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Origin of eroded soil organic matter in low-order catchments of the southern Sierra Nevada

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Publication Date
2014

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Origin of eroded soil organic matter in low-order catchments of the southern Sierra Nevada

A Thesis submitted in partial satisfaction of the requirements for the degree of Master of Science

in

Environmental Systems

by

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University of California, Merced
2014
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ACKNOWLEDGEMENTS

Many people contributed to the completion of this work, most notably my advisors, Drs. Asmeret Asefaw Berhe and Stephen C. Hart. Other mentors including Drs. Carolyn Hunsaker, Marilyn Fogel, and Karis McFarlane all guided me during the course of this work. Funding was provided by NSF through the Southern Sierra CZO and UC Merced awarded to Drs. Asmeret Asefaw Berhe and Stephen C. Hart.

Field and lab work would not have been possible without the hard work of many friends and colleagues including Rebecca Lever, Erin Stacy, Chelsea Carey, Abby Dziegiel, Nick Marlowe, and Grant Ivison-Lane. The staff of the U.S. Forest Service was amazing to work with and provided the sediment samples used in the study. The tireless effort of my non-academic support network of family and friends was invaluable. Thank you to everyone who helped in any way in this research.
INTRODUCTION

The carbon (C) cycle is currently experiencing perturbations largely due to human activities. Anthropogenic activities, such as logging and transportation emissions, are releasing large amounts of carbon dioxide (CO$_2$) into the atmosphere. In addition to CO$_2$, other greenhouse gases such as methane and nitrous oxide are contributing to global climate shifts. Anticipated changes in air temperature and precipitation linked to the greenhouse effect are expected to influence rates of biological processes (Kim et al. 2013). However, there are major uncertainties in our identification and quantification of fluxes in the global C cycle. For example, the role that lateral transport (i.e., erosion) of topsoil plays in redistributing C on the landscape and the fate of the eroded soil organic matter (SOM) are currently poorly understood (Van Hemelryck et al. 2010, Van Oost et al. 2012). Soil erosion can lead to changes in the rate of organic matter (OM) decomposition, depending on the burial rates of eroded topsoil downslope and photosynthetic recharge of the soil C pool from eroded sites at the hillslope scale. Soil erosion controls soil C concentrations and composition across different settings in eroding landscapes, which affects local C stocks and can influence global cycling of C (Nadeu et al. 2012).

BACKGROUND

Soils as a carbon reservoir

Globally, soils contain more than 2500 petagrams (Pg = 10$^{15}$g) of carbon (C) in the top 3 m, with over half of that contained within the top 1 m (Tarnocai et al. 2009). Factors of soil formation (parent material, climate, topography, biota, and time) and the resultant soil physical and chemical properties determine the amount of C that soil in any given ecosystem can store (Six et al. 2006, Condron et al. 2010). The amount of C stored in soil is a balance between C input through the processes of photosynthesis, and deposition of eroded material (from water and wind erosion or dust input) and C output through decomposition, leaching, and erosion (Hu et al. 1999).

Carbon storage in soil is critical for regulation of the concentrations of carbon dioxide (CO$_2$) and other greenhouse gases (GHGs) in the atmosphere. Given the large amount of C stored in the soil, even small perturbations in the physical system due to changes in climate, land use, erosion, and fire, can have a profound impact on concentrations of GHGs in the atmosphere. Furthermore, physical perturbations can affect the stock, composition, stability, and stabilization mechanisms of C in soil, leading to changes in the persistence and fate of soil C (Berhe et al. 2007, Van Hemelryck et al. 2010, Van Oost et al. 2012, Berhe and Kleber 2013). The role of physical perturbations in the environment affecting soil C dynamics will become even more important in the future with global climate change. Anticipated changes include precipitation amount and intensity, as well as temperature, with resultant implications for changing rates of soil OM input and output from the soil system (Kim et al. 2013, Berhe et al. 2014).
Erosion and soil carbon dynamics

Erosion is the lateral redistribution of soil from one area to another as a result of water, wind (Lal 2003), or gravity driven mass transport (Berhe and Kleber, 2013). Because the Earth's land surface is dominated by sloping landscapes, processes of soil erosion are ubiquitous. However, the natural rate of soil erosion can increase significantly due to anthropogenic influences, such as land use change, logging, and/or fire (Montgomery 2007). Currently, soil erosion laterally distributes on the order of 30-100 Pg of soil each year (Stallard 1998) containing 1-5 Pg of C (Berhe et al. 2007). Of this transported material, 70-90% is deposited in the same or nearby watersheds (Berhe et al. 2012). As long as the eroded C is replaced by photosynthesis and some of the eroded SOM is also stabilized in depositional settings, erosion can induce a terrestrial sink for CO₂ with respect to the atmosphere (Berhe et al. 2007). This erosion-induced terrestrial C sequestration can potentially account for up to 75% of the estimated “missing sink” or “residual terrestrial C sink.” However, until recently, the role of soil erosion in the global C cycle was not considered in C models and budgets.

Previously, there has been some disagreement in the literature over the validity of an erosion induced terrestrial sink. The prevailing paradigm that soil erosion can only constitute a CO₂ source to the atmosphere, due to the presumed large fluxes of CO₂ to the atmosphere from decomposition of C in topsoil during and after erosional transport (Lal 2004, 2005a, b, Pimentel 2006). However, other studies (Stallard 1998, Harden et al. 1999, Berhe et al. 2007, Berhe et al. 2008, Harden et al. 2008, Dymond 2010, Berhe et al. 2012) have shown that erosion can constitute a C sink if the criteria are met (Berhe et al. 2007).

Controls on erosional transport of topsoil and associated organic matter

The rate of soil erosion and the fate of eroded SOM in any given catchment is a function of a range of environmental variables, such as climate, topography, disturbance, vegetative cover, and lithology. Climate is typically a primary factor that controls rates of soil erosion because the type, amount, intensity, timing, and frequency of precipitation, as well as temperature and sun exposure, control the amount of energy that is available to mobilize soil material and the ease with which the soil can be transported (Pimentel 2006, Fu et al. 2010, Cantón et al. 2011, Nadeu et al. 2012, Kim et al. 2013). Water erosion mainly occurs first by detachment of aggregates due to impact (kinetic energy) of rain or hail, and afterwards detached topsoil material is transported with runoff and ultimately deposited in downslope settings. The travel distance of mobilized topsoil and associated soil organic carbon (SOC) is determined by the intensity of the precipitation, but generally most of the material remains near (< 1.5 m) to where it was detached (Chaplot and Poesen 2012). The same variables that affect the rate of erosion also influence the type and growth patterns of vegetation, soil development, and physical stability and chemical composition of soil (Kim et al. 2013).
The type of vegetation cover largely determines the quantity and quality of OM input to the soil. In turn, the amount and composition of the OM influences a suite of soil physical and chemical properties, including aggregation, C and N stocks in soil, and cation exchange capacity (Cantón et al. 2011). In conjunction with the above variables, geomorphology of the landscape (including slope steepness and length, size of a catchment, spatial heterogeneity, and soil physical properties) affect the amount of sediment that can be redistributed by erosion (Cantón et al. 2011). Slope, topography, and soil texture govern how prone a soil is to erosion (erodibility), as they determine the ease with which sediment can be mobilized on a hillslope and the rate of water flow in soil (Berhe 2012).

