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Tracing the source of soil organic matter eroded from temperate forest catchments using carbon and nitrogen isotopes

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A B S T R A C T

Soil erosion continuously redistributes soil and associated soil organic matter (SOM) on the Earth's surface, with important implications for biogeochemical cycling of essential elements and terrestrial carbon sequestration. Despite the importance of soil erosion, surprisingly few studies have evaluated the sources of eroded carbon (C). We used natural abundance levels of the stable and radioactive isotopes of C (^{13}C and ^{14}C) and stable isotope of nitrogen (^{15}N) to elucidate the origins of SOM eroded from low-order catchments along the western slopes of the Sierra Nevada of California, USA. Our work was conducted in two relatively undisturbed catchments (low elevation = 1800 m, and high elevation = 2300 m) of the Kings River Experimental Watersheds (KREW) in the Sierra National Forest. Sediment captured in basins at the outlet of each gauged watershed were compared to possible source materials, which included: upland surficial organic horizons (i.e., forest floor) and mineral soils (0–0.6 m) from three landform positions (i.e., crest, backslope, and toeslope), stream bank soils (0–0.6 m), and stream-bed materials (0–0.05 m). We found that most of the organic matter (OM) in the captured sediments was composed of O-horizon material that had high C concentrations. Radiocarbon analyses also showed that the captured OM is composed of modern (post-1950) C, with fraction modern values at or above 1.0. Our results suggest that surface (sheet) erosion, as opposed to channeling through established streams and episodic mass wasting events, is likely the largest source of sediment exported out of these minimally disturbed, headwater catchments. The erosional export of sediment with a high concentration of C, especially in the form of relatively undecomposed litter from the O horizon, suggests that a large fraction of the exported C is likely to be decomposed during or after erosion; hence, it is unlikely that soil erosion acts as a significant net sink for atmospheric CO_2 in these low-order, temperate forest catchments.

1. Introduction

Quantification of the fluxes of carbon (C) within and from terrestrial ecosystems is critical for generating an accurate global C budget and predicting future effects of soil C storage on the Earth's climate system (Falkowski, 2000). Soil is the third largest global C reservoir (>2500 Pg) and contains nearly double the C found in terrestrial living biomass (560 Pg) and the atmosphere (750 Pg) combined (Condron et al., 2010; Dungait et al., 2012; Smith, 2012; Tarnocai et al., 2009). Much of the C held within soil exists as free organic particles, or organic matter (OM) bonded to mineral particles or encapsulated in aggregates (De Baets et al., 2012).

Globally, soil erosion is an important and widespread flux in the C cycle that redistributes on the order of 75 Gt of soil and 1–5 Gt soil organic carbon annually (Battin et al., 2009; Berhe et al., 2007; Regnier et al., 2013; Stallard, 1998). In uplands, erosional transport of topsoil and associated OM can lead to loss of soil organic matter (SOM) through direct removal of soil mass. About 70–90% of the eroded topsoil material is redistributed downhill or downstream, while the rest is exported out of the source watersheds (Doetterl et al., 2016; Gregorich et al., 1998; Lal, 2003; Rumpel et al., 2006). For the material deposited locally, erosion leads to stabilization of at least some of the eroded SOM in depositional landforms through burial, and new and reconfigured associations of the eroded OM with soil minerals (Berhe et al., 2012a; Sharpley, 1985). Soil erosion and subsequent deposition lead to a net global

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terrestrial sink of 0.12–1.5 Gt C yr^{-1}, particularly when their contribution to C storage is assessed at a watershed scale and dynamic replacement of eroded OM by production of new photosynthate is taken into account (Berhe et al., 2008; Berhe et al., 2007; Doetterl et al., 2016; Harden et al., 2008; Harden et al., 1999; Stallard, 1998; Van Oost et al., 2007). Considerable uncertainty, however, still remains as to the fate of the OM that is exported out of eroding catchments.

Lateral distribution of topsoil by erosion can significantly alter the fate of soil C. The amount and composition of SOM found in soil-mantled hillslopes reflect the balance between inputs of OM, mainly from net primary productivity and deposition from erosion upslope, and outputs and transformations of OM by processes including decomposition, leaching, and erosion downslope (Berhe et al., 2007). The composition of SOM transported by erosion can be similar to the SOM in hillslope soil profiles (Bellanger et al., 2004). But, several variables control the rate and nature of soil erosion, including climate (specifically precipitation amount, distribution, and intensity), slope (angle and length), land use, and lithology. All of these variables can impose important controls on the amount and composition of eroded C and its post-erosional transport (Berhe et al., 2014; Kim et al., 2013; Nadeu et al., 2012).

Source pools of SOM from different landform positions can affect the persistence of eroded C post-erosion. For example, soil material eroded from the surficial soil horizons (e.g., O or A horizons) is likely to have a large proportion of relatively undecomposed OM with a high carbon to nitrogen ratio (C:N). Organic horizon material will also be largely freely associated rather than associated with soil minerals, making this SOM relatively easy to decompose and mineralize during transport or after deposition downstream. On the other hand, soil material eroded from deeper soil layers, such as B horizons by gully erosion, is likely to have a low proportion of relatively undecomposed OM, with much of the eroded C associated with soil minerals through aggregation or sorptive interactions. As a result, a higher proportion of the material eroded from deeper soil layers should be stabilized after it has been transported by erosion and deposited downslope, compared to material eroded from surficial soil horizons. Hence, tracing the movement of SOM from eroding landform positions, as well as identification of its ultimate depositional setting within a catchment, are essential for quantifying the fate of laterally transported C on the landscape (Berhe and Kleber, 2013).

Stable isotope analysis is a potentially powerful tool for sourcing eroded SOM in natural and agricultural landscapes. Traditionally, radioactive isotopes are used to estimate the rate of soil erosion (Heimsath et al., 1997; Kim, 2008; Kirchner et al., 2006; Kirchner et al., 2001; Portenga and Bierman, 2011; Riebe et al., 2001) and to infer associated rates of soil C erosion (Klaminder et al., 2009; Yoo et al., 2005). Some radionuclides (i.e., ^{137}Cs, ^{7}Be) are useful tracers for estimating the rate of soil erosion because they bind tightly to soil minerals and are only redistributed when the soil minerals are mobilized vertically or laterally (Alewell et al., 2009; Kim et al., 2013). Soil movement, relative to an undisturbed reference site, can then be estimated by enrichment (i.e., deposition) or depletion (i.e., erosion) of these isotopes in a soil profile. However, short half-lives (e.g., 30 years for ^{137}Cs and 53 days for ^{7}Be) and a lack of reference sites (i.e., sites with no erosion or deposition) make using radionuclides unreliable over time or in some environments. Furthermore, the relatively high cost of radionuclide analyses in soils and sediments limits their broad application across ecosystems (Schaub and Alewell, 2009).

Using the natural abundance of stable isotopes in the environment offers a more consistent and less expensive alternative to radionuclides for tracing pathways of erosion (Meusburger et al., 2013). Stable isotope compositions of C (δ^{13}C) and nitrogen (δ^{15}N) vary significantly among ecosystem pools because of isotopic fractionation during the cycling of C and N (Hogberg, 1997; O'Leary, 1988). Differences in the δ^{13}C and δ^{15}N values of OM pools in upland soils and floodplains have been used to follow the transport of soil materials mobilized by precipitation events in catchments and deposited in accumulated sediments (Bellanger et al., 2004; Fox and Papanicolaou, 2007). In addition, stable isotopes have been used to trace the sources of eroding material occurring on catchment hillslopes (Hilton et al., 2012; Meusburger et al., 2013) and to follow SOM bound to stream gravels (Collins et al., 2013). Radiocarbon (^{14}C) is routinely used to determine turnover and stability of SOM within an ecosystem (Torn et al., 2009; Trumbore, 2009; Trumbore et al., 1996; Trumbore et al., 1989; Wang et al., 1996) because the amount of ^{14}C remaining in SOM reflects the time since the original material last exchanged carbon dioxide (CO_2) with the atmosphere (Trumbore, 2009). The lower the amount of ^{14}C remaining in OM, the longer the material has been present within the soil. In stable landscape positions, SOM generally becomes older with depth, but if erosion leads to deep soil removal from hillslopes and subsequent burial downslope, SOM age can decrease with depth (Berhe et al., 2012a; Torn et al., 2009).

We studied the sources of OM transport by soil erosion, and how these sources varied as a function of elevation (proxy for type of precipitation), in two catchments within the rain-snow transition zone of the Sierra Nevada, California, USA. The higher elevation catchment receives the vast majority of its annual precipitation as snow, while the lower elevation catchment receives approximately half of its precipitation as rain and half as snow. These differences in precipitation type may alter the sources of erosion because snow-dominated catchments generate higher peak stream flows in the spring during snowmelt, while lower elevation catchments have greater potential for rain-on-snow events, possibly leading to the occurrence of overland flow during the winter months (Bales et al., 2006; Hunsaker and Neary, 2012). Furthermore, because these catchments are similar in their other state factors (i.e., parent material, vegetation, topography, and landsurface age), they are an excellent surrogate experimental system for evaluating possible changes in soil erosion processes as the climate warms (Bales et al., 2011; Jenny, 1941).

Rates of sediment yield and the C and N contents of these sediments were characterized previously in these catchments over multiple years by Stacy et al., (2015). They found that sediment yield and C and N concentrations showed high interannual variation. Annual sediment export was positively correlated with stream discharge, while both C and N concentrations were negatively correlated with stream discharge. They speculated that the primary source of the organic materials captured in sediment traps at the outlet of the catchments was OM-rich litter material and topsoil. Stream discharge, more than sediment source, was the primary factor controlling the magnitude of C and N export from these upland forest catchments. In addition, no significant difference in the amount or C and N content of sediment derived from catchments of different elevations was found, suggesting that differences in precipitation type may not be an important driver of sediment yield. The present study builds upon this previous investigation by evaluating if the sources of the SOM found in sediments across contrasting years and catchment elevations are indeed similar, and consist primarily of surficial SOM.

