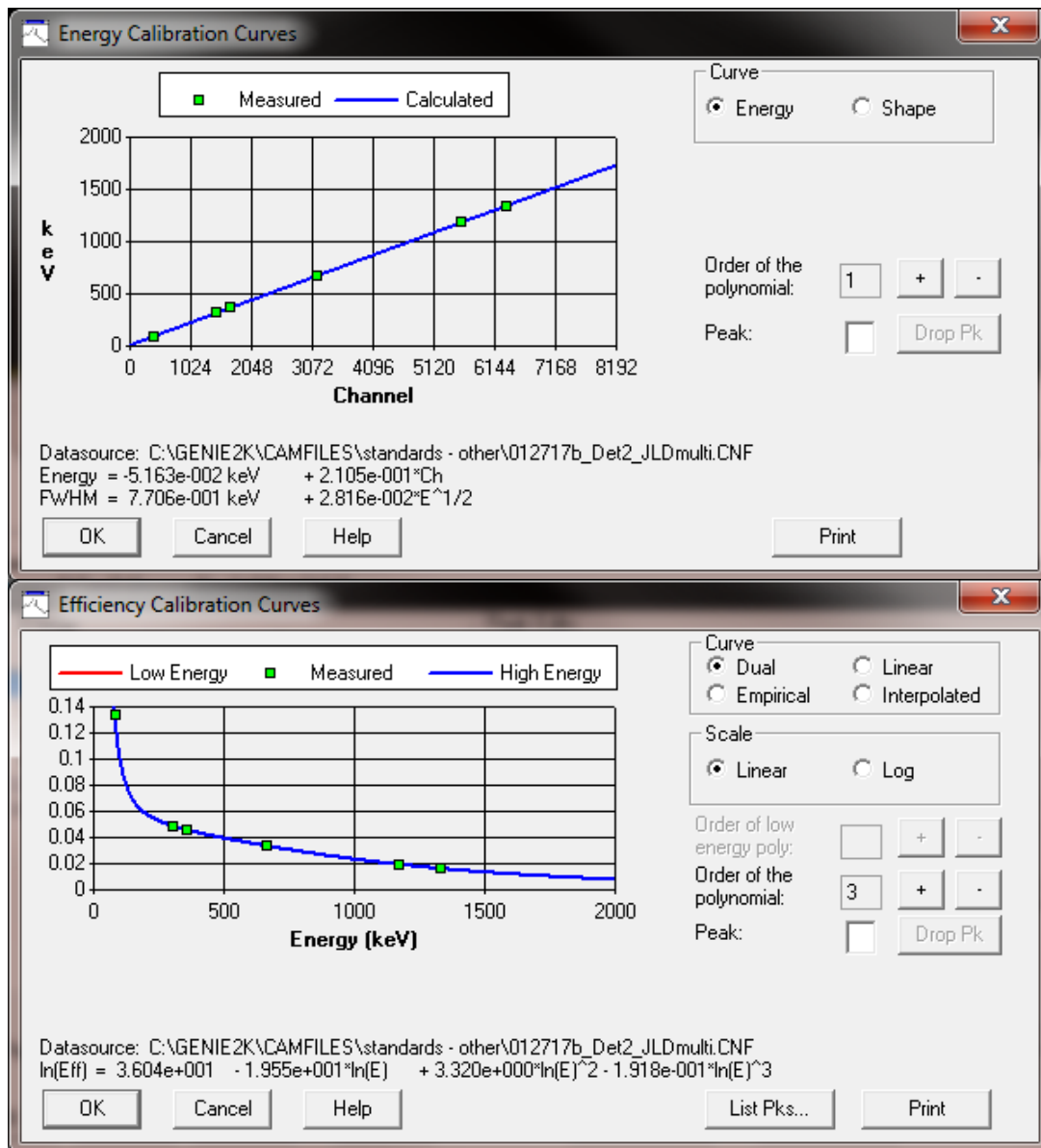


Figure A2. Examples of calibrations performed periodically throughout data collection. Measured value must lie on the calculated curves for proper calibration.