Reduction of vegetation cover due to land cover change generally leads to destabilization of soil and can alter the soil system irrevocably. Removal of vegetation causes death and decay of roots that provide stability and the continuous input of OM in the form of litter and exudates, which help bind the soil and provide structure (Torn et al. 2009). Loss of aboveground vegetation also removes the protective vegetation that serves to slow down the energy with which raindrops impact the ground surface. The type of bedrock the soil is derived from is also important, as soils derived from some parent materials (for example poorly consolidated sediments) tend to be more prone to erosion (Kögel-Knabner et al. 2008, Bellin et al. 2011). All of the above variables that influence rate of soil erosion also influence soil physical and chemical conditions (including texture and mineralogy), plant productivity, and the amount and composition of OM in soil (Berhe et al. 2014).

### Stabilization mechanisms of OM

Stabilization mechanisms of SOM determine how long C is stored in the soil. Soil OM is protected from loss, or stabilized, through a combination of physical and chemical mechanisms (Goh 2004, Berhe et al. 2012, Dungait et al. 2012). Aggregates physically protect SOM through encapsulation that limit access of SOM by microbes (Dungait et al. 2012), along with reducing diffusion of oxygen, water and extracellular enzymes into SOM (Berhe and Kleber 2013). Aggregates can become destabilized with erosion, increasing the availability of C for decomposition. A range of soil properties, such as clay content, availability of particle microsites, OM composition, and strength of bonds between minerals and OM, dictate the strength of physical and chemical stabilization mechanisms of OM. The stability determines the amount of OM that is available for mineralization as a result of microbially mediated processes (Kögel-Knabner et al. 2008, Torn et al. 2009, Keil and Mayer 2013). The above listed factors that influence soil development and erosion also influence the formation of aggregates. The aggregates protect C from loss from oxidative decomposition by determining the availability of reactive surface area and microsites where bonding between C and mineral surfaces can occur (Goh 2004, Berhe 2012).

Up to 90% of C in soil exists as C chemically bound to soil mineral particles (Berhe et al. 2012). The clay content and mineralogy of soil, especially concentration of metal oxides, have a strong effect on stabilization of OM in soils. These small sized particles
have high surface area to volume ratio and possess effective surfaces for bonding of organic functional groups to soil minerals. Iron (Fe) and aluminum (Al) oxides have surface charge that depends on soil pH. The pH at which they have no net charge is called the Point of Zero Charge (PZC, about pH 7.8 for some oxides). Below their PZC, oxides are capable of directly bonding with negatively charged organic compounds through formation of ionic or covalent bonds. This can lead to effective stabilization of SOM over decadal to century timescales (Kögel-Knabner et al. 2008, Keil and Mayer 2013). These bonds are the backbone of aggregate creation and enable the physical and chemical protection of C from decomposition.

**Isotopes**

Stable and radioactive isotopes (e.g., \(^{14}\)N, \(^{13}\)C, and \(^{14}\)C) are useful ecological tracers for photosynthetic pathways of plants, food webs in ecosystems, and for determining sources of water, soil, and OM in ecosystems (Dawson et al. 2002, Kuzyakov 2010). Isotopes of any given atom have different numbers of neutrons in their nucleus leading to changes atomic mass. Some isotopes are radioactive or unstable, meaning they decay into other elements by the emission of particles or gamma rays. Other isotopes are stable and maintain their atomic structure over time.

Typically, in order to identify sources of OM or other compounds in any given reservoir, the stable isotope composition of a given material is analyzed along with potential sources that may have at least partially contributed to the material in the reservoir. At natural abundance level, the isotope composition of a given substance is given in terms of delta (\(\delta\)) notation by comparing the sample with a known standard:

\[
\delta(\%o) = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000
\]

where \(\delta\) of the isotope (in per mille or \(\%o\)) is the mole fraction ratio of the heavy to light isotope of the sample (\(R_{\text{sample}}\)) divided by the same ratio in the standard (\(R_{\text{standard}}\)) minus 1. The value is multiplied by 1000 to magnify the small differences in isotope ratios among different natural substances. Isotope ratios are found in the atmosphere and biosphere for different isotopes due to preferential use (or loss from a reservoir) of lighter isotopes by biotic and other reactions that involve phase transformations (Fogel and Cifuentes 1993). A positive value indicates more of the heavy isotope compared to the standard and a value that is negative indicates more of the light isotope as compared to the standard (Sulzman 2007).

**Using \(^{13}\)C and \(^{15}\)N isotopes to trace sources of SOM**

Even though stable isotopes (e.g., \(^{13}\)C, \(^{17}\)O, and \(^{2}\)H) are regularly used in ecology and biogeochemistry to trace sources of OM, determine OM turnover, and to indicate pathways of water in soil, their use in soil erosion related studies has been relatively limited. Tracing sources of eroded topsoil material (by comparison with isotope compositions of likely source locations or reservoirs) is possible because the isotopic
values of eroded topsoil will largely reflect those of the constituents of topsoil in the source areas of the eroded material (Nadelhoffer and Fry 1988, Amundson et al. 2003, Meusburger et al. 2013) in a manner that is analogous to using stable isotopes of C and N to trace nutrients, and water source use (Nadelhoffer and Fry 1988, Vitousek et al. 1989, Hart and Classen 2003, Papanicolaou et al. 2003, Bird and Torn 2006, Fox and Papanicolaou 2007, Alewell et al. 2009, Schaub and Alewell 2009, Fox et al. 2010, Nadeu et al. 2012). Radioactive isotopes have traditionally been used to help trace soil movement by taking advantage of known radioisotope additions to the environment. Major contributors, such as atmospheric thermonuclear testing around 1960, the Chernobyl disaster or the recent accident at Fukushima Daiichi in Japan, have all released large amounts of radioactive materials to the environment (Schaub and Alewell 2009, Kim et al. 2013). The most commonly used isotopes in soil erosion work are radioisotopes such as $^{137}$Cs, $^7$Be or $^{10}$Be (Alewell et al. 2009, Meusburger et al. 2013). These radioisotopes are deposited from the atmosphere and typically uniformly distributed on the soil surface where they form strong chemical bonds with soil minerals. Hence, any observed changes over time in the concentrations of these isotopes in soil are attributed to lateral distribution of soil minerals due to erosion or deposition. Recently, other tracers, such as $^{13}$C and $^{15}$N, have been sought that have better integrity over time because of the short half-lives of many of the previously used elements (Alewell et al. 2008).