We used stable and radioactive isotopes of SOM to trace the origin of eroded materials within two, low-order montane watersheds of contrasting elevation in the Sierra Nevada. Our objectives were to: 1) evaluate the efficacy of stable isotopes of C and N as erosion tracers, 2) determine the origin of eroded SOM collected in sediment traps at the outlet of each catchment, and 3) infer fate of eroded SOM using ^{14}C at the catchment scale. By applying this approach to two similar, low-order catchments within the rain–snow transition zone, we were able to determine if precipitation type affects the sources of sediment leaving these catchments, and whether it altered the fate of the eroded OM.

2. Methods

2.1. Site description

Field sampling was conducted in the Kings River Experimental Watersheds (KREW) located within the Sierra National Forest of

California (Fig. 1). The KREW project began in 2000 and is operated by the United States Forest Service (USFS). The entire KREW project consists of eight low-order (i.e., headwater) catchments: four at lower elevations (Providence Watersheds, 1485–2115 m.a.s.l.) and four at higher elevations (Bull Watersheds, 2050–2490 m.a.s.l.; Hunsaker et al., 2007). The lower elevation sites are also co-located with the National Science Foundation-supported Southern Sierra Critical Zone Observatory (SSCZO; Bales et al., 2011). Our study used two of these catchments: one at higher elevation (Bull 203) and one at lower elevation (Providence 303). These two catchments were selected because they are the most similar across the altitudinal gradient in size and topography (Table 1).

The study sites are part of the western Sierra climosequence (Dahlgren et al., 1997), where all state factors other than climate (i.e., parent material, vegetation, topography, and landsurface age) are comparable (Hunsaker and Neary, 2012; Jenny, 1941). The catchments have a Mediterranean-type climate characterized by warm, dry summers and cool, wet winters. Differences in elevation between the two catchments result in contrasting temperature and precipitation regimes. Mean (± standard deviation) annual air temperatures range from 11.3 ± 0.8 °C in Providence to 7.8 ± 1.4 °C in Bull (years 2004–2007; Johnson et al., 2011). Both catchments receive similar amounts of precipitation during the year (mean annual precipitation ± standard deviation = 1549 ± 547 mm for Providence and 1583 ± 507 mm for Bull; Water Years 2005–2011; WY, October 1–September 30). However, the higher elevation catchment receives the vast majority (75–90%) of its precipitation as snow, whereas the lower elevation catchment receives about half of its precipitation as rain and half as snow (35–60% snow; Hunsaker and Neary, 2012; Climate and Hydrology Database Projects, www.fsl.orst.edu/climhy).

The poorly developed soils within these catchments are derived from granite or granodiorite, and are dominated by the Shaver soil series in the Providence catchment (coarse-loamy, mixed mesic Pachic Xerumbrepts), and the Cagwin soil series in the Bull catchment (mixed, frigid Dystric Xeropsamments). Smaller amounts of the Cagwin series is found in Providence, and smaller amounts of the Gerle soil series (coarse-loamy, mixed, frigid Typic Xerumbrepts) is found in Bull. Vegetation type is predominantly Sierran mixed conifer at Providence (94% cover). Sierran mixed conifer consists of 1/3 each pine (*Pinus lambertiana*, *Pinus ponderosa*), fir (*Abies concolor*), and cedar (*Libocedrus decurrens*) species. The Bull catchment is co-dominated by Sierran mixed conifer with the addition of *Pinus jeffreyi* (47% cover) and red fir (44% cover; *Abies magnifica*; Hunsaker et al., 2007; Johnson et al., 2011). Canopy cover varies within both catchments but generally increases as elevation decreases or in closer proximity to streams.

2.2. Sampling design

We established six, spatially independent hillslope transects in each of the two catchments. Replicate mineral soil samples were collected at each of three landform positions along the transects: crest, backslope, and toeslope/deposition (Fig. 2). The landform positions encompassed the path of eroded material moving downhill.

Four of the Providence transects (P1–P4) and one of the Bull transects (B6) were also studied by Stacy et al. (2015). We used the same

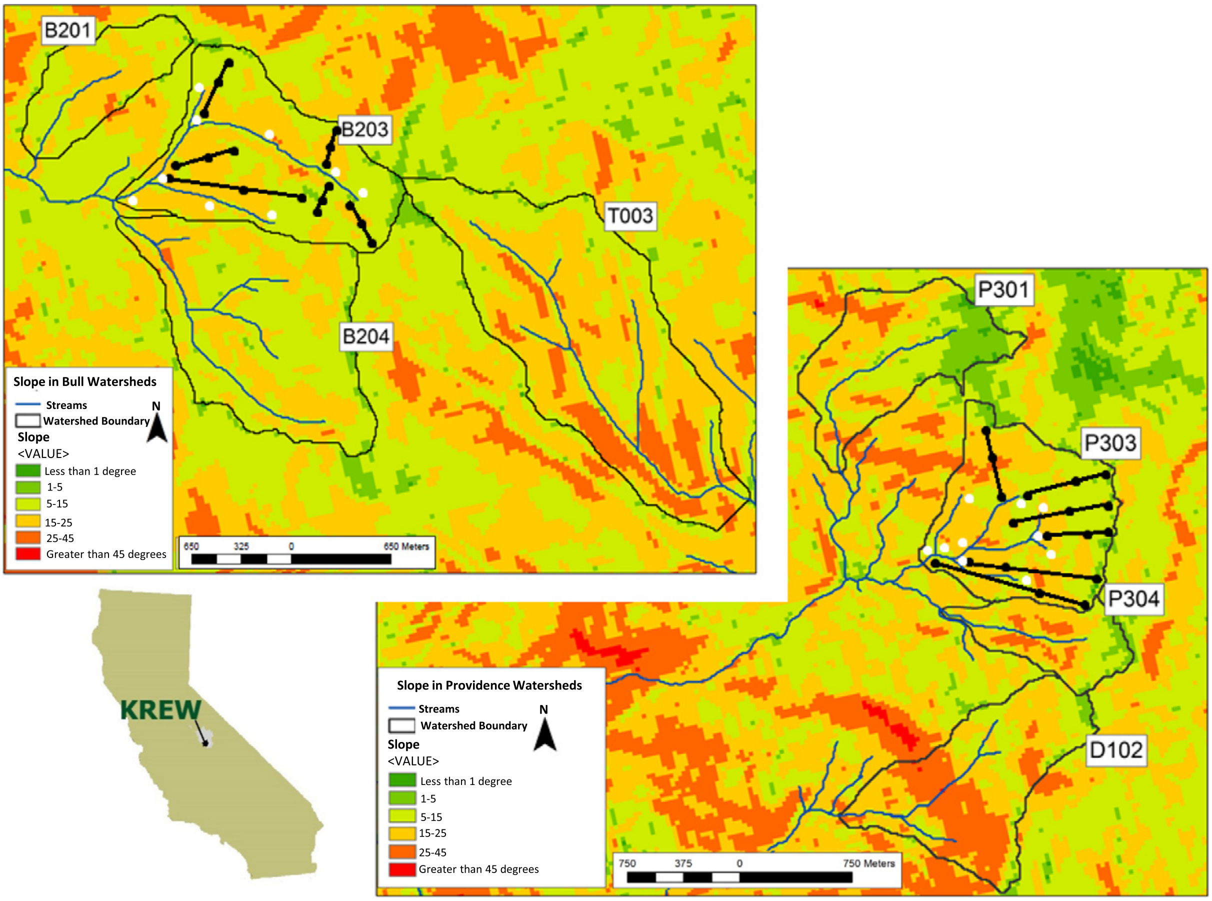

Fig. 1. Location of the two study catchments (B203 and P303) within the Kings River Experimental Watersheds project. Sampled soil transects are shown with black lines and black dots denote sampling points. White points are stream bank and stream bed points sampled within the catchments. The map shows degree of slope designated by cool to warm colors that denote low to high slope, respectively.
Base maps from Stacy et al. (2015).

Table 1
Selected characteristics of the two, low-order western Sierra Nevada catchments used in this study.[a]

Catchment	Elevation (m.a.s.l.)	Size (ha)	Temperature (°C)	Precipitation (mm)	Dominant soil series
P303 in Providence	1730–1990	132.3	11.3 ± 0.8	1549 ± 547 (35–60% as snow)	Shaver
B203 in Bull	2185–2490	138.4	7.8 ± 1.4	1583 ± 507 (75–90% as snow)	Cagwin

[a] Temperature is the average of four years (2004–2007) with interannual standard deviation shown. Precipitation is the amount of rain and snow from WY 2005–2011 with the percentage that occurs as snowfall (from Hunsaker and Neary, 2012 and Johnson et al., 2011). Elevation and catchment size are from the KREW Study Plan (Hunsaker et al., 2007).

site selection criteria as Stacy et al. (2015) to establish additional hillslope transects for a total of six per catchment to capture spatial heterogeneity. 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. Each transect is oriented on a western facing slope with points selected in open areas free from any recreational or grazing disturbance observed at time of establishment. Western facing aspects were chosen as they dominate both catchments. By placing transects in relatively open areas, we chose sites that would be possible sources of eroded sediment, i.e., material that is likely to be mobilized by water-based soil erosion.

Soil samples were collected at each landform position using a 1-m^2 plot with three sample points chosen at random. The surficial O horizon was collected at each point within a 10 × 20 cm area and composited within each plot 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 up to 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 boulders or large roots impeded the auger at depths less than 40 cm, other points were chosen at random within 20 cm of the original point. After six attempts, mineral soil successfully cored to the desired depth classes was retained and the deepest depth reached was recorded.

In addition to upland soils, stream bank soils and stream bed materials were also collected to evaluate these locations as possible sources of eroded material captured as sediment leaving the catchment. Sampling points were chosen along the 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 (see below). Stream bank samples were collected on an approximately 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 samples were taken with stream bank samples because O horizons were not found at these landform positions. The stream bank mineral soils were hand augered only to 60 cm depth (0–10, 10–20, 20–40, and 40–60 cm) because soils were water-saturated below this depth. Stream bed material was collected perpendicular to the bank sampling point in the middle of the stream path. Bed material was collected within a 10 × 10 cm area to a depth of 5 cm using a flat trowel. The bed material depth was chosen to include only those bed materials that would freely move with the flowing stream.