FRN activity profiles give us a wealth of information regarding the isotopic character of soils underlying different snowpack heights. To better understand the timescales and transported processes we coupled these data with advection-diffusion numerical models. The diffusive processes in soils have been found to be significant drivers of soil evolution over a variety of short timescales (Dixon et al., 2009a; Kaste et al., 2007; Tyler et al., 2001), and therefore it is important to view our measured data with the ideal curve for the FRNs Cs-137 and Pb-210. Here, our unique field site presented challenges for modeling FRN data. Since the control plot has been left without manipulation, steady-state conditions could be assumed for modeling purposes. This steady-state conditions assumes that unsupported Pb-210 meteoric inputs are balanced by losses from radioactive decay. The depth distribution of Pb-210 in a soil profile can then be described by the following advection-diffusion model (DeMaster and Cochran, 1982))

$$A(z) = A_0 \left(\frac{V - \sqrt{V^2 + 4\lambda D}}{2D} (z) \right) \quad (\text{Equation 1})$$

Where:

- A(z) = nuclide activity at depth
- A₀ = surface nuclide activity
- V = advection velocity
- λ = nuclide decay
- D = diffusion coefficient

The advection velocity is generally determined by the placement of the subsurface peak in Cs-137, however we were not able to determine clear peaks in subsurface. This peak corresponds with the peak deposition time of 1964 from thermonuclear testing during the Cold War (citation). Since we could not find clear subsurface peaks, in this scenario, (V) was assumed to be zero. Inventories at depth were converted to percent-inventory profiles by dividing the activity at a certain depth by the total inventory for that

core. Then, a best-fit curve was modeled by minimizing the sum of residuals and a diffusion coefficient (D) is calculated, giving us an idea of the intensity of diffusion like soil mixing.

Table A6. Fallout Radionuclide data for Control plot soil cores.

Sample		Activity (Bq/cm ²)				Inventory (Bq/cm ³)			
	Depth	²¹⁰ Pb	¹³⁷ Cs	²¹⁰ Pb-excess	²¹⁰ Pb-excess (alt)	²¹⁰ Pb	¹³⁷ Cs	²¹⁰ Pb-excess	²¹⁰ Pb-excess (alt)
C1									
C1a1	0.75	0.20	0.01	0.09	0.16	0.29	0.02	0.14	0.23
C1a2	2.25	0.07	0.01	0.00	0.03	0.11	0.02	0.00	0.05
C1a3	3.75	0.08	0.02	0.00	0.04	0.12	0.02	0.00	0.06
C1a4	5.25	0.40	0.05	0.30	0.36	0.60	0.07	0.45	0.54
C1a5	6.75	0.08	0.03	0.00	0.04	0.12	0.04	0.00	0.06
C1a6	8.25	0.08	0.02	0.00	0.04	0.12	0.03	0.00	0.06
C1a7	9.75	0.06	0.02	0.00	0.02	0.09	0.03	0.00	0.03
C1a8	11.25	0.06	0.02	0.00	0.02	0.09	0.03	0.00	0.03
C1a9	12.75	0.06	0.01	0.00	0.02	0.08	0.01	0.00	0.02
C1a10	14.25	0.04	0.00	0.00	0.00	0.05	0.01	0.00	0.00
C1a1i	16.25	0.04	0.00	0.00	0.00	0.10	0.00	0.00	0.00
C1a2i	18.75	0.05	0.00	0.00	0.01	0.11	0.00	0.00	0.01
C1b1	20.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b2	22.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b3	23.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C1b4	25.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b5	26.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b6	28.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b7	29.75	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C1b8	31.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C1b9	32.75	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
C1b10	34.25	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
C2									
C2a1	0.75	1.04	0.03	1.04	1.00	1.56	0.04	1.56	1.50
C2a2	2.25	0.28	0.07	0.09	0.24	0.42	0.11	0.14	0.36
C2a3	3.75	0.05	0.03	0.00	0.01	0.07	0.04	0.00	0.01
C2a4	5.25	0.05	0.01	0.00	0.01	0.08	0.02	0.00	0.02
C2a5	6.75	0.04	0.00	0.00	0.00	0.07	0.01	0.00	0.01
C2a6	8.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2a7	9.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2a8	11.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2a9	12.75	0.05	0.00	0.00	0.01	0.08	0.00	0.00	0.02
C2a10	14.25	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
C2a.1i	16.25	0.03	0.00	0.00	0.00	0.09	0.00	0.00	0.00
C2a.2i	18.75	0.04	0.00	0.00	0.00	0.11	0.00	0.00	0.01