Differences in isotopic composition of SOM can be due to many factors, including the vegetation type and tissue that the organic matter is derived from, the microbial products that are synthesized during SOM formation, and preferential association of certain OM components with soil minerals (Dawson et al. 2002, Amundson et al. 2003). Much of the isotopic work done in soils has focused on using historic or experimental changes in vegetation type over time to characterize the rate of C or nitrogen (N) cycling in soil. The most commonly used stable isotopes for this work are $^{13}$C and $^{15}$N that show isotope ratios depending on the source of OM or soil, water use efficiency, and transformations associated with mineralization of OM (Fogel and Cifuentes 1993). Identifying differences in isotope compositions are possible because of the preferential use of lighter isotopes over the heavier ones (e.g., $^{14}$N over $^{15}$N and $^{12}$C over $^{13}$C) due to their weaker bonds and faster reaction rates, resulting in isotopic fractionation between the reactant and product (Kendall and Caldwell 1998). Plant material and, subsequently, OM derived from the three photosynthetic pathways exhibit distinct ranges in $\delta^{13}$C values (Wang et al. 1998, Sulzman 2007, Torn et al. 2009) associated with the degree of C isotope discrimination during C fixation. After the OM enters the soil system, microbial decomposition can further fractionation the C isotopes, resulting in successively higher $\delta^{13}$C values over time. Stable isotopes of N are also fractionated during decomposition of OM, resulting in progressively higher $\delta^{15}$N values of OM as the lighter $^{14}$N is converted preferentially from organic to inorganic forms (i.e., during N mineralization). Isotopic composition of C and N in soil is also influenced by additional factors, such as soil depth, texture, porosity, OM stock, type of vegetation, climate and land use changes (Kendall 1998).
Application of stable C and N isotopes to trace the movement of soil material has been successful in a few areas in both agricultural and minimally disturbed catchments (Fox and Papanicolaou 2007, Schaub and Alewell 2009, Nadeu et al. 2012). In each study, a depositional pool was identified (e.g., time-averaged or event-based sediments, wetlands) and soils collected from contributing hillslopes. In all studies the isotopic ratios of the depositional material reflected the upland material regardless of which depositional pool was used. These findings also showed erosion was driven by catchment characteristics, such as topography or soil type for time-averaged deposition, but event-based deposition was controlled by precipitation.

**Radiocarbon and residence time**

The average time an atom spends in a given reservoir is defined as its mean residence time (MRT). For C atoms in an ecosystem, their MRT can be determined using a radioisotope of C or radiocarbon ($^{14}$C). Radiocarbon ($^{14}$C) is a common cosmogenic radionuclide formed continually in the upper atmosphere when cosmic rays interact with dinitrogen gas. Once formed, $^{14}$C combines with oxygen to form $^{14}$CO$_2$ that is taken up by plants during photosynthesis, and gets incorporated into biomass further up the food chain (i.e., animals consume plants and animals consume other animals). When living things that incorporated the $^{14}$CO$_2$ or resultant biomass die, they stop exchanging $^{14}$C with the atmosphere. When living biomass dies and the remains enter the soil system, the $^{14}$C contained within the detritus begins to undergo radioactive decay to $^{14}$N at a constant rate (half-life of 5730 years), enabling determination of the time when the organism died and stopped exchanging $^{14}$C with the atmosphere (Trumbore 2009).

Radioactive isotopes of C (and other elements) have also been released into the atmosphere as a result of anthropogenic activities boosting the natural, background signal of $^{14}$C. For example, the 1960s atmospheric testing of thermonuclear weapons added massive amounts of radiocarbon to the atmosphere (Trumbore 2009), enabling scientists to be able to discern between “pre-bomb” or “post-bomb” SOM depending on how long the OM has been in the soil reservoir (Torn et al. 2009). This is similar to how one can use the half-life of $^{14}$C to determine when specific groups of C atoms entered the soil system. For example, the decay that occurs naturally leads to radiocarbon depleted soils deeper in the profile (i.e., older, more radioactive decay) than the shallower soils that have recently fixed radiocarbon and are closer in $^{14}$C concentration to our current atmosphere (Trumbore 2009). The $^{14}$C profile in soil can then be used to determine the MRT of C in soils. Generally, the MRT of C atoms in soil increases with depth (i.e., $^{14}$C values decrease with depth, indicating increasing age of C; Connin et al, 1997; Torn et al, 2002; Trumbore, 2009).
Origin of eroded soil organic matter in low-order catchments of the southern Sierra Nevada

ABSTRACT

Topsoil and associated soil organic matter are continuously laterally distributed on the landscape by the process of soil erosion. Soil erosion has been shown to lead to a sink of atmospheric carbon dioxide in soil if at least some of the eroded carbon is replaced by production of new photosynthate and some of the eroded carbon is stabilized in downslope depositional landform positions. However, until recently most of the work in this area has been focused on agricultural or grassland systems. Here, we present results from temperate forested catchments in the southern part of the Sierra Nevada Mountains. We found that most of the soil organic matter eroded from low-order catchments in the western slope of the Sierra Nevada Mountains is composed of forest floor material that has high concentrations of carbon. The steep slopes of these catchments also contribute to export of large proportion of the eroded forest floor material out of the catchments. Our radiocarbon analyses showed that the soil organic matter in the eroded material is composed of modern (post-1950) carbon with fraction modern values at or above 1.0. We also found that neither elevation, nor climate (across six years that we investigated) led to significant changes in the composition and mean residence times of the eroded material, despite considerable differences in mean residence time of soil organic matter in soil profiles of the contributing hillslopes in the high vs. low elevation (low elevation: 1800 m, high elevation: 2300 m) catchments we studied. Our findings suggest that soil organic matter eroded from upland forested catchments was substantially more susceptible to decomposition, compared to organic matter eroded from agricultural or grassland systems, leading to a less significant role of soil erosion on terrestrial carbon sequestration in forested catchments.
INTRODUCTION

Quantifying the fluxes of carbon (C) in terrestrial ecosystems is critical for understanding dynamic processes in the global C cycle and possible future effects of soil C storage on the global climate system (Falkowski 2000). Soil is the third largest C reservoir (2500 Pg) and is nearly double what is contained in terrestrial living biomass (560 Pg) and the atmosphere (750 Pg) combined (Tarnocai et al. 2009, Condron et al. 2010, Dungait et al. 2012, Smith 2012). Much of the C held within soil exists as organic matter (OM), either free, bonded to mineral particles, or encapsulated in aggregates (De Baets et al. 2012). The amount and composition of soil organic carbon (SOC) found in eroded soils reflects the origins of the soil organic matter (SOM) itself that has been transported by erosion (Bellanger et al. 2004). Erosion drivers such as climate (specifically precipitation amount, distribution, and intensity), slope, land use or change, and lithology can lead to selective transport of OM leading to distinct amounts and composition of eroded C found in captured sediments (Nadeu et al. 2012, Kim et al. 2013). With a warming climate and therefore changing precipitation regime (i.e., from snow dominated to rain dominated), determining erosion patterns in montane ecosystems is critical for predicting changes to hillslope processes.

Source pools of SOM from different landform positions can regulate the stability of transported C by governing the availability of protective surfaces for stabilization of OM. Tracing the movement of soil C in OM from eroding landform positions to the ultimate depositional settings (sediments) within a catchment is essential for quantifying fate of laterally transported C on the landscape (Berhe and Kleber 2013). Transport and fate of C are important factors in an ecosystem and especially for the C cycle. Release and stabilization of C within the landscape influences where and how C is stored within the landscape (Berhe et al. 2014).