In 2003 and 2004, sediment basins were constructed in both the Providence and Bull catchments at a natural depositional area at the

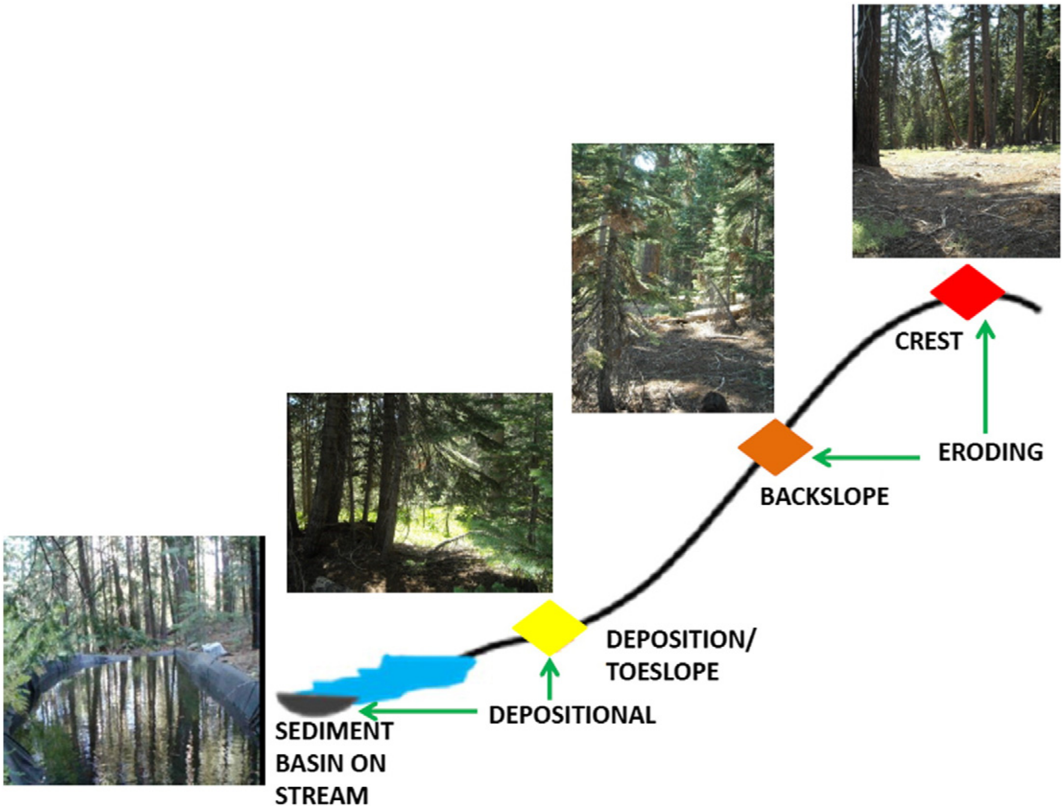

Fig. 2. Schematic showing a typical hillslope-based transect within one of the low-order catchments. The types of landforms sampled and their relative positions along the hillslope are shown, along with photographs of representative sampling points at each type of landform.
Adapted from Stacy et al. (2015). Photo credit: E.P. McCorkle and E.M. Stacy.

outlet of each catchment (Hunsaker et al., 2007). The area was deepened and enclosed with logs held in place by rebar at the outlet, and the streambed and both banks were 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 column. The sediment basins are emptied annually close to the end of the WY (August or September) by hand after temporarily diverting the incoming stream. Material collected from these basins is referred to as "sediment" throughout this paper. The wet sediment was weighed on a spring scale to the nearest kg and large woody debris (longer than 0.5 m) was collected on a tarpaulin. Large woody debris was weighed wet and a subsample air dried to get the total dry weight. A subsample was taken from each bucket of wet sediment removed and composited to create one representative sample of the entire basin. This sample was then mixed and subsampled again for analysis in this study. The United States Pacific Southwest Forest Service Research Station in Fresno, CA, processed several subsamples from each basin to determine mineral and organic fractions of the sediment, and stored dried archives from each year (Hunsaker and Neary, 2012). For this study, the samples for WY 2012 and 2013 (October 1, 2011–September 30, 2012 and October 1, 2012–September 30, 2013, respectively) were taken from the field site and stored at 4 °C until analyzed (~two weeks), while samples from WY 2009–2011 were taken from the air-dried Forest Service archives.

2.3. Chemical analyses

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 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. Large woody debris was excluded from this study because field observations found that most of the large woody debris falls directly into the basin from the surrounding trees.

Soil and sediment samples were measured for physical and chemical soil characteristics, including: gravimetric water content, bulk density, and pH. For gravimetric water content, samples were homogenized and approximately 10 g dried at 105 °C for 48 h. 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 × 10 cm core. The soil was then sieved to <2 mm, dried at 105 °C, and weighed; reported bulk density values are the dry mass of the <2 mm material divided by the core volume. 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 (mass to volume) soil to deionized water suspension. The pH of upland soils was measured on field-moist soil, while 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 et al., 2015).

Sieved subsamples of all soils and stream bed material were left to air dry before being ground to a fine powder using a mortar and pestle. Air-dried, O-horizon samples were ground using a Wiley mill (Thomas Scientific, Swedesboro, NJ) to <425 μm. Ground samples were analyzed for $\delta^{13}C$, $\delta^{15}N$, and elemental C and N at the Geophysical Laboratory of the Carnegie Institute in Washington, D.C. on a Carlo Erba NC 2500 elemental analyzer (Carlo Erba, Milan, Italy) connected to a Delta V Plus Isotope Ratio Mass Spectrometer (ThermoFisher Scientific, Waltham, MA) via a Conflo III interface (Thermo Finnegan, San Jose, CA). Samples were measured relative to an internal acetanilide standard with an analytical precision of ±0.01‰ for $\delta^{13}C$ and $\delta^{15}N$. Replicate analyses of a representative soil sample were analyzed in each run; reproducibility of this homogenized soil was ±0.20‰ for $\delta^{13}C$ and ±0.25‰ for $\delta^{15}N$. The relatively low pH of soil and sediments in both of these catchments (pH in water generally <6.0) suggested that inorganic C was not present; spot checks using 1 M HCl (and observing for any effervescence) conducted previously on similar materials from these catchments confirmed the lack of the presence of carbonates. Hence, no pretreatment was conducted on these samples to remove carbonates before elemental and isotopic analyses.

Stable isotope analyses were performed on samples of sediments from both catchments collected in all years, but radiocarbon analysis was only performed on one representative transect and one stream bank point in each catchment because of cost constraints. The transects and stream banks analyzed were chosen using $\delta^{13}C$ values considered representative of the catchments (i.e., similar to median values of all transects and bank points within that catchment). These samples were used to augment the stable isotope results because the age of SOM contained within the soils and sediments should relate to their origin (e.g., shallow vs. deep soil depths). For radiocarbon analyses, 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). Radiocarbon data in this work is presented as Fraction Modern (FM; Reimer et al., 2004), which was corrected for mass dependent fractionation with measured $\delta^{13}C$ for a given sample. Analytical precision averaged 0.0031 FM. The FM value of a sample is then used to determine fraction of C that is modern (post-1950) versus old (pre-1950) depending on whether the value is greater than or less than 1 (FM = 1 indicating C from 1950), respectively.

2.4. Data analyses

We present our data in this manuscript as mean ± standard error (or occasionally, ±95% Confidence Interval). We refrained from using inferential statistics to test for significant differences between and among means because, even though we used replicated transects within a catchment, only one catchment at each elevation range was assessed. We graphed $\delta^{13}C$ versus $\delta^{15}N$ to identify potential sources of eroded organic matter (i.e., upland soils, stream bank soils, and stream bed materials) for the sediments captured within basins at the outlet of the catchments. Source isotopic values most similar to sediment isotopic values were considered the dominant potential source of the eroded OM. Additionally, sediment isotope C and N values that plotted between or among two or more source isotopic compositions suggested mixed sources of the eroded OM. We used 95% confidence intervals about mean values to distinguish unique isotope signatures among potential sources and collected sediments. This "fingerprinting" technique assumes that relatively little isotopic fractionation occurred in these materials during transport and storage within the sediment basins before collection. The assumption of little isotopic fractionation occurring during transport is reasonable for our study site where the sediment (and associated OM) was only mobilized over short transport distances, owing to the steep topography of the landscape (we discuss this assumption further in the Discussion section). A two-end member mixing model was then used to determine the contribution of the different sources to sediment OM for each WY (Phillips and Koch, 2002). We estimated source contributions using C and N isotopic values separately because in the Providence watershed sediment $\delta^{13}C$ values (and in one case $\delta^{15}N$) were typically outside the range of the measured source value means (see Discussion section).

3. Results

3.1. Soil characteristics

Tables 1 and 2 give descriptions of environmental conditions at the study sites, and physical and chemical properties of our samples.

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 unit less than crest or backslope in both catchments; Table 2).

At the comparable times of sampling (fall 2012 for Providence and summer 2013 for Bull), Providence soils had higher gravimetric water contents, averaged at all landform positions and depths (0.15 kg/kg ± 0.01), than Bull (0.10 kg/kg ± 0.01). Deposition sites were the wettest landform positions in both catchments (Table 2). Soil gravimetric water content declined with depth at all landform positions and in both catchments. Bulk density values (0–10 cm depth) ranged from 0.75 to 0.94 Mg/m^3, with higher values in Bull than in Providence at all landform positions.

The relative contribution of organic and mineral materials to sediments collected within the sediment basins appeared correlated with annual precipitation (Table 3). In both catchments, the proportion of the sediment collected in the sediment basin as OM (based on dry mass) generally decreased as annual precipitation increased. However, in both catchments, these relationships were not statistically significant, possibly the result of low statistically power (linear or log-linear regressions, $P > 0.10$, $n = 5$ per catchment).

3.2. Elemental C and N

Elemental concentrations are presented as percent C and N (rather than nutrient mass per soil volume) because our samples were obtained by a soil auger. We determined soil bulk density only for the 0–10 cm depth. Soil bulk density in these catchments and others of KREW can be found in Stacy et al. (2015) and Johnson et al. (2011). Concentrations of C in the O horizon and the underlying 0–10 cm mineral soil layer (Fig. 3) increased from the crest to depositional landforms in both catchments. Mean C 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, and 0.03% at deposition). In contrast, 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, 16.50% ± 3.62 for stream bed and stream bank, respectively; Fig. 3). Depth profiles also show a decrease in both percent C and N with depth in both catchments (Fig. 4).