C2b1	20.75	0.04	0.00	0.00	0.00	0.07	0.00	0.00	0.01
C2b2	22.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2b3	23.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C2b4	25.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C2b5	26.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C2b6	28.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C2b7	29.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
C2b8	31.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2b9	32.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C2b10	34.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C3									
C3a1	0.75	1.45	0.01	1.45	1.41	2.17	0.01	2.17	2.11
C3a2	2.25	0.52	0.02	0.42	0.48	0.77	0.02	0.63	0.71
C3a3	3.75	0.11	0.01	0.01	0.07	0.17	0.02	0.02	0.11
C3a4	5.25	0.06	0.01	0.00	0.02	0.10	0.02	0.00	0.04
C3a5	6.75	0.05	0.01	0.00	0.01	0.08	0.02	0.00	0.02
C3a6	8.25	0.06	0.02	0.00	0.02	0.09	0.02	0.00	0.03
C3a7	9.75	0.06	0.01	0.00	0.02	0.09	0.02	0.00	0.03
C3a8	11.25	0.06	0.01	0.00	0.02	0.09	0.02	0.00	0.03
C3a9	12.75	0.06	0.01	0.00	0.02	0.09	0.02	0.00	0.03
C3a10	14.25	0.06	0.01	0.00	0.02	0.09	0.01	0.00	0.03
C3a1i	16.25	0.06	0.01	0.00	0.02	0.15	0.02	0.00	0.05
C3a2i	18.75	0.04	0.00	0.00	0.00	0.10	0.01	0.00	0.00
C3b1	20.75	0.07	0.01	0.00	0.03	0.10	0.02	0.00	0.04
C3b2	22.25	0.05	0.01	0.00	0.01	0.07	0.01	0.00	0.01
C3b3	23.75	0.04	0.00	0.00	0.00	0.06	0.01	0.00	0.00
C3b4	25.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
C3b5	26.75	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
C3b6	28.25	0.05	0.00	0.00	0.01	0.07	0.00	0.00	0.01

[illegible]

L2b7	29.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L2b8	31.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L2b9	32.75	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
L2b10	34.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
SM2x-3									
L3a1	0.75	0.10	0.02	0.03	0.06	0.15	0.03	0.04	0.09
L3a2	2.25	0.17	0.02	0.09	0.13	0.25	0.03	0.14	0.19
L3a3	3.75	0.17	0.02	0.07	0.13	0.25	0.03	0.10	0.19
L3a4	5.25	0.12	0.02	0.04	0.08	0.18	0.03	0.05	0.12
L3a5	6.75	0.17	0.02	0.07	0.13	0.25	0.03	0.10	0.19
L3a6	8.25	0.07	0.02	0.00	0.03	0.11	0.02	0.00	0.05
L3a7	9.75	0.07	0.02	0.00	0.03	0.10	0.02	0.00	0.04
L3a8	11.25	0.06	0.02	0.00	0.02	0.10	0.03	0.00	0.04
L3a9	12.75	0.05	0.01	0.00	0.01	0.08	0.02	0.00	0.02
L3a10	14.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L3a1i	16.25	0.04	0.00	0.00	0.00	0.09	0.00	0.00	0.00
L3a2i	18.75	0.03	0.00	0.00	0.00	0.08	0.00	0.00	0.00
L3b1	20.75	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
L3b2	22.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
L3b3	23.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
L3b4	25.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L3b5	26.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L3b6	28.25	0.04	0.00	0.00	0.00	0.07	0.00	0.00	0.01
L3b7	29.75	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
L3b8	31.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L3b9	32.75	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
L3b10	34.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00