Stable isotope analysis is a potentially powerful tool for sourcing eroded material in natural and agricultural landscapes. Traditionally, radioactive and cosmogenic isotopes were used to estimate the rate and source pools of erosion. Radiogenic nuclides (i.e., $^{137}$Cs, $^7$Be) released from nuclear power plant incidents (i.e., Chernobyl, Fukushima) bind tightly to soil minerals and create a concentration of these isotopes in soil (Alewell et al. 2009, Kim et al. 2013). Soil movement can then be estimated by enrichment (i.e., deposition) or depletion (i.e., erosion) of these elements on soil profiles along a hillslope. However, short half-lives (e.g., 30 years for $^{137}$Cs) and a lack of reference sites (i.e., sites with no erosion or deposition) make using radiogenic nuclides unreliable over time. Also, the assumption that atmospheric fallout is uniform and substantial across the landscape is not necessarily robust. Lastly, the measurements are time consuming and expensive when compared to stable isotopes (Schaub and Alewell 2009).

Stable isotopes of C and nitrogen (N) are present and cycled in large quantities (relative to other elements) in all terrestrial ecosystems, making them more reliable than radiogenic nuclides for tracing pathways of erosion. Stable isotopes are present in ratios governed by biological and chemical processes that preferentially use lighter isotopes in reactions or discrimination resulting in fractionation. Fractionation creates an isotopic ratio (concentration of the light to heavy isotopes) that can become enriched or depleted.
in the heavier isotopes of both elements in plant material and consequently the SOM (O’Leary 1988, Hogberg 1997). These ratios are well defined in carbon because the greatest fractionation occurs during photosynthesis. Nitrogen, due to the many microbial transformations that occur, has many fractionations that influence the isotopic ratio (Fogel and Cifuentes 1993). Using the stable isotopes of C and N together create a more robust fingerprint of the source of SOM. Recently, $^{13}$C and $^{15}$N have been used together and separately as a tool for following the transport of soil materials driven by precipitation events in catchments (Bellanger et al. 2004, Fox and Papanicolaou 2007) and deposited in accumulated sediments or wetlands (Alewell et al. 2009, Nadeu et al. 2012). In addition, stable isotopes have been used to trace the sources of erosion occurring on catchment hillslopes (Hilton et al. 2012, Meusburger et al. 2013) and as SOM bound to stream gravels (Collins et al. 2013).

Radiocarbon ($^{14}$C) can also be used to determine SOM stabilization within an ecosystem. The amount of radiocarbon remaining in the SOM illustrates the time since the original material last exchanged CO$_2$ with the atmosphere (Trumbore 2009). The less radiocarbon present in the material the longer it has been in the soil system. Without erosion SOM becomes older with depth, but if erosional redistribution of SOM lead to removal from hillslopes and burial in downslope soil profiles, the observed radiocarbon depth profiles can become inverted (Torn et al. 2009).

The objectives of this study are to: 1) test the viability of using stable isotopes as an erosion tracer, 2) determine the origin of eroded SOM, and 3) infer fate of eroded SOM using radiocarbon. This study is unique in that we are using stable isotopes and radiocarbon to determine the sources of eroded SOM on a catchment scale. By studying two similar catchments along the rain-snow transition, we were able to determine if precipitation type has an effect on natural erosion rates. Using this technique in two low-order catchments of the Sierra Nevada, we identified the main sources of collected sediments and distribution of C on the landscape. By characterizing the type of material that constitutes the sediments and where/how long C resides, natural patterns of erosion were delineated in these catchments. To do this, we sampled upland soils, O horizon, and stream bank samples as source pools for captured sediment to run bulk analysis of $^{13}$C and $^{15}$N as our natural tracer with a subset run of radiocarbon to determine residence time of C.

METHODS

Site description

Field sampling was conducted in the Kings River Experimental Watersheds (KREW) located within the Sierra National Forest of California (Figure 1). The KREW project is run by the United States Forest Service (USFS) and was started in 2000. The project consists of eight low-order (i.e., headwater) catchments at high and low elevations (Bull: 2050 to 2490 m; Providence: 1485 to 2115 m.a.s.l.; (Hunsaker et al. 2007). The lower elevation sites are also co-located with the Southern Sierra Critical Zone Observatory (SSCZO; (Bales et al. 2011)). Each catchment has one of four treatments: control, prescribed burning, mechanical thinning, and a combination of thinning and burning.
(Hunsaker et al. 2007). We sampled prior to the prescribed burn treatment: surficial organic (O) horizons, mineral soils, and stream bank and bed material in two catchments in Providence and Bull (Table 1). Samples were collected once during fall 2012 for Providence and summer 2013 for Bull.

The study sites are part of a climosequence with all other factors such as parent material, vegetation, topography, and soil age being comparable (Hunsaker and Neary 2012). Parent material is granite or granodiorite with the weakly developed soils dominated by three soil families: Shaver (Providence); Cagwin, and Gerle-Cagwin (Bull). Vegetation type is predominantly Sierran mixed conifer at both sites with differences in dominant species with elevation. Sugar pine (Pinus lambertiana) occurs at both sites while Providence also has ponderosa (P. ponderosa) and Bull has Jeffrey pine (P. jeffreyi). White fir (Abies concolor) is present at both sites, but Bull is dominated by red fir (A. magnifica). Incense cedar (Libocedrus decurrens) is also abundant at both sites (Johnson et al. 2011). The catchments have a Mediterranean climate characterized by warm, dry summers and cool, wet winters. Differences in elevation between the two catchments result in contrast in temperature and precipitation regime. Mean (± standard deviation) of annual air temperatures range from 11.3 ± 0.8°C in Providence to 7.8 ± 1.4°C in Bull (mean of October 1 to September 30 (Water Year) 2004-2007; (Johnson et al. 2011). Both catchments receive similar amounts of precipitation during the year (1549 ± 547 mm for Providence and 1583 ± 507 mm for Bull), but the high elevation receives mainly snow due to lower temperatures whereas the low elevation is half rain and half snow (Climate and Hydrology Database Projects, www.fsl.orst.edu/climhy).