Table 3
KREW sediment basin mineral and organic proportions (percentages based on dry mass).

		P303			B203		
Water Year	Average precipitation[a] (mm)	Mineral	Organic coarse[b] and fine	Organic LWD[c]	Mineral	Organic coarse and fine	Organic LWD
2009	1260	37.6	61.5	0.9	66.1	32.5	1.4
2010	1630	44.1	54.6	1.3	87.4	11	1.6
2011	2280	85.9	11.5	2.6	75.1	21.4	3.5
2012	1020	74.1	25.4	0.6	69.3	30.7	0
2013	880	38.3	61.4	0.3	45.9	51	3.1

[a] Data from upper Providence meteorology station.

[b] Coarse organic material is pieces of bark, wood, and cones that can be easily picked out of the sediment.

[c] LWD is large woody debris (wood longer than 50 cm in length).

Providence sediments overall had higher C concentrations than sediments from Bull. Differences in sediment C were greater among years than between catchments. The highest C concentrations were measured in samples collected in WY 2009 (17.85%) and 2013 (19.91%) in Providence and Bull, respectively. The lowest C concentrations were measured in samples collected in WY 2011 (2.10%) and 2010 (2.06%) in Providence and Bull, respectively. Note that there is no standard error associated with the sediment values because there was only one sediment basin per catchment per year.

In Providence, the atomic carbon to nitrogen ratio (C:N$_{at}$) showed a similar pattern to the C concentrations of the surface soils at all landform positions. In Bull, C:N$_{at}$ ratios for the stream bank and stream

Table 2
Characteristics of soils sampled in fall 2012 for Providence and summer 2013 for Bull catchments.[b]

Catchment	Landscape position	Depth (m)	Gravimetric water content (kg water/kg soil)	Bulk density (Mg/m^3)	pH (in water)
Providence	Crest	0.0–0.1	0.12 (±0.03)	0.77 (±0.07)	5.87 (±0.06)
		0.1–0.2	0.11 (±0.05)	ND	6.06 (±0.08)
		0.2–0.4	0.10 (±0.02)	ND	6.01 (±0.09)
		0.4–0.6	0.08 (±0.02)	ND	5.95 (±0.11)
		0.6–0.8	0.09 (±0.04)^4	ND	6.03 (±0.17)^4
	Backslope	0.0–0.1	0.16 (±0.04)	0.75 (±0.09)	6.25 (±0.21)
		0.1–0.2	0.14 (±0.03)	ND	6.20 (±0.19)
		0.2–0.4	0.11 (±0.03)^5	ND	6.32 (±0.14)^5
		0.4–0.6	0.08 (±0.01)^4	ND	6.02 (±0.18)^4
		0.6–0.8	0.06 (±0.00)^2	ND	6.09 (±0.18)^2
	Deposition	0.0–0.1	0.28 (±0.05)	0.75 (±0.08)	5.58 (±0.24)
		0.1–0.2	0.22 (±0.04)	ND	5.74 (±0.19)
		0.2–0.4	0.23 (±0.04)	ND	5.72 (±0.16)
		0.4–0.6	0.20 (±0.04)	ND	5.82 (±0.15)
		0.6–0.8	0.27 (±0.08)^3	ND	5.59 (±0.15)^3
Bull	Crest	0.0–0.1	0.04 (±0.01)	0.94 (±0.03)	5.48 (±0.11)
		0.1–0.2	0.06 (±0.01)	ND	5.51 (±0.12)
		0.2–0.4	0.06 (±0.00)^5	ND	5.51 (±0.09)^5
		0.4–0.6	0.06 (±0.01)^3	ND	5.61 (±0.05)^3
	Backslope	0.0–0.1	0.07 (±0.02)	0.82 (±0.06)	5.28 (±0.09)
		0.1–0.2	0.07 (±0.01)	ND	5.31 (±0.08)
		0.2–0.4	0.08 (±0.01)^4	ND	5.31 (±0.07)^4
		0.4–0.6	0.06 (±0.02)^4	ND	5.22 (±0.06)^4
	Deposition	0.0–0.1	0.19 (±0.09)	0.85 (±0.07)^5	5.20 (±0.19)
		0.1–0.2	0.13 (±0.06)	ND	5.26 (±0.15)
		0.2–0.4	0.14 (±0.05)	ND	5.19 (±0.11)
		0.4–0.6	0.17 (±0.05)^4	ND	5.11 (±0.08)^4
		0.6–0.8	0.17 (±0.01)^3	ND	5.04 (±0.12)^3
		0.8–1.0	0.17 (±0.03)^3	ND	4.93 (±0.11)^3

[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. ND denotes not determined.

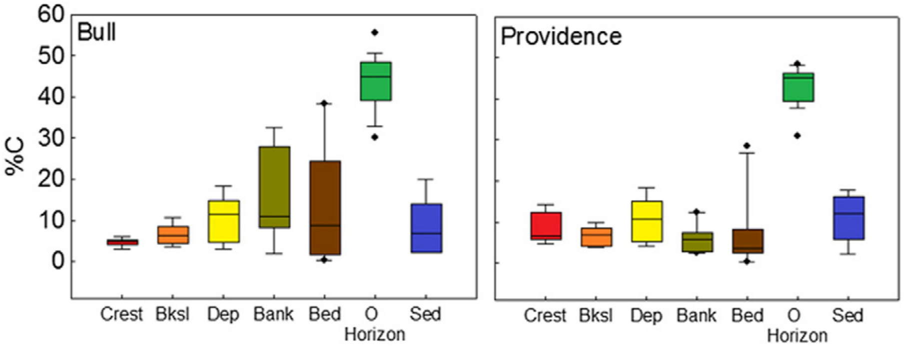

Fig. 3. Boxplots of carbon (C) concentrations in the top 10 cm of mineral soil and O horizons from all sample locations for both catchments. Crest (Bull and Providence n = 6), Bksl = backslope (Bull and Providence n = 6), Dep = deposition (Bull and Providence n = 6) and O-horizon (Bull n = 17; Providence n = 18) from each landform position. Stream samples: Bank = stream bank (Bull n = 11; Providence n = 10), Bed = stream bed (Bull n = 11; Providence n = 10). Sed = Sediment, where interannual variation among WY 2009–2013 is shown (Bull and Providence n = 5). Boxes are the first and third quartiles with the middle line denoting the second quartile, or median. Whiskers on either side of the box are ±1.5 times the interquartile range (IQR = quartile 3–quartile 1) higher or lower than the box, with points outside of the whiskers considered as outliers.

bed were lower than in the 0–10 cm mineral soils (Fig. 5). Soils in the Bull catchment had lower N concentrations than Providence leading to the observed higher C:N_{at} ratios in soil samples from Bull (Providence: 24.3 ± 1.2, 24.3 ± 1.3, 29.4 ± 3.2 and Bull: 24.8 ± 2.4, 27.7 ± 3.0, 30.2 ± 5.0 for crest, backslope, and deposition landform positions, respectively; Fig. 5). The C:N_{at} ratios in the 0–10 cm mineral soils were more similar to those of sediments than C:N_{at} values of the O horizon (i.e., forest floor) in both catchments. Sediments had lower interannual variation of C:N_{at} ratios in Bull than in Providence among years (Fig. 5).

3.3. Stable carbon ($\delta^{13}C$) and nitrogen ($\delta^{15}N$) isotopes in SOM

Pools of C and N from potential sources of SOM and collected sediments had isotopic compositions that were distinct from each other (Fig. 6). Isotopic compositions of upland soils (both O-horizon and 0–10 cm mineral soil) and sediment from years 2010, 2012, and 2013 had higher values in both δ^{13}C and δ^{15}N by 1 to 2‰ at the higher elevation catchment (Bull) compared to the lower elevation catchment (Providence). In the both catchments, the means of δ^{13}C and δ^{15}N in the stream bed and bank samples were statistically similar based on

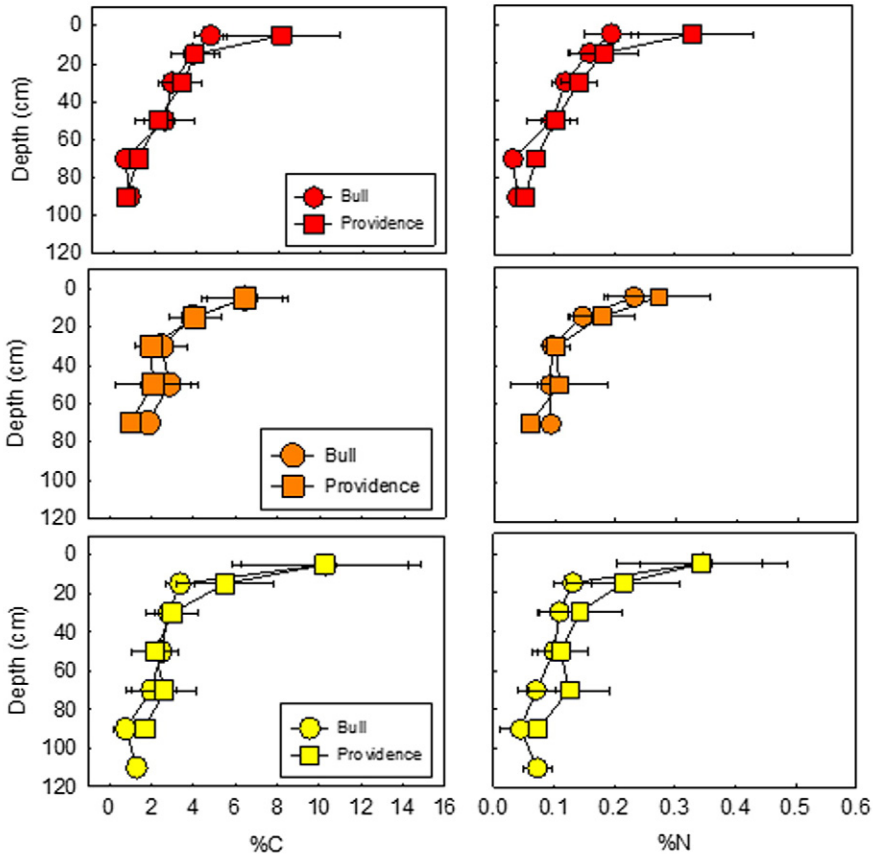

Fig. 4. Depth profiles of percent C and N of both catchments at the upland landform positions. Mean values for Providence samples are marked by squares and mean values for Bull samples are marked by circles. Red = crest, orange = backslope, and yellow = deposition/toeslope. All samples from the 6 transects are included in the averages. Error bars denote standard error of the mean for that depth class across transects within a catchment.