Table A8. Fallout Radionuclide data for SM4x plot soil cores.

Sample		Activity (Bq/cm ²)				Inventory (Bq/cm ³)			
	Depth	²¹⁰ Pb	¹³⁷ Cs	²¹⁰ Pb-excess	²¹⁰ Pb-excess (alt)	²¹⁰ Pb	¹³⁷ Cs	²¹⁰ Pb-excess	²¹⁰ Pb-excess (alt)
SM4x-1									
H1a1	0.75	0.23	0.02	0.12	0.19	0.35	0.03	0.18	0.29
H1a2	2.25	0.16	0.02	0.05	0.12	0.24	0.02	0.08	0.18
H1a3	3.75	0.14	0.01	0.01	0.10	0.21	0.02	0.01	0.15
H1a4	5.25	0.12	0.01	0.04	0.08	0.19	0.02	0.06	0.13
H1a5	6.75	0.12	0.01	0.01	0.08	0.18	0.02	0.02	0.12
H1a6	8.25	0.12	0.02	0.02	0.08	0.17	0.02	0.03	0.11
H1a7	9.75	0.09	0.01	0.01	0.05	0.13	0.02	0.01	0.07
H1a8	11.25	0.04	0.01	0.00	0.00	0.05	0.01	0.00	0.00
H1a9	12.75	0.08	0.01	0.00	0.04	0.11	0.02	0.00	0.05
H1a10	14.25	0.05	0.01	0.00	0.01	0.07	0.01	0.00	0.01
H1a1i	16.25	0.04	0.00	0.00	0.00	0.11	0.01	0.00	0.01
H1a2i	18.75	0.03	0.00	0.00	0.00	0.08	0.00	0.00	0.00
H1b1	20.75	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b2	22.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b3	23.75	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b4	25.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b5	26.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b6	28.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H1b7	29.75	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
H1b8	31.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
SM4x-2									
H2a1	0.75	0.14	0.02	0.00	0.10	0.20	0.03	0.00	0.14
H2a2	2.25	0.13	0.02	0.04	0.09	0.19	0.02	0.06	0.13
H2a3	3.75	0.21	0.01	0.09	0.17	0.32	0.02	0.14	0.26
H2a4	5.25	0.33	0.02	0.20	0.29	0.50	0.02	0.30	0.44
H2a5	6.75	0.14	0.02	0.05	0.10	0.22	0.03	0.08	0.16
H2a6	8.25	0.12	0.02	0.00	0.08	0.18	0.03	0.00	0.12
H2a7	9.75	0.10	0.02	0.00	0.06	0.15	0.03	0.00	0.09
H2a8	11.25	0.09	0.03	0.00	0.05	0.14	0.04	0.00	0.08
H2a9	12.75	0.07	0.01	0.00	0.03	0.10	0.02	0.00	0.04
H2a10	14.25	0.05	0.00	0.00	0.01	0.07	0.01	0.00	0.01
H2a1i	16.25	0.06	0.00	0.00	0.02	0.15	0.01	0.00	0.05
H2a2i	18.75	0.06	0.00	0.00	0.02	0.16	0.01	0.00	0.06
H2b1	20.75	0.04	0.00	0.00	0.00	0.07	0.00	0.00	0.01
H2b2	22.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H2b3	23.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00