**Sampling design**

We established six spatially independent hillslope transects in the two catchments. We sampled prior to treatment implementation to evaluate sources of natural erosion. Soil samples were taken at each of three landform positions along each transect: crest, backslope, and toeslope/deposition (Figure 2). The landform positions encompass the path of eroded material moving downhill to the possible terrestrial depositional areas. Four of the Providence transects (P1-P4) and one of the Bull transects (B6) were established previously by Stacy (2012). We used the same site selection criteria as Stacy (2012) to establish additional hillslope transects within each catchment for a total of six transects. Transects were placed a minimum distance of 20 m from each other while avoiding crossing roads and streams. Bull had some old logging skid trails that were still visible on the landscape and were avoided because these features are not representative of the entire catchment. Transects were spread throughout both catchments to capture the spatial heterogeneity in the catchments. Each transect is oriented on a western facing slope with points selected in open areas free from any recreational or grazing disturbance that could be observed at time of establishment. Western facing aspects were chosen as they dominate both catchments. By placing the transects in open areas, we chose sites that would be likely sources of eroded sediment which could move downhill with water to the depositional areas.
Soil was sampled at each landform position using a 1-m² plot with three sample points chosen at random. The surficial O horizon was collected at each point within a 10 x 20 cm area and composited before soil coring. Mineral soil samples were taken at multiple depths until saprolite contact (0-10, 10-20, 20-40, 40-60, 60-80, 80-100, and 100-120 cm) using a 5 cm diameter closed bucket hand auger, and the samples collected at the three points were then composited by depth. If obstructions such as rocks or large roots impeded the auger at depths less than 40 cm, other points were chosen within 20 cm of the original point within the plot. After six attempts, soil successfully cored to the desired depth classes was retained and the deepest depth reached was recorded. In addition to soils, streambank and streambed materials were also collected to evaluate these locations as possible sources of eroded sediment. Sampling points were chosen along the one main perennial stream and branches in both catchments at the headwaters and any landscape features (i.e., large meadows) to just above the sediment basins. Stream bank samples were collected on a level surface within 10 cm of the exposed bank face of the stream to maintain soil profile continuity at all augered points. No O horizon was taken with stream bank samples as there was none to be taken due to living biomass present that would not be washed downstream. The stream bank mineral soils were hand augered only to 60 cm depth (0-10, 10-20, 20-40, and 40-60 cm) due to the shallower groundwater than in upland locations. Streambed material was collected perpendicular to the bank sampling point in the middle of the stream path. Bed material was collected in a 10x10 cm area to a depth of 5 cm using a flat trowel. The bed material depth was chosen to include only the amount of gravel that would freely move with the flowing stream.

In 2003 and 2004, sediment basins were constructed in a natural depositional area below the flumes of each catchment (Hunsaker et al. 2007). The area was deepened and enclosed with logs held in place by rebar at the outlet and both banks then covered with a thick rubber pond liner. The stream runs into the basin where the velocity of the water slows, allowing sediment to settle out of the water. The Forest Service empties sediment basins annually close to the end of the water year (August or September) by hand after diverting the stream. The wet sediment was weighed on a spring scale to the nearest kilogram and large woody debris (longer than 0.5 m) were collected on a tarpaulin. A representative sample of the large wood was air-dried and weighed. A subsample was taken from each bucket of wet sediment and composited to create one representative sample of the entire basin. Large woody debris was weighed wet and a subsample air dried to get the total dry weight. This sample was then mixed and subsampled for analysis in this study. The Forest Service research station in Fresno, CA, processes several subsamples from each basin to determine mineral and organic content and stores archives for each year (Hunsaker and Neary 2012). For this study, the samples for WY 2012 and 2013 were taken from the field site and stored at 4°C until analyzed, while sample years prior to that were taken from the air-dried Forest Service archives.
Analyses

Laboratory analyses

All samples were stored field moist at 4 °C in sealed plastic bags until processed within two weeks of collection. Upland mineral soils were sieved (2 mm) field-moist, and the > 2 mm fraction discarded. Sediment and stream bank samples were dried at 40 °C in a forced air oven to evaporate some of the water contained in the samples to facilitate sieving. After reaching a constant weight, these samples were sieved through a 2-mm mesh sieve and the > 2-mm fraction discarded. For the surficial O horizon, materials >1 cm in diameter were discarded, and the remaining material cut into 2-cm lengths to facilitate subsampling. The subsamples were then ground with a Wiley mill (Thomas Scientific, Swedesboro, NJ) to <420 µm.

Transect soil samples were measured for physical and chemical soil characteristics including gravimetric water content, bulk density, and pH. Stream bank samples were analyzed for bulk density and pH but not gravimetric water content since they were dried prior to processing. For gravimetric water content, upland mineral soil and unground forest floor samples were homogenized and approximately 10 g dried at 105°C and 70°C until constant weight, respectively. Bulk density cores were taken at all sampling points in summer 2013 from the top of the mineral soil (0 to 10 cm) using a rubber mallet and a metal 5 cm by 10 cm core. Upland and stream bank soils were measured for pH using an Orion Dual Star pH/ISE benchtop meter with an Orion Ross Sure-flow combination epoxy body electrode (Thermo Scientific, Waltham, MA) and an Orion Star ATC probe (Thermo Scientific, Waltham, MA) in a 1:2 soil to deionized water slurry. The pH of upland soils was measured on field moist soil, and the pH of stream bank soils was measured on air-dry soil. Bulk density and pH were analyzed for comparison to previous work on similar soils (Johnson et al. 2011, Stacy 2012).

Subsamples of all soils, and streambed material were left to air dry before being ground using a mortar and pestle for stable isotope analysis. Air dry O horizon samples were ground using a Wiley mill. Ground samples were run for $^{13}$C, $^{15}$N, and elemental C and N at the Geophysical Laboratory at the Carnegie Institute in Washington, D.C. on an NC 2500 elemental analyzer (Carlo Erba, Milan, Italy) connected to a Delta V Plus isotope ratio mass spectrometer (Thermo Scientific, Waltham, MA) via a Conflo III interface (Thermo Finnegan, San Jose, CA). Samples were run against an acetanilide standard with an analytical precision of ± 0.01‰ for $^{13}$C and $^{15}$N.

Radiocarbon analysis was performed on all sediments, but radiocarbon analysis could only be performed on one transect and one stream bank point in each catchment because of cost limitations. The transects and stream banks analyzed were chosen using $^{13}$C values considered representative of the catchments (i.e., similar to median values of all transects and bank points). These samples were used to augment the stable isotope results as age of SOM contained within the soils and sediments should relate to origin (e.g., shallower soils = younger soils). Samples were prepared according to Vogel et al. (1984) and run at the Center for Accelerated Mass Spectrometry at Lawrence Livermore National Laboratory on a 10 MV Model FN Tandem accelerator (High Voltage Engineering Corporation, Burlington, MA).
Statistical Analyses

All data is presented as mean ± standard error. Inferential statistics were not used in this study due to the nature of the data set. Many of the statistical tests that would ordinarily be used (e.g., ANOVA or t-tests) violated assumptions with the varying numbers of each sample type (e.g., sediment had an n = 5, O horizon had an n= 18, 0 - 10 cm mineral soil had an n = 6). There was also only one replicate of each catchment. Instead, means and standard errors were compared between catchments. The stable isotopes were compared using biplots to show spatial similarities between sources and the end pool.

RESULTS

Soil Characteristics

Mineral soils were slightly acidic with Bull (pH = 5.31 ± 0.05) being more acidic than Providence (5.94 ± 0.05). Samples from depositional landforms of both catchments were the most acidic at all depths (from 0.16 to 0.49 pH less than crest or backslope in both catchments; Table 2). Providence soils had higher gravimetric water contents averaged across depths (0.15 kg/kg ± 0.01) than Bull (0.10 kg/kg ± 0.01) at all landform positions and depths. Deposition sites were the wettest landform positions in both catchments by 0.10 kg/kg as compared to the crest and backslope of both catchments (Table 2). Gravimetric water content declined with depth at all landform positions in both catchments. Bulk density values at all depths were 0.3 Mg/m³ higher in Bull than Providence at all landform positions. Soils were wetter in Providence than in Bull but were most likely affected more by the timing of soil collection (Providence sampled in fall, Bull sampled in summer).