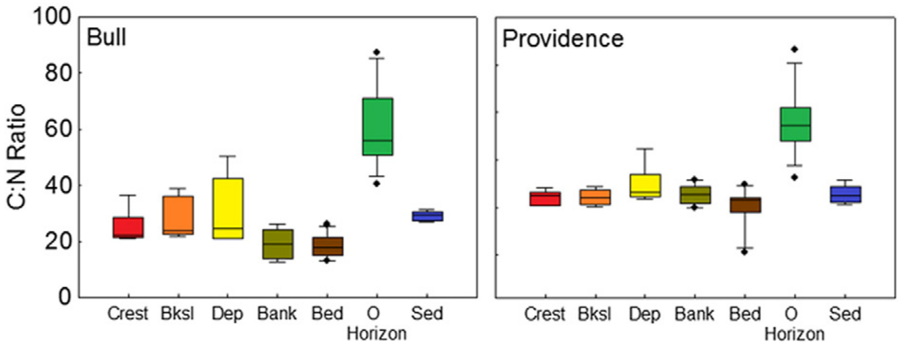

Fig. 5. Boxplots of the carbon to nitrogen mass ratios of the top 10 cm of mineral soil and O horizons within each sample type of both catchments. Crest (Bull and Providence n = 6), Bksl = backslope (Bull and Providence n = 6), Dep = deposition (Bull and Providence n = 6) and O-horizon (Bull n = 17; Providence n = 18) from each landform position. Stream samples: Bank = stream bank (Bull n = 11; Providence n = 10), Bed = stream bed (Bull n = 11; Providence n = 10). Sed = Sediment, where interannual variation of WY 2009–2013 is shown (Bull and Providence n = 5). Boxes are the first and third quartiles with the middle line denoting the second quartile, or median. Whiskers on either side of the box are ±1.5 times the interquartile range (IQR = quartile 3–quartile 1) higher or lower than the box, with points outside of the whiskers considered as outliers.

overlapping 95% confidence intervals. The upland soils were isotopically more similar to the stream bed in Providence than in Bull. In Bull, stream bed and bank materials had intermediate values between the upland soils and the sediments and O horizon materials.

Within the Providence catchment, we found a considerable range in δ^{13}C (5‰) and δ^{15}N (6‰) values. Here, mineral soils from all landscape positions have similar δ^{13}C and δ^{15}N values (Fig. 6). The O horizon was averaged within a catchment because there was very little variation in stable isotope values among landscape positions or transects. Stable isotopic values of stream bank and bed materials were similar to each other, and tended to have lower δ^{13}C lower values (but similar δ^{15}N values) than mineral soils. Sediments displayed considerable interannual variation in both δ^{13}C and δ^{15}N. Sediment isotopic values tended to be intermediate between O horizon and stream bank and bed values (Fig. 6).

Similar variations in isotopic compositions were found in the higher elevation Bull catchment. Surface mineral soils (0–10 cm) from Bull were generally more enriched in ^{13}C and ^{15}N than comparable soil from Providence (Fig. 6). As in the Providence catchment, O-horizon stable isotope values were averaged within a catchment because there was little variation among landscape positions or transects. Also similar to the Providence catchment, stream bank and bed had very similar δ^{13}C and δ^{15}N values; however, for the Bull Catchment, both of these isotope values were lower than in mineral soils. The stable isotopic composition of sediments had less interannual variation than in Providence, but these values were still intermediate between the O horizon and stream bank and bed materials (Fig. 6).

The δ^{13}C values showed enrichment with depth at all landform positions and in both catchments (Fig. 7). Values of δ^{15}N were more varied than δ^{13}C with depth at all landform positions and in both catchments (Fig. 7). Some depth profiles even showed ^{15}N depletion with depth (e.g., crest landform position in both catchments and the backslope landform position in the Providence catchment).

3.4. Mixing model

The mixing model results support the findings depicted in our isotopic C vs. N graphs (Table 4). Based on the δ^{15}N values, sediments transported from the Bull catchment show that 58 to 100% of the OM in the sediments is sourced from O-horizon material, with the rest attributed to stream bed and/or stream bank. The lowest contribution of the O-horizon material to transported OM (58%) was in 2013, and the highest in WY 2010 (100%). The contribution of O-horizon material to sediment OM was lower in the Providence than in the Bull catchments. The mixing model calculations attribute more than 100% of the sediment OM to O-horizon in WY 2011 (the wettest year in the study period) in the Providence catchment. For the other years, O-horizon material contributed 42–69% of the OM to the sediments in the Providence catchment.

3.5. Radiocarbon

Radiocarbon data from the representative soil samples analyzed showed mostly modern material in Providence and a split between pre-modern and modern SOM in Bull (Fig. 8).

Providence had more modern organic C in the soils at depth than Bull, and C in shallow mineral soil layers (<20 cm deep) had modern radiocarbon signature at all landform positions in both catchments.

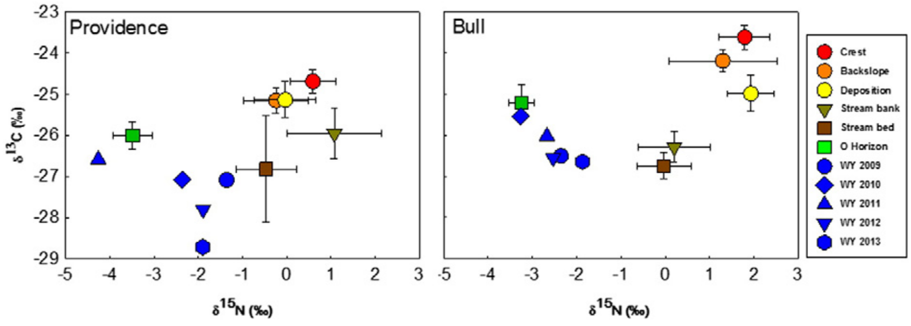

Fig. 6. Stable isotope (δ^{13}C and δ^{15}N) graphs of potential organic matter source locations and collected sediments for Providence (left panel) and Bull (right panel) catchments. Soil samples from potential source locations were collected in fall 2012 from Providence and in summer 2013 from Bull; 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 sediments (shown by individual Water Year; n = 1 composite per year). 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.

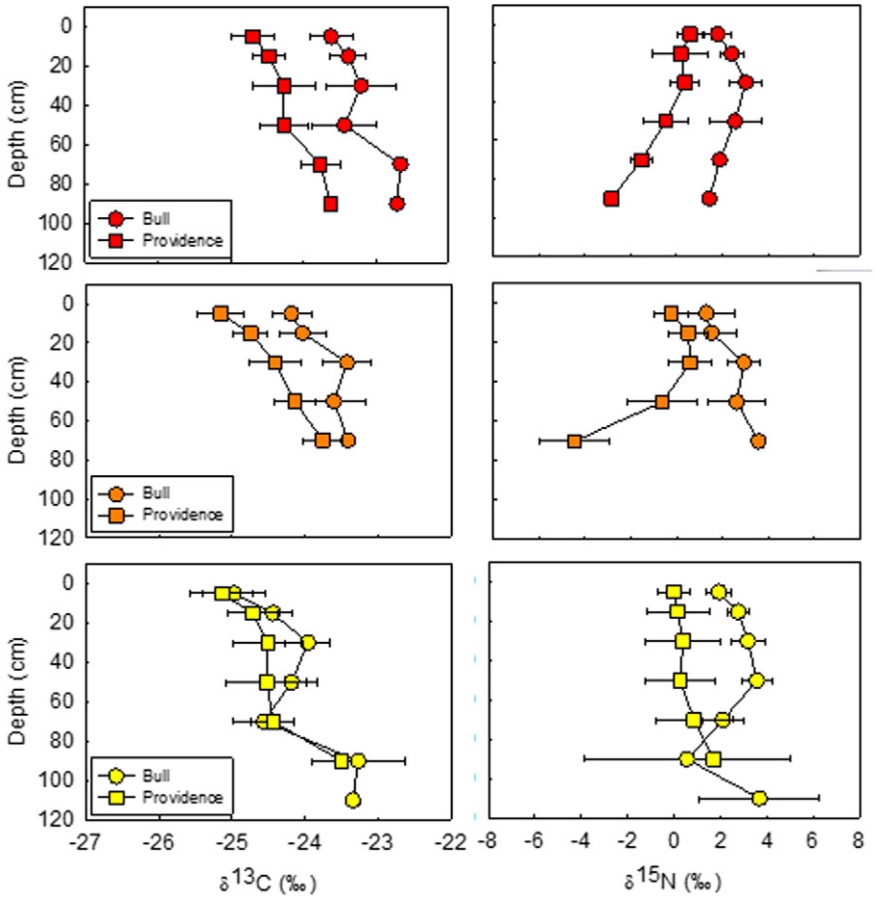

Fig. 7. Depth profiles of $\delta^{13}C$ and $\delta^{15}N$ of mineral soil from both catchments at the upland landform positions. Mean values for Providence samples are marked by squares and mean values for Bull samples are marked by circles. Red = crest, orange = backslope, and yellow = deposition/toeslope. All samples from the 6 transects are included in the averages. Error bars denote standard error of the mean for that depth class across transects within a catchment.

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 had modern radiocarbon values in the Bull catchment. Providence soils showed comparatively little change in radiocarbon values with depth (Fig. 8). In contrast, the organic C in Bull soils and stream bank samples generally became progressively older with depth (Fig. 8). All sediment samples were modern and consistently had FM values between 1.05 and 1.10 for sediments from both catchments across all years (Fig. 9).

Table 4
Two end member $\delta^{15}N$ mixing model results.