SM4x- 3									
H3a1	0.75	0.40	0.02	0.31	0.36	0.60	0.03	0.47	0.54
H3a2	2.25	0.25	0.02	0.13	0.21	0.37	0.02	0.19	0.31
H3a3	3.75	0.18	0.02	0.08	0.14	0.27	0.02	0.12	0.21
H3a4	5.25	0.19	0.02	0.06	0.15	0.29	0.03	0.10	0.23
H3a5	6.75	0.18	0.02	0.07	0.14	0.27	0.03	0.10	0.21
H3a6	8.25	0.13	0.02	0.03	0.09	0.20	0.03	0.05	0.14
H3a7	9.75	0.09	0.02	0.00	0.05	0.14	0.02	0.00	0.08
H3a8	11.25	0.06	0.01	0.00	0.02	0.09	0.01	0.00	0.03
H3a9	12.75	0.05	0.01	0.00	0.01	0.08	0.01	0.00	0.02
H3a10	14.25	0.06	0.00	0.00	0.02	0.09	0.01	0.00	0.03
H3a1i	16.25	0.05	0.00	0.00	0.01	0.12	0.01	0.00	0.02
H3a2i	18.75	0.04	0.00	0.00	0.00	0.09	0.00	0.00	0.00
H3b1	20.75	0.05	0.00	0.00	0.01	0.07	0.00	0.00	0.01
H3b2	22.25	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
H3b3	23.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
H3b4	25.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H3b5	26.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
H3b6	28.25	0.03	0.00	0.00	0.00	0.04	0.00	0.00	0.00
H3b7	29.75	0.04	0.00	0.00	0.00	0.06	0.00	0.00	0.00
H3b8	31.25	0.04	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H3b9	32.75	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00
H3b10	34.25	0.03	0.00	0.00	0.00	0.05	0.00	0.00	0.00

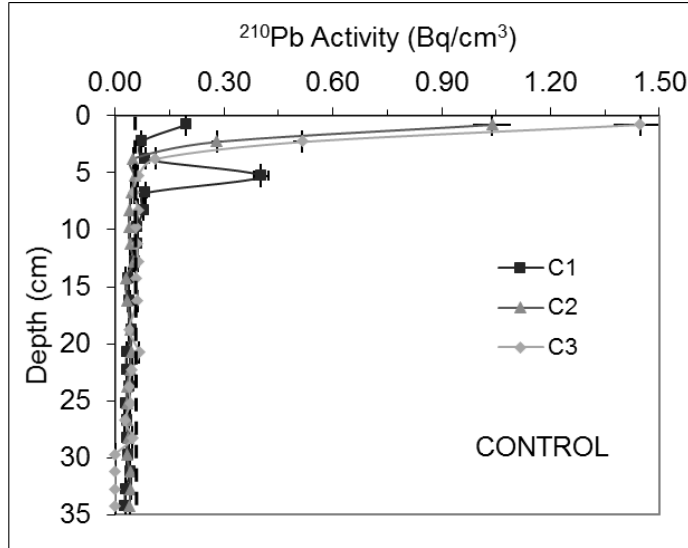
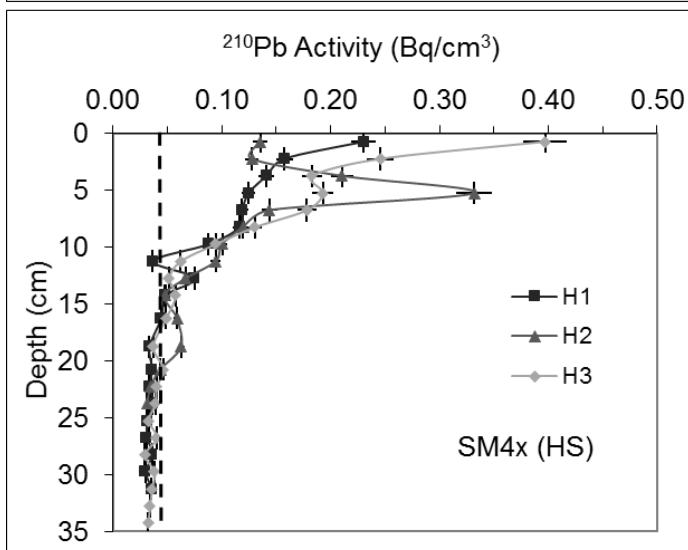
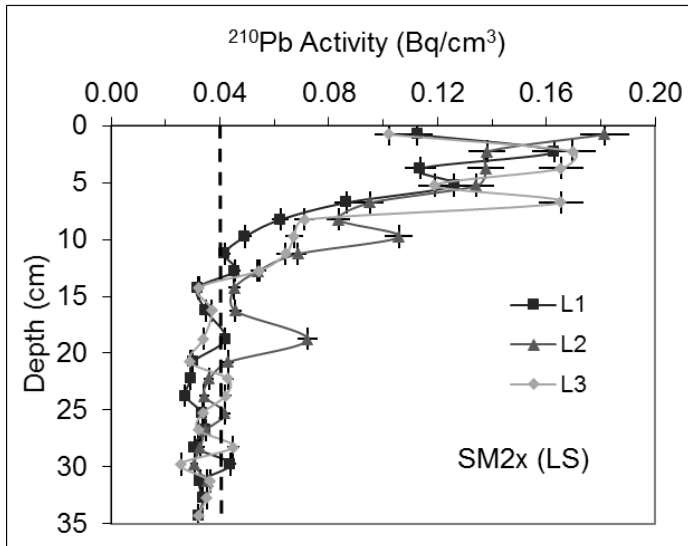