Elemental C and N

Concentrations of C in the 0-10 cm layer and the O horizon (Figure 3) increased from the crest to the deposition in both catchments. Carbon concentrations in Providence were higher than in Bull in the 0-10 cm layer at each landform position (higher at Providence by: 3.61% at crest, 0.04% at backslope, 0.03% at deposition). Values had little variability across transects at each catchment as shown by the standard error values. The bed and stream bank samples had about twice as much C in Bull than in Providence (Providence: 6.71% ± 2.62, 5.87% ± 1.10 and Bull: 13.97% ± 0.14, 16.49% ± 3.62 for streambed and stream bank, respectively; Figure 3).

Providence sediments overall had a higher C concentration than sediments from Bull. Sediment had the most difference among years more than between catchments with WY 2009 (17.85%) and 2013 (19.91%) having the highest C concentrations in Providence and Bull, respectively. The lowest C concentrations were WY 2011 (2.10%) and 2010 (2.06%) in Providence and Bull, respectively. Note there is no standard error associated with the sediment values since there was only one sediment basin per catchment per year unlike the soils that had multiple points were replicated within catchments.

In Providence, the carbon to nitrogen (C: N) mass ratios showed a similar pattern to the C concentrations of the surface soils at all landform positions. In Bull, there was
much lower C:N mass ratios for the stream bank and streambed in relation to the 0-10 cm mineral soils (Figure 4). Soils in the Bull catchment had lower N than Providence shown in the higher C:N ratios present (Providence: 24.31 ± 1.20, 24.29 ± 1.27, 29.36 ± 3.21 and Bull: 24.83 ± 2.42, 27.73 ± 3.01, 30.17 ± 4.97 for crest, backslope, deposition; Figure 4). The C:N ratio also showed greater similarities between the 0-10 cm mineral soils and sediments than in the O horizon (forest floor) in both catchments. Sediments had smaller ranges of C:N ratios in Bull than in Providence among all years (Figure 4).

$^{13}$C and $^{15}$N

Unique signatures were found in the stable isotope biplots among the different source pools and sediments analyzed (Figure 5). Values of upland soils (both surficial O and mineral horizons (0 – 10 cm)) and sediment years 2010, 2012, and 2013 were more enriched in both $^{13}$C and $^{15}$N by 1 to 2‰ at the higher elevation compared to the lower elevation. In the both catchments, there were no statistically significant differences based on confidence intervals between the means of $^{13}$C and $^{15}$N in the streambed and bank samples. The upland soils were more similar isotopically to the stream bank in Providence than in Bull. In both catchments, stream bank soils and bed material had intermediate values between the upland soils, the sediments and O horizon materials. Sediments varied among years in both catchments with no discernable pattern in Bull or Providence with water year.

In Providence, mineral soils from all landscape positions were similar isotopically (Figure 5). Organic horizons were grouped together because there was very little variation in stable isotope values between landscape positions or transects. Stream bank and bed materials of both catchments were similar to each other and tended to be lower in $^{13}$C than mineral soils. Sediments displayed considerable interannual variation in both $^{13}$C and $^{15}$N. Sediment isotopic signatures tended to be intermediate between O horizon and stream bed (Figure 5).

In Bull, surface mineral soils had some variation but were different from all other samples (Figure 5). Organic horizon values were again grouped together as there was little variation between landscape positions or transects. Stream bank and bed had very similar values and, like Providence, had lower $^{13}$C values than mineral soils. Sediments had less interannual variation than in Providence, but still were intermediate between the O horizon and stream bank and bed materials (Figure 5).

The $^{13}$C values showed enrichment with depth at all landform positions at both catchments (Figure 6). Values of $^{15}$N are more varied with depth at both sites than $^{13}$C at all landform positions (Figure 6). The enrichment of $^{13}$C and $^{15}$N with depth illustrates why only the values for surface soils were used in the biplot analysis; deeper soils do not appear to contribute to the sediments.

Radiocarbon

Radiocarbon data of the representative samples analyzed showed mostly modern material in Providence and a split between pre-modern and modern in Bull (Figure 7). Providence had more modern organic C in the soils than Bull. Shallow mineral soils (< 20 cm deep) had modern C at all landform positions in both catchments. Stream bank
samples within the Providence catchment were all modern except at the 10-20 cm depth, while only organic C from the upper 20-cm of the mineral soil were modern in the Bull catchment. Providence soils showed little change in SOC age with depth (Figure 7). In contrast, the SOC contained in Bull soils and stream bank samples became progressively older with depth (Figure 7). Bull soils and stream banks that we analyzed have longer residence times in soils at depths below 20 cm in backslope and stream bank locations, but for the deposition position this break occurs below 10 cm. All sediment samples were modern and had similar values in both catchments (Figure 8).

DISCUSSION

Using a combination of stable isotopes and radiocarbon is an effective method to source eroded SOM back to origin materials. Our stable isotope results clearly delineated the sources of eroded sediments, and were consistent with previous studies (Alewell et al. 2009, Schaub and Alewell 2009, Nadeu et al. 2012). The stable isotope technique was efficient and relatively easy to use compared to the more expensive radioisotopes (e.g., $^{137}$Cs or $^7$Be) usually used in erosion studies. Also, using stable isotopes negates the need for non-eroded reference sites or a half-life imposed time limit (Schaub and Alewell 2009). To overcome the high cost associated with radiocarbon, here we show that a small set of samples can be used effectively in conjunction with stable isotopes.

In our low-order Sierra catchments, we found that O horizon along with stream bank and bed materials contribute the most to eroded sediments. We concluded this based on the similarity of the stable isotope values between the O horizon, stream materials, and all years of the sediment from both catchments. Our findings build upon to previous work with sediment fences placed on hillslopes in these catchments. For example, Stafford (2011) found that established paths (e.g., perennial streams) had more material mobilized to sediment than overland flow. Much of the sediment transport occurs during snowmelt, which causes a large pulse or pulses of water to flow to and in the streams (Iida et al. 2012). Precipitation timing and intensity have been shown to have an important effect on the amount and type of eroded C that collects as sediment (Gomez et al. 2003). Sediment yield and amount of precipitation were positively correlated in all of the KREW catchments over a seven year period (Hunsaker and Neary 2012, Stacy 2012). Catchments with large climatic seasonal changes (i.e., warm, dry summers and cold, wet winters) and soils that have high water repellency can cause more material to move due to low initial soil moisture (Cantón et al. 2011, Kim et al. 2013). Our study sites exhibit both of these qualities and could explain the importance of surficial material moving downslope to the basins.