Water Year	Stream bed	O horizon	Stream bank	O horizon	Average	O horizon
Bull						
2009	0.28	0.72	0.26	0.74	0.27	0.73
2010	0.00	1.00	0.00	1.00	0.00	1.00
2011	0.18	0.82	0.17	0.83	0.18	0.82
2012	0.23	0.77	0.21	0.79	0.22	0.78
2013	0.44	0.56	0.40	0.60	0.42	0.58
Providence						
2009	0.70	0.30	0.46	0.54	0.58	0.42
2010	0.37	0.63	0.25	0.75	0.31	0.69
2011	−0.26	1.26	−0.17	1.17	−0.21	1.21
2012	0.53	0.47	0.35	0.65	0.44	0.56
2013	0.52	0.48	0.35	0.65	0.44	0.56

Two end members consist of stream bed and O horizon, stream bank and O horizon, and an average of all source materials and O horizon. The table shows the proportion of the transported organic matter in sediment attributed to be sourced from each endmember in a given WY.

4. Discussion

Of the sampled potential sources of erosion within our low-order (i.e., headwater) Sierra catchments, we found that O horizons along with stream bank and bed materials likely contribute most to eroded organic matter contained within the sediments. The $\delta^{15}N$ values suggest that exported sediments from the Bull catchment are composed mostly of O horizon material, while those from the Providence catchment generally suggest more of a mixture of the two sources, with considerable interannual variability in their relative contributions. We make these conclusions based on the similarity of the stable isotope values between the O horizon, stream materials, and sediment from both catchments across several years of contrasting annual precipitation. Our mixing model results support these qualitative graphical analyses. Deep soils were not considered as potential sources of sediment OM because both $\delta^{13}C$ and $\delta^{15}N$ values increased with soil depth beyond the values observed for sediments, suggesting that deeper soils do not contribute significantly to sediment OM. The lack of deeper mineral soil contributions to the sediments suggests that sheet and stream erosion, as opposed to rill or gully erosion, were likely the dominant means of sediment production in both our study catchments across all years.

Our data suggest that the quantity of precipitation has important effects on the amount and nature of SOM transported by soil erosion. The amount of precipitation was positively correlated to sediment yield in all of the KREW catchments over a seven-year period (Hunsaker and

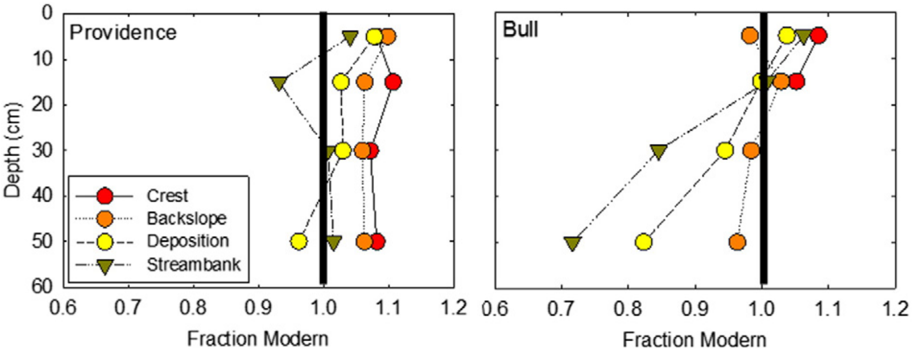

Fig. 8. Fraction modern of soil carbon (C) based on radiocarbon at representative (median values) soil and stream bank sites in Providence and Bull catchments. One transect and one stream bank point were chosen from each catchment. The vertical black line denotes the divide between modern (>1.0) and pre-modern (<1.0) values. There is no standard error associated with these measurements as only one sample from each depth was analyzed.

Neary, 2012; Stacy et al., 2015). In WY 2011, which had relatively high annual precipitation and stream discharge, the P303 stream moved more mineral sediment than the following two years (WY 2012 and 2013) of below normal rainfall. The B203 stream moved the most mineral sediment in WY 2010, which was also above normal rainfall; this amount of mineral sediment was one to two orders of magnitude larger than the following two years (i.e., WY 2011 and 2012). In addition, the effect of precipitation on amount and composition of sediment (i.e., % OM) is also likely to be modified by local topography and surface roughness; these factors can have important implications on the travel distance, velocity of mass transport by overland flow or from within-channel scouring, and fraction of eroded material that is exported from the catchments.

Typically, headwater streams are expected to have a high proportion of organics in their sediment (Hunsaker and Neary, 2012). Consistent with this expectation, we found that over the WY 2009–2013 period collected sediments in Providence and Bull catchments averaged 44% and 31% OM by mass, respectively. We also found that the proportion of OM in the sediment tended to be higher when annual precipitation was low. Stacy et al. (2015) also found that the concentration of C and N in the sediments from the KREW watersheds decreased as annual stream discharge increased. This pattern is consistent with deeper or more channelized sources of sediment material (that has lower OM concentrations) as catchment runoff increases.

The type of erosion affects the concentration of C in eroded soil. For instance, gullies can produce a large amount of sediment but the eroded material is likely to contain very little C (deep soil and associated OM), and overland flow (resulting in sheet erosion) mobilizes material of relatively smaller mass but higher in C concentration (e.g., litter and free or light OM fractions from the O horizon and surface mineral soil; Gomez et al., 2010). Other types of erosion also mobilize soil material with different C and OM concentrations. For example, landslides can mobilize material that includes soil from the entire profile, along with C-depleted saprolite and bedrock (Hilton et al., 2008). In our catchments, we observed similarities between the elemental C and N concentrations of sediments and mineral soils. Soils from both catchments we studied are developed on relatively stable land surfaces with high water infiltration rates and low landslide potential. Comparison of the $\delta^{13}C$ and $\delta^{15}N$ values of soil samples with sediment shows that, in all cases, sediment SOM is more depleted in the heavier isotopes of both elements compared to the mineral soils. The $\delta^{13}C$ and $\delta^{15}N$ values of the sediments are closer to the surficial O horizon stable isotopic values than other measured sources. These results are consistent with the observations that the amount of coarse and fine particulate organic matter in the sediments was generally between 30 and 50% (Table 3; Hunsaker and Neary, 2012).

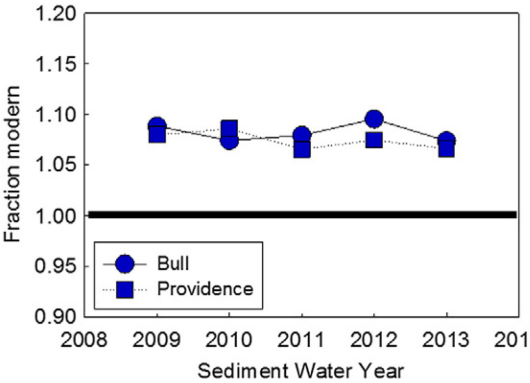

Fig. 9. Fraction modern of all sediments (WY 2009–2013) analyzed for radiocarbon in both catchments. The horizontal black line denotes the divide of modern (>1.0) and pre-modern (<1.0) values. There is no standard error associated with these measurements as only one sample from each depth was analyzed. All sediment samples are modern and show little variation between catchments and years.

4.1. Efficacy of ^{13}C and ^{15}N as tracers of eroded SOM and soil minerals

We found that combining measurements of both C and N stable isotopes was an effective method to source eroded SOM back to the location in the landscape it originated from. The stable isotope technique was efficient and relatively easy to use compared to the more expensive radioisotope methods (e.g., ^{137}Cs or ^{7}Be) that are generally employed in erosion studies. Also, using stable isotopes negates the need for non-eroded reference sites or a half-life imposed time limit needed for radioisotopes (Schaub and Alewell, 2009). Radiocarbon data can potentially provide additional insight to the origin of the collected sediments, especially when $\delta^{13}C$ and $\delta^{15}N$ values of potential sources are similar across locations and depths. Furthermore, radiocarbon can provide insight to the possible fate of these eroded materials (see discussion below). Our findings and conclusions are largely consistent with previous work in these catchments that determined sediment yield, sediment composition, and erosion patterns (Hunsaker and Neary, 2012; Stacy et al., 2015; Stafford, 2011).

Using $\delta^{13}C$ and $\delta\ ^{15}N$ to source sediments relies on several assumptions that may not always be met in erosion studies. The approach requires that significant differences in stable isotopic values among the potential sources of these materials occur within the catchment. Our study catchments were dominated by C_3 vegetation leading to relatively small $\delta^{13}C$ isotopic separation between the different potential sources we analyzed. Nevertheless, we still found sufficient differences among these sources and sediments for this approach to be applied successfully.

The method also requires that little isotopic fractionation, via decomposition or other processes, occurs in the eroded material during transport or post-deposition within the basins. Although we cannot quantify the magnitude of these compositional changes, it is clear that some alteration in the source materials occurred if O horizon and stream bank and bed materials are the sources of the captured sediments. For example, sediment OM from WY 2013 was significantly more depleted in ^{13}C than any measured source in the Providence catchment, and sediment OM from WY 2011 was significantly more depleted in ^{15}N than any measured source in Bull catchment (Fig. 6). Additionally, although potential stream sources (bank and bed) had C and N concentrations similar to the sediments, C concentrations of O horizon sources were substantially higher than collected sediments (Figs. A.3 and A.5). Preferential C utilization of labile organic functional groups (e.g., simple carbohydrates) by microbial decomposers and selective preservation of plant lignin, which is depleted in ^{13}C, could account for these changes in elemental and ^{13}C values of the sediments (Fine and Carter, 2013). It is also possible that wetting and drying cycles in the sediment basins could affect the elemental and isotopic composition of the collected OM (Berhe et al., 2012b; Kaiser et al., 2015). The sediment collected in WY 2011 that was more ^{15}N-depleted than any of our measured sources might have occurred from bacterial growth adding ^{15}N-depleted biomass to the residual material during anoxic decay (Lehmann et al., 2002). Despite these compositional changes in source materials, using multiple isotopes and elements as tracers allowed identification of the primary sources of erosion within the two studied catchments.

For OM to be representative of the origin of mineral sediment, OM must be associated with the mineral material originally mobilized, and no new OM-mineral associations can be created during transport or following deposition in the sediment basins. Given the coarse-texture of the mineral soils within these granitic catchments and the particulate nature of the eroded OM collected in the sediment basins, OM was likely eroded somewhat independently from mineral materials. Furthermore, for these same reasons, we speculate that re-association of OM with new mineral matter during transport or post-deposition was not significant enough to confound the interpretation of the erosion sources in our study. The contrasting proportion of OM in the sediments among years and between catchments suggests somewhat divergent pathways of OM and mineral soil transport via erosion from these catchments.