Figure A3. Combined profile plots from the control, SM2x, and SM4x conditions. Core C1 showed evidence of anthropogenic disturbance, with a near surface reduction to the detection limits of our BEGe detectors. This suggested that “dead” soil was deposited on top of the surface, most likely by shoveling during the implementation of various instruments or field site maintenance.



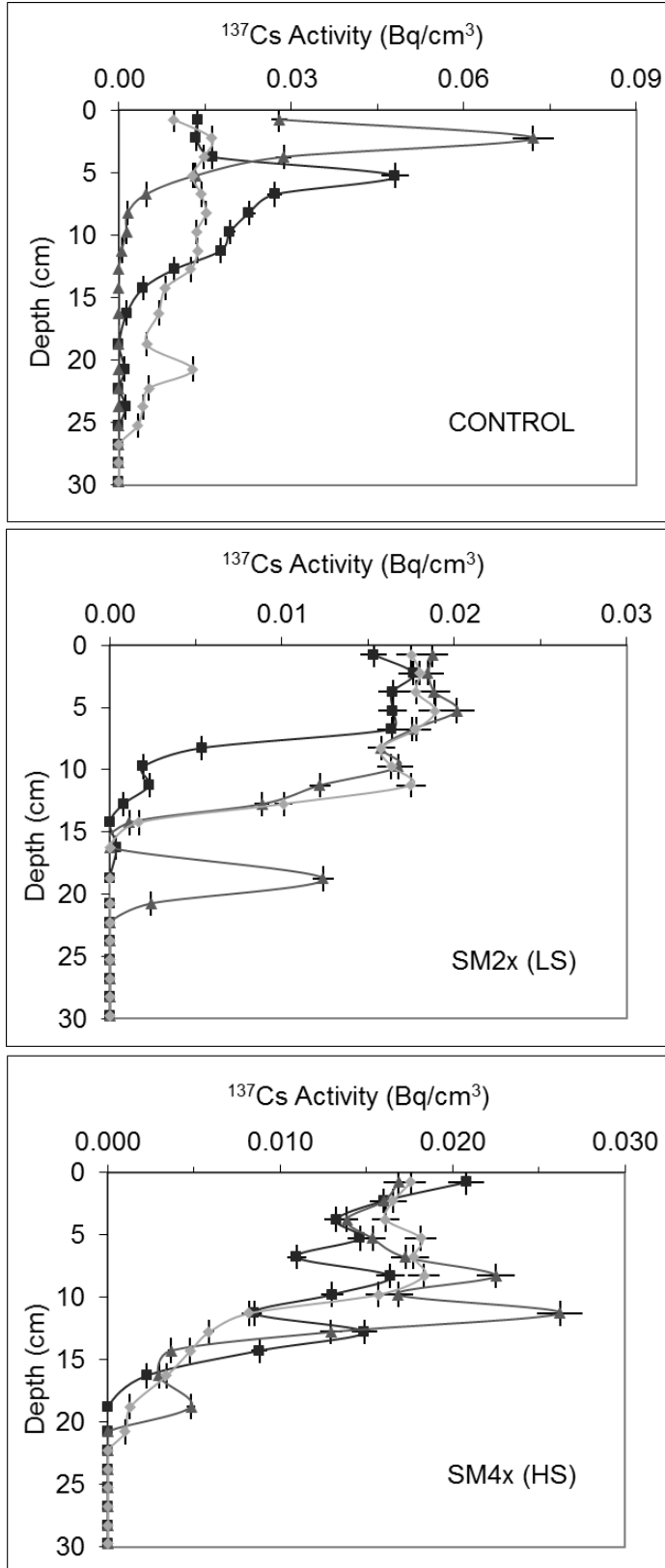


Figure A4. Combined profile plots from the control, SM2x, and SM4x conditions showing ^{137}Cs activity.

References Cited

- Dixon, J. L., Heimsath, A. M., and Amundson, R., 2009, The critical role of climate and saprolite weathering in landscape evolution: *Earth Surface Processes and Landforms*, v. 34, no. 11, p. 1507-1521.
- Kaste, J. M., Heimsath, A. M., and Bostick, B. C., 2007, Short-term soil mixing quantified with fallout radionuclides: *Geology*, v. 35, no. 3, p. 243-246.
- He, Q., and Walling, D., 1996, USE OF FALLOUT Pb-210 MEASUREMENTS TO INVESTIGATE LONGER-TERM RATES AND PATTERNS OF OVERBANK SEDIMENT DEPOSITION ON THE FLOODPLAINS OF LOWLAND RIVERS: *Earth Surface Processes and Landforms*, v. 21, no. 2, p. 141-154.
- Ligero, R., Barrera, M., Casas-Ruiz, M., Sales, D., and López-Aguayo, F., 2002, Dating of marine sediments and time evolution of heavy metal concentrations in the Bay of Cádiz, Spain: *Environmental Pollution*, v. 118, no. 1, p. 97-108.
- Tyler, A. N., Carter, S., Davidson, D. A., Long, D. J., and Tipping, R., 2001, The extent and significance of bioturbation on ¹³⁷Cs distributions in upland soils: *Catena*, v. 43, no. 2, p. 81-99.

APPENDIX E

METHOD FOR pH ANALYSIS

The same three soil cores used for total C, N, and particle size analysis were used to measure pH changes among snowpack plots. A total of six samples from each core were measured to produce pH depth profiles. For this analysis, only unpulverized material was used due to the limited supply of milled, homogenized material. Each sample was placed into a 1:2 soil DI water mixture and stirred with a magnetic stir bar until all sediments were in suspension and a mixed slurry formed. A VMR symPHony pH probe was placed into this slurry until a stable pH value was measured. We took 3 measurements per sample and calculated average sample pH from these values. Error bars in Figure D were calculated by running three separate mixtures of the same sample.

Table A9. Summative soil pH data.

Soil pH			Depth
Control	SM2x	SM4x	
5.65	5.39	5.35	3.75
5.49	5.56	5.29	6.75
5.63	5.36	5.27	14.25
5.63	5.60	5.59	26.75
5.66	5.62	5.69	31.25
5.70	5.59	5.75	34.25