Radiocarbon values also substantiate the stable isotope results as the sediments are composed of modern material much like the surface soils. In an unstable landform position, the soil ages would likely be inverted or mixed. In our catchments, the Bull backslope values may indicate erosion where the top few layers have inverted ages (Gomez et al. 2010). The age of the SOM increases with depth at the backslope and deposition in the Bull catchment and is modern at least to a depth of 0.6 m in the Providence catchment. The soils in Providence may show the vestiges of logging activity...
or some other disturbance that would cause a mixing of the soils and SOM ages. Even with mixing at Providence, the stable isotopic signatures of deep mineral soils are more enriched in both heavier isotopes at all depths as compared to the sediments. Most soil profiles show an increase in $^{13}$C with depth as over time more decomposed or transformed material is moved down the profile as water infiltrates deeper (Nadelhoffer and Fry 1988, Bol et al. 1999, Bellanger et al. 2004, Schaub and Alewell 2009, Meusburger et al. 2013). The data indicate that in the Providence catchment, soil layers below 0.6 m have not been disturbed and may show increasing age with depth.

For erosion, the size and momentum of rain drops has been shown to correlate to the amount of soil detachment (Salles and Poesen 2000). One of the advantages of working in catchments along the rain-snow transition is the differences in precipitation regimes that work as a climate change gradient. The effect of precipitation differences that occurs at each site can be seen in the differences between isotopic and elemental values. Similar sediment isotopic values across all years in Bull indicate more consistent erosion patterns than in Providence (Figure 5). This is most likely due to the majority of precipitation falling as snow in Bull, which creates little movement to surface or deep material until spring snowmelt. Providence receives rain and snow events throughout the winter that have the ability to destabilize the soil surface through rain splash or stabilize with snow cover (Hunsaker and Neary 2012). Frequent flood events show more negative $^{13}$C values while large singular events are opposite (Gomez et al. 2010). Recurrent flood events can continually carry shallow (i.e., fresh, isotopically light) material, while singular large events can create gullies and move deeper material (i.e., older, isotopically heavy). Using the same principles, intense precipitation events yield material that is more enriched in heavier isotopes while less intense events will transport material more depleted in heavier isotopes (Bellanger et al. 2004). Changes to the precipitation regime with climate change (i.e., more rain, less snow at higher elevations) can change the type of precipitation and therefore the composition of eroded material. In the future, the Bull catchment can have the same varied isotopic values that Providence currently shows if the precipitation regime shifts with anticipated climate change.

The type of erosion affects the concentration of C in eroded soil. For instance, gullies can produce a large amount of sediment but the material contains very little C, and overland flow creates small amounts of high C content sediment (Gomez et al. 2010). Other types of erosion also have different nutrient contents. For example, landslides can mostly contain nutrient depleted bedrock (Hilton et al. 2008). In our catchments, similarities between the elemental C and N concentrations of sediments and mineral soils are a possible indicator of processing during transport or after deposition. We assume this because isotopically we identified the sediment SOM as originating from the O horizon, but the nutrient contents are very different. The O horizon has a much higher C concentration than the sediment which is closer to the mineral soils. The mineral soils have had SOM mineralized and leached which lowers the nutrient concentration. The sediment would need a more quick and direct route to the basin for the sediment C concentrations to match the original O horizon.

Differences in the C concentrations of each landform position and among sediment years were also observed. As expected, soils sampled in the deposition sites are more
saturated and have higher concentrations of C and N compared to crest and backslope positions. The stream banks also show this pattern as many of the higher elevation sampling points were still saturated from flowing streams and have higher C and C: N ratios compared to the lower elevation, drier stream sites. These wetter sites have the ability to store more C than the upslope well-drained soils due to the inhibition of decomposition (Schaub and Alewell 2009). Abiotic factors such as slope, aspect, and climate could all account for the lack of C accumulation in these catchments (Lal 2005a). Above- and below-ground productivity of plants could also influence turnover times and inputs to the soil (Chapin et al. 2009, Fu et al. 2010). Nutrient concentrations (C and N) found in sediments was higher in low water years (low water year ~1000 mm/yr, high water year ~2000 mm/yr) suggesting that the sediment basins were more effective at capturing smaller nutrient rich particles with less precipitation. Less precipitation caused the stream water to remain in the basin longer, resulting in more material settling into the basin (Stacy 2012).

While our findings are compelling, limitations do exist with the technique we used. The study catchments were dominated by C\textsubscript{3} vegetation meaning there was very little \textsuperscript{13}C isotopic separation between the different samples we analyzed. This technique would be more robust with a combination of photosynthetic pathways such as \text{C\textsubscript{4}} and \text{C\textsubscript{3}} to create larger isotopic separation (Fogel and Cifuentes 1993). Using annual collections sediments also creates some uncertainty as to the arrival and path of eroded material. Because each precipitation event moves some material, the timing and composition of sediment delivery could be spread out during the year or even across years. This uncertainty in connectivity of the slopes to the stream could create variability within the catchment and the basin.

Our results suggest that isotopic tracers are powerful tools for elucidating sources of eroded materials such as sediments. By combining isotope analyses (\textsuperscript{13}C, \textsuperscript{14}C, and \textsuperscript{15}N), we were able to determine the main sources of sediment of the catchments. Our findings build upon previous work in these catchments outlining sediment yield, sediment composition, and erosion patterns (Stafford 2011, Hunsaker and Neary 2012, Stacy 2012). The captured sediments were shown to be largely O horizon and stream material in origin through stable isotopes and augmented by radiocarbon age. Mineral soils and stream banks in the Bull catchment show an increase in age with depth. Providence catchment SOC is modern through much of the depth, but the stable isotope values do not support deep soil contributions to the sediments. These findings show that even with limitations such as a \text{C\textsubscript{3}} dominated catchment that our technique has been validated and could be utilized in other systems on a catchment scale.

**CONCLUSION**

This study shows that, in upland temperate forested ecosystems, the OM transported by erosion is mainly topsoil material that is mobilized by surface erosion, and to a smaller degree by channel erosion in established streams. Combined with previous work in the same study catchments, our results suggest that precipitation type has the largest influence on erosion. Our findings of distinct characteristics of SOM eroded from low vs.
high elevation catchment suggests that future changes to erosion patterns may occur in the snow-dominated catchment as the climate warms and more precipitation is delivered as rain rather than snow.

Furthermore, high, at least episodic, rates of topsoil and associated SOM redistribution (within source catchment) and export (out of source catchment) erosion in these types of upland temperate forests and the nature of the eroded SOM appear to be driven by the nature of the landscape (i.e. steep slopes), high OM in top soil and litter, and preferential lateral mobilization of light carbonaceous materials and small mineral particles by water erosion. The nature of the eroded SOM (modern, organic matter rich) also suggests that a large fraction of the SOM eroded from upland temperate ecosystems can be lost during transport or after deposition. In future studies, it will be important to determine the contribution of upland temperate forests to the erosion induced C sink (compared to croplands and grasslands).
REFERENCES


Stafford, A. K. 2011. Sediment production and delivery from hillslopes and forest roads in the southern Sierra Nevada California. Colorado State University, Fort Collins, CO.


TABLES AND FIGURES

| Table 1 Selected site characteristics of the two low-order western Sierra Nevada catchments.\(^a\) |
|---|---|---|---|---|---|
| **Catchment** | **Elevation (m a.s.l.)** | **Size (ha)** | **Temperature (°C)** | **Precipitation (mm)** | **Dominant Soil Family** |
| Providence | 1730-1990 | 132.3 | 11.3 ± 0.8 | 1549 ± 547 | Shaver |
| Bull | 2185-2490 | 138.4 | 7.8 ± 1.4 | 1583 ± 507 | Cagwin, Gerle-Cagwin |

\(^a\) Temperature is the average of 4 years (2004-2007) with standard deviation shown. Precipitation is the amount of rain and snow from WY 2005-2011 with the percentage that occurs as snowfall (from Bales et al. 2011 and Johnson et al. 2011). Elevation and catchment size are from the KREW Study Plan (Hunsaker et al. 2007).
Table 2: Characteristics of soils sampled in fall 2012 for Providence and summer 2013 for Bull catchments\(^b\).

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Landscape Position</th>
<th>Depth (m)</th>
<th>Gravimetric Water Content (kg water/kg soil)</th>
<th>Bulk Density (M/m(^3))</th>
<th>pH (in water)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Providence</td>
<td>Crest</td>
<td>0.0-0.1</td>
<td>0.12 (± 0.03)</td>
<td>0.77 (± 0.07)</td>
<td>5.87 (± 0.06)</td>
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<td></td>
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<td>0.1-0.2</td>
<td>0.11 (± 0.05)</td>
<td>ND</td>
<td>6.06 (± 0.08)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.2-0.4</td>
<td>0.10 (± 0.02)</td>
<td>ND</td>
<td>6.01 (± 0.09)</td>
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<td></td>
<td>0.4-0.6</td>
<td>0.08 (± 0.02)</td>
<td>ND</td>
<td>5.95 (± 0.11)</td>
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<td>0.6-0.8</td>
<td>0.09 (± 0.04)</td>
<td>ND</td>
<td>6.03 (± 0.17)</td>
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<td></td>
<td>Backslope</td>
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<td>0.16 (± 0.04)</td>
<td>0.75 (± 0.09)</td>
<td>6.25 (± 0.21)</td>
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<td>6.20 (± 0.19)</td>
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<td>6.02 (± 0.18)</td>
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<td>Deposition</td>
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<td>0.75 (± 0.08)</td>
<td>5.58 (± 0.24)</td>
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<td>Bull</td>
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<td>Backslope</td>
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<td>5.28 (± 0.09)</td>
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<tr>
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<td>Deposition</td>
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<td>1.01 (± 0.10)</td>
<td>5.20 (± 0.19)</td>
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<td>0.13 (± 0.06)</td>
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<td>5.26 (± 0.15)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.2-0.4</td>
<td>0.14 (± 0.05)</td>
<td>ND</td>
<td>5.19 (± 0.11)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.4-0.6</td>
<td>0.17 (± 0.05)</td>
<td>ND</td>
<td>5.11 (± 0.08)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.6-0.8</td>
<td>0.17 (± 0.01)</td>
<td>ND</td>
<td>5.04 (± 0.12)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.8-1.0</td>
<td>0.17 (± 0.03)</td>
<td>ND</td>
<td>4.93 (± 0.11)</td>
</tr>
</tbody>
</table>

\(^b\) All values are averaged across transects by depth shown with standard error in parentheses. The number of samples used in calculations is denoted by the superscript outside the parenthesis if sample size was not 6. Samples that have ND were not determined.
Figure 1: Location of the two study catchments (B203 and P303) within the Kings River Experimental Watersheds project. Sampled soil transects are shown with black lines and black dots denote sampling points. White points are stream bank and streambed points sampled within the catchments. The map shows degree of slope of all catchments designated by cool to warm colors that denote low to high slope, respectively. *Base maps from Stacy, 2012.*
Figure 2: Schematic of the profile illustration of landform positions sampled in the catchments and photographs of representative sampling points of each type of landform position. Profile adapted from Stacy, 2012. Picture credit: E.P. McCorkle and E.M. Stacy.
Figure 3: Boxplots of carbon (C) concentrations in the top 10 cm of mineral soil from all sample locations for both catchments. Crest (Bull n= 22; Providence n= 29), Bksl = backslope (Bull n= 21; Providence n= 25), Dep = deposition (Bull n= 30; Providence n= 28) and O Horizon (Bull n= 17; Providence n= 18) from each landform position. Stream samples: Bank = streambank (Bull n= 34; Providence n= 39), Bed = streambed (Bull n= 11; Providence n= 10). Sed = Sediment, where interannual variation of water years 2009-2013 is shown (Bull n= 5; Providence n= 5). Boxes are the first and third quartiles with the middle line denoting the second quartile or median. Lines on either side of the box show ±1.5 interquartile range (median) with points outside of the box showing outliers that fall outside of that range.
Figure 4: Boxplots of the carbon to nitrogen mass ratios of the top 10 cm of mineral soil within each sample type of both catchments. Crest (Bull n= 22; Providence n= 29), Bksl = backslope (Bull n= 21; Providence n= 25), Dep = deposition (Bull n= 30; Providence n= 28) and O Horizon (Bull n= 17; Providence n= 18) from each landform position. Stream samples: Bank = stream bank (Bull n= 34; Providence n= 39), Bed = stream bed (Bull n= 11; Providence n= 10). Sed = Sediment, where interannual variation of water years 2009-2013 is shown (Bull n= 5; Providence n= 5). Boxes are the first and third quartiles with the middle line denoting the second quartile or median. Lines on either side of the box show ±1.5 interquartile range (median) with points outside of the box showing outliers that fall outside of that range.
Figure 5: Stable isotope ($^{13}$C and $^{15}$N) biplots of potential organic matter source locations and collected sediments for Providence (left panel) and Bull (right panel) catchments. Soil samples from potential source locations were collected in fall 2012 from Providence and in summer 2013 from Bull; all stream bank and bed samples from both catchments were collected in the summer 2013. Symbols denote mean values for each potential source (0-10 cm depth) or collected sediment. For upland sources, n=6 for each landform position. For stream banks and beds, n = 10 for each in Providence and n = 11 for each in Bull). Surficial organic (O) horizon values are means across all transects and landform positions in each watershed (n = 18 for Providence, n = 17 for Bull). Horizontal and vertical bars denote ± 95% confidence intervals. Values for collected sediments are shown by individual water year (October 1 – September 30; n = 1 composite per year).
Figure 6: Depth profiles of $^{13}$C and $^{15}$N of both catchments at the upland landform positions. Providence samples are marked by squares and Bull samples are marked by circles. Error bars denote standard error of the depth class. All samples of each depth from each transect were averaged together for each point.
Figure 7: Fraction modern of subset analyzed for radiocarbon at chosen soil and stream bank sites in both catchments. The vertical black line denotes the divide between modern (> 1.0) and pre-modern (< 1.0) values. There is no standard error associated with these measurements as only one sample from each depth was analyzed.
Figure 8: Fraction modern of all sediments (WY 2009-2013) analyzed for radiocarbon in both catchments. The horizontal black line denotes the divide of modern (> 1.0) and pre-modern (< 1.0) values. There is no standard error associated with these measurements as only one sample from each depth was analyzed.