Finally, the materials collected in the sediment basins may not be representative of the substances eroded from the uplands. For instance, any autochthonous OM produced within the stream and deposited within the sediment basin would potentially alter the stable isotope values of the eroded OM deposited in the sediment basin, confounding the interpretation of the source of the eroded material. However, autochthonous production in headwater catchments is relatively low (Vannote et al., 1980); hence, at least in our catchments, it is unlikely that this mechanism confounded our results. Any selectivity of sediment capture within the basins could also alter the isotopic composition of the deposited and collected OM. Previous studies in the catchments showed that nutrient concentrations (C and N) found in sediments were higher in low water years (low precipitation year ~1000 mm/y; high precipitation year ~2000 mm/y) suggesting that the sediment basins were more effective at capturing smaller, nutrient-rich particles transported during years of lower annual precipitation. Less precipitation results in lower stream flows and more settling of material into the basin (Stacy et al., 2015). Despite this apparent selectivity in the OM collected within sediment basins within and among years, our results suggest that the fidelity of the coupled stable isotope — sediment basin approach is high enough to provide valuable insight into the potential sources of eroded material from montane watersheds.

4.2. Utility of radiocarbon for assessing fate of eroded SOM

Radiocarbon values support the stable isotope results that the OM contained in the collected sediments in both catchments and across all years are composed of modern, presumably less decomposed organic material. Like the stable isotope values, the sediment radiocarbon values are similar to those found for the surface upland soils and stream bank and bed materials in each catchment.

In the Bull catchment, the FM values in upland soils and stream banks decrease with soil depth, a pattern commonly observed in upland soils from stable landscape positions (Trumbore, 2000). Soil organic matter is generally older deeper in the profile because most new organic matter is deposited in the upper soil horizons and, over time, the more decomposed material is moved down the profile through leaching and vertical colloid mobilization (Bellanger et al., 2004; Bol et al., 1999; Meusburger et al., 2013; Nadelhoffer and Fry, 1988; Schaub and Alewell, 2009). Furthermore, greater association of OM with clay surfaces at depth stabilizes OM physically from further decomposition (Schmidt et al., 2011). However, in the Providence catchment, there was little change in FM of the SOM with depth down to 60 cm for stream banks and for upland soils in all landform positions. This pattern could be the result of recent profile mixing from past logging activity, historic gold mining, bioturbation, or some other disturbance that would cause deep soil mixing, or could be reflective of deeper rooting patterns within this lower elevation catchment than found at the higher elevation Bull catchment. In unstable landform positions where significant erosion or deposition is occurring, radiocarbon ages of OM with depth can also be inverted or mixed (Berhe et al., 2007). This explanation may account for the radiocarbon — depth patterns observed for the Bull backslope or Providence stream bank soils (Fig. 8). However, soil mixing is not apparent in the $\delta^{13}C$ and $\delta^{15}N$ depth profiles from upland soils in both catchments; generally, both isotopes become more enriched in the heavier isotopes of C and N with increasing soil depth, as has been found in many other upland soils from stable landscape positions (e.g., Nadelhoffer and Fry, 1988).

The recent ages of the OM collected as sediments in both catchments suggest that the OM in these eroded sediments are likely to be readily decomposed, and the eroded C returned to the atmosphere as CO_2 following catchment export. We base this conclusion on our assumption that this relatively young OM is likely weakly or unassociated with clay minerals that would stabilize these materials from further decomposition. However, if these materials are buried by additional deposition following catchment export, or if they remain in an anoxic environment where microbial decomposition is impeded, the OM exported as sediments from these catchments may remain in a similar form for extended periods of time (Berhe et al., 2007).

4.3. Soil erosion in minimally disturbed headwater catchments in a warmer world

Precipitation type (i.e., snow versus rain), along with size and momentum of rain drops, has been shown to correlate to soil erosion (Dalzell et al., 2007; Salles and Poesen, 2000). Higher soil erosion rates generally occur in montane catchments that receive more of their annual precipitation as intense rain storms, or where rain-on-snow events are common. The Providence catchment receives rain and rain-on-snow events throughout the winter that have the ability to destabilize the soil surface through rain splash and associated runoff from rapid snowmelt (Hunsaker and Neary, 2012). Recurrent flood events can continually carry shallow (i.e., relatively undecomposed, isotopically light) material, while singular large events (e.g., from rapid snowmelt due to rain-on-snow) can create gullies and move deeper material (i.e., older, more decomposed and isotopically heavy OM; Gomez et al., 2010). Similarly, intense rain events should yield material that is more enriched in heavier isotopes because it can mobilize more total soil material (less dense, C-rich material from forest floor and denser, C-poor minerals), while less intense rain events are more likely to transport less dense material that is depleted in heavier isotopes (Bellanger et al., 2004). In contrast, soil erosion in montane catchments with snow-dominated precipitation tends to be significant only briefly during snowmelt, which causes a large pulse or pulses of water to flow to and in streams (Iida et al., 2012). We used the different elevations of the two catchments as a surrogate experimental system of how climatic warming (shifts in precipitation from snow to rain) may affect soil erosion in headwater catchments of the western Sierra Nevada.

Contrary to our expectations, the annual sediment yield (kg ha^{-1}) from the snow-dominated Bull catchment was usually similar to or higher than sediment yield from the Providence catchment located within the rain-snow transition zone (Hunsaker and Neary, 2012; Stacy et al., 2015; unpublished data). Additionally, the average proportion, and interannual variation in the proportion, of sediment that is mineral matter are both higher in the Bull catchment than in the Providence catchment (Table 3). Taken together, these results may suggest that most of the mineral sediment in the snow-dominated Bull catchment originates from stream bank and bed erosion during peak stream flows following snowmelt, rather than more dispersed surficial sheet erosion throughout the catchment that is likely responsible for most of the exported OM. Furthermore, the similarity in sediment stable isotopic values among all years in the snow-dominated, Bull catchment relative to the rain–snow transition, Providence catchment suggests more consistent erosion sources in Bull than in the Providence catchment (Fig. 6). Such consistency in the isotopic values of sediments among years would be expected if erosion sources were more localized than dispersed. Given these contrasts, and those from other similar catchments within KREW (Hunsaker and Neary, 2012), we speculate that the sources and composition of soil material eroded will be more variable and dispersed as the climate warms and the rain–snow transition zone moves upwards in elevation. However, the total amount of sediment lost will still be largely determined by future changes in total precipitation amount (Hunsaker and Neary, 2012; Stacy et al., 2015) and rainfall intensity (Dalzell et al., 2007; Salles and Poesen, 2000). The shallower soils at these higher elevation catchments (Dahlgren et al., 1997), which tend to intensify the erosive energy of rainfall, along with the higher frequency of high-intensity rainfall events predicted over the Sierra as the climate warms (Das et al., 2011), suggest that we can also expect greater total erosion loses from Sierran catchments in the future.

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References

Alewell, C., Schaub, M., Conen, F., 2009. A method to detect soil carbon degradation during soil erosion. Biogeosciences 6, 2541–2547.

Bales, R.C., et al., 2006. Mountain hydrology of the western United States. Water Resour. Res. 42.

Bales, R.C., et al., 2011. Soil moisture response to snowmelt and rainfall in a Sierra Nevada mixed-conifer forest. Vadose Zone J. 10 (3), 786–799.

Battin, T.J., et al., 2009. The boundless carbon cycle. Nat. Geosci. 2 (9), 598–600.

Bellanger, B., et al., 2004. Monitoring soil organic carbon erosion with δ^{13}C and δ^{15}N on experimental field plots in the Venezuelan Andes. Catena 58 (2), 125–150.

Berhe, A.A., Kleber, M., 2013. Erosion, deposition, and the persistence of soil organic matter: mechanistic considerations and problems with terminology. Earth Surf. Process. Landf. 38 (8), 908–912.

Berhe, A.A., Harte, J., Harden, J.W., Torn, M.S., 2007. The significance of the erosion-induced terrestrial carbon sink. Bioscience 57 (4), 337–346.

Berhe, A.A., Harden, J.W., Torn, M.S., Harte, J., 2008. Linking soil organic matter dynamics and erosion-induced terrestrial carbon sequestration at different landform positions. J. Geophys. Res. Biogeosci. 113, G4. http://dx.doi.org/10.1029/2008jg000751.

Berhe, A.A., et al., 2012a. Persistence of soil organic matter in eroding versus depositional landform positions. J. Geophys. Res. 117.

Berhe, A.A., Suttle, K.B., Burton, S.D., Banfield, J.F., 2012b. Contingency in the direction and mechanics of soil organic matter responses to increased rainfall. Plant Soil 358 (1–2), 371–383.

Berhe, A.A., et al., 2014. Soil erosion controls on biogeochemical cycling of carbon and nitrogen. Nat. Educ. Knowl. 5 (8), 2.

Bol, R., Harkness, D.D., Huang, Y., Howard, D.M., 1999. The influence of soil processes on carbon isotope distribution and turnover in the British uplands. Eur. J. Soil Sci. 50, 41–51.

Collins, A.L., et al., 2013. Catchment source contributions to the sediment-bound organic matter degrading salmonid spawning gravels in a lowland river, southern England. Sci. Total Environ. 456-457, 181–195.

Condron, L., Stark, C., O'Callaghan, M., Clinton, P., Huang, Z., 2010. The role of microbial communities in the formation and decomposition of soil organic matter. In: Tilston, E.I. (Ed.), Dixon, G.R.Soil Microbiology and Sustainable Crop Production, Springer Science, pp. 81–118

Dahlgren, R.A., Boettinger, J.L., Huntington, G.L., Amundson, R.G., 1997. Soil development along an elevational transect in the western Sierra Nevada, California. Geoderma 78, 207–236.

Dalzell, B.J., Filley, T.R., Harbor, J.M., 2007. The role of hydrology in annual organic carbon loads and terrestrial organic matter export from a midwestern agricultural watershed. Geochim. Cosmochim. Acta 71 (6), 1448–1462.

Das, T., Dettinger, M.D., Cayan, D., Hidalgo, H., 2011. Potential increase in floods in California's Sierra Nevada under future climate projections. Clim. Chang. 109 (Suppl. 1), S71–S94.

