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**A method to detect
soil carbon
degradation during
soil erosion**

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A method to detect soil carbon degradation during soil erosion

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Abstract

Soil erosion has been discussed intensively but controversial both as a significant source or a significant sink of atmospheric carbon possibly explaining the gap in the global carbon budget. One of the major points of discussion has been whether or not carbon is degraded and mineralized to CO₂ during detachment, transport and deposition of soil material. By combining the caesium-137 (¹³⁷Cs) approach (quantification of erosion rates) with stable carbon isotope signatures (process indicator of mixing versus degradation of carbon pools) we were able to show that degradation of carbon occurs during soil erosion processes at the investigated mountain grasslands in the central Swiss Alps (Urseren Valley, Canton Uri). Transects from upland (erosion source) to wetland soils (erosion sinks) of sites affected by sheet and land slide erosion were sampled. Analysis of ¹³⁷Cs yielded an input of 2 and 2.6 t ha⁻¹ yr⁻¹ of soil material into the wetlands sites. Assuming no degradation of soil organic carbon during detachment and transport, carbon isotope signature of soil organic carbon in the wetlands could only be explained with an assumed 800 and 400 years of erosion input into the wetlands. The latter is highly unlikely with alpine peat growth rates indicating that the upper horizons might have an age between 7 and 200 years. While we do not conclude from our data that eroded soil organic carbon is generally degraded during detachment and transport, we propose this method to gain more information on process dynamics during soil erosion from oxic upland to anoxic wetland soils, sediments or water bodies.

1 Introduction

The soil organic carbon (SOC) pool contains twice the amount of the atmospheric carbon pool and is thus considered a potential major source to drive global climate change. Soil erosion is the most widespread form of soil degradation and significant amounts of carbon are either relocated to lower situated soils, water bodies and sediments or de-

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graded to CO₂ during soil erosion. The latter has been a point of intensive discussion over the last 5 years (Lal, 2004; Lal and Pimentel, 2008; Lal et al., 2004a,b; Harden et al., 2008; Renwick et al., 2004; Smith et al., 2001, 2007; Stallard, 1998; van Oost et al., 2004, 2007, 2008). While some scientists argue that erosion can be considered to decrease carbon