De Baets, S., et al., 2012. Lignin signature as a function of land abandonment and erosion in dry luvisols of SE Spain. Catena 93, 78–86.

Doetterl, S., et al., 2016. Erosion, deposition and soil carbon: a review of process-level controls, experimental tools and models to address C cycling in dynamic landscapes. Earth Sci. Rev. 154 (102–122).

Dungait, J.A.J., Hopkins, D.W., Gregory, A.S., Whitmore, A.P., 2012. Soil organic matter turnover is governed by accessibility not recalcitrance. Glob. Chang. Biol. 18 (6), 1781–1796.

Falkowski, P., 2000. The global carbon cycle: a test of our knowledge of earth as a system. Science 290 (5490), 291–296.

Fine, S.T., Carter, B.J., 2013. Effect of sedimentation on soil organic carbon content and δ^{13}C values after 7 years of burial. Soil Sci. 178 (12), 647–653.

Fox, J.F., Papanicolaou, A.N., 2007. The use of carbon and nitrogen isotopes to study watershed erosion processes. J. Am. Water Resour. Assoc. 43 (4), 1047–1064.

Gomez, B., Baisden, W.T., Rogers, K.M., 2010. Variable composition of particle-bound organic carbon in steepland river systems. J. Geophys. Res. 115.

Gregorich, E., Greer, K., Anderson, D., Liang, B., 1998. Carbon distribution and losses: erosion and deposition effects. Soil Tillage Res. 47 (3–4), 291–302.

Harden, J.W., et al., 1999. Dynamic replacement and loss of soil carbon on eroding cropland. Glob. Biogeochem. Cycles 13 (4), 885–901.

Harden, J.W., et al., 2008. Soil erosion: data say C sink. Science 320 (5873), 178–179.

Heimsath, A., Dietrich, W., Nishiizumi, K., Finkel, R., 1997. The soil production function and landscape equilibrium. Nature 388 (6640), 358–361.

Hilton, R.G., Galy, A., Hovius, N., 2008. Riverine particulate organic carbon from an active mountain belt: importance of landslides. Glob. Biogeochem. Cycles 22 (1).

Hilton, R.G., et al., 2012. Climatic and geomorphic controls on the erosion of terrestrial biomass from subtropical mountain forest. Glob. Biogeochem. Cycles 26 (3), 1693–1705.

Hogberg, P., 1997. 15N natural abundance in soil–plant systems. New Phytol. 137, 179–203.

Hunsaker, C.T., Neary, D.G., 2012. Sediment loads and erosion in forest headwater streams of the Sierra Nevada. California, Revisiting Experimental Catchment Studies in Forest Hydrology, Melbourne, pp. 195–203.

Hunsaker, C.T., Adair, J., Auman, J., Weidich, K., Whitaker, T., 2007. Kings river experimental watershed research study plan, Fresno, CA. p. 134.

Iida, T., Kajihara, A., Okubo, H., Okajima, K., 2012. Effect of seasonal snow cover on suspended sediment runoff in a mountainous catchment. J. Hydrol. 428-429, 116–128.

Jenny, H., 1941. Factors of soil formation: a system of quantitative pedology. Dover Publications, Inc., New York.

Johnson, D.W., Hunsaker, C.T., Glass, D.W., Rau, B.M., Roath, B.A., 2011. Carbon and nutrient contents in soils from the Kings River Experimental Watersheds, Sierra Nevada Mountains, California. Geoderma 160 (3–4), 490–502.

Kaiser, M., Kleber, M., Berhe, A.A., 2015. How air-drying and rewetting modify soil organic matter characteristics: an assessment to improve data interpretation and inference. Soil Biol. Biochem. 80, 324–340.

Kim, W., 2008. Cosmogenic isotopes to disentangle natural and anthropogenic effects on erosion. pp. 1–78.

Kim, J.K., Onda, Y., Yang, D.-Y., Kim, M.S., 2013. Temporal variations of reservoir sediment sources in a small mountainous catchment in Korea. Earth Surf. Process. Landf. 38 (12), 1380–1392.

Kirchner, J.W., et al., 2001. Mountain erosion over 10 yr, 10 k.y., and 10 m.y. time scales. Geology 29 (7), 591.

Kirchner, J., Riebe, C., Ferrier, K., Finkel, R., 2006. Cosmogenic nuclide methods for measuring long-term rates of physical erosion and chemical weathering. J. Geochem. (…).

Klaminder, J., Yoo, K., Giesler, R., 2009. Soil carbon accumulation in the dry tundra: important role played by precipitation. Vol. 114 (G4), G04005.

Lal, R., 2003. Soil erosion and the global carbon budget. Environ. Int. 29 (4), 437–450.

Lehmann, M.F., Bernasconi, S.M., Barbieri, A., McKenzie, J., 2002. Preservation of organic matter and alteration of its carbon and nitrogen isotope composition during simulated and in situ early sedimentary diagenesis. Geochim. Cosmochim. Acta 66 (20), 3573–3584.

Meusburger, K., Mabit, L., Park, J.H., Sandor, T., Alewell, C., 2013. Combined use of stable isotopes and fallout radionuclides as soil erosion indicators in a forested mountain site, South Korea. Biogeosciences 10 (8), 5627–5638.

Nadelhoffer, K.J., Fry, B., 1988. Controls on natural nitrogen-15 and carbon-13 abundances in forest soil organic matter. Soil Sci. Soc. Am. J. 52 (6), 1633–1640.

Nadeu, E., Berhe, A.A., de Vente, J., Boix-Fayos, C., 2012. Erosion, deposition and replacement of soil organic carbon in Mediterranean catchments: a geomorphological, isotopic and land use change approach. Biogeosciences 9 (3), 1099–1111.

O'Leary, M.H., 1988. Carbon isotopes in photosynthesis. Bioscience 38 (5), 328–336.

Phillips, D.L., Koch, P.L., 2002. Incorporating concentration dependence in stable isotope mixing models. Oecologia 130, 114–125.

Portenga, E.W., Bierman, P.R., 2011. Understanding earth's eroding surface with 10Be. GSA Today 21 (8), 4–10.

Regnier, P., et al., 2013. Anthropogenic perturbation of the carbon fluxes from land to ocean. Nat. Geosci. 6 (8), 597–607.

Reimer, P.J., Brown, T.A., Reimer, R.W., 2004. Discussion: reporting and calibration of post-bomb 14C data. Radiocarbon 46 (3), 1299–1304.

Riebe, C., Kirchner, J., Granger, D., Finkel, R., 2001. Minimal climatic control on erosion rates in the Sierra Nevada, California. Geology.

Rumpel, C., et al., 2006. Preferential erosion of black carbon on steep slopes with slash and burn agriculture. Catena 65 (1), 30–40.

Salles, C., Poesen, J., 2000. Rain properties controlling soil splash detachment. Hydrol. Process. 14, 271–282.

Schaub, M., Alewell, C., 2009. Stable carbon isotopes as an indicator for soil degradation in an alpine environment (Urseren Valley, Switzerland). Rapid Commun. Mass Spectrom. 23 (10), 1499–1507.

Schmidt, M.W., et al., 2011. Persistence of soil organic matter as an ecosystem property. Nature 478 (7367), 49–56.

Sharpley, A., 1985. The selection erosion of plant nutrients in runoff. Soil Sci. Soc. Am. J. 49 (6), 1527–1534.

Smith, P., 2012. Soils and climate change. Curr. Opin. Environ. Sustain. 4 (5), 539–544.

Stacy, E.M., Hart, S.C., Hunsaker, C.T., Johnson, D.W., Berhe, A.A., 2015. Soil carbon and nitrogen erosion in forested catchments: implications for erosion-induced terrestrial carbon sequestration. Biogeosciences 12 (16), 4861–4874.

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 (197 pp).

Stallard, R.F., 1998. Terrestrial Sedimentation and the carbon cycle: coupling weathering and erosion to carbon burial. Glob. Biogeochem. Cycles 12 (2), 231–257.

Tarnocai, C., et al., 2009. Soil organic carbon pools in the northern circumpolar permafrost region. Glob. Biogeochem. Cycles 23 (2).

Torn, M.S., Swanston, C.S., Castanha, C., Trumbore, S.E., 2009. Storage and turnover of organic matter in soil. In: Senesi, N., Xing, B., Huang, P.M. (Eds.), Biophysico-Chemical Processes Involving Natural Nonliving Organic Matter in Environmental Systems. John Wiley and Sons, Inc., pp. 215–268.

Trumbore, S.E., 2000. Age of soil organic matter and soil respiration: radiocarbon constraints on belowground C dynamics. Ecol. Appl. 10 (2), 399–411.

Trumbore, S., 2009. Radiocarbon and soil carbon dynamics. Annu. Rev. Earth Planet. Sci. 37, 47–66.

Trumbore, S.E., Vogel, J.S., Southon, J.R., 1989. AMS 14C measurements of fractionated soil organic matter: an approach to deciphering the soil carbon cycle. Radiocarbon 31, 644–654.

Trumbore, S.E., Chadwick, O.A., Amundson, R., 1996. Rapid exchange between soil carbon and atmospheric carbon dioxide driven by temperature change. Science 272 (5260), 393–396.

Van Oost, K., et al., 2007. The impact of agricultural soil erosion on the global carbon cycle. Science 318 (5850), 626–629.

Vannote, R.L., Minshall, G.W., Cummins, K.W., Sedell, J.R., Cushing, C.E., 1980. The river continuum concept. Can. J. Fish. Aquat. Sci. 37 (1), 130–137.

Vogel, J.S., Southon, J.R., Nelsonm, D.E., Brown, T.A., 1984. Performance of catalytically condensed carbon for use in accelerated mass spectrometry. Nucl. Inst. Methods Phys. Res. B 5 (2), 289–293.

Wang, Y., Amundson, R., Trumbore, S., 1996. Radiocarbon dating of soil organic matter. Quat. Res. 45 (3), 282–288.

Yoo, K., Amundson, R., Heimsath, A., Dietrich, W., 2005. Erosion of upland hillslope soil organic carbon: coupling field measurements with a sediment transport model. Glob. Biogeochem. Cycles 19 (3), GB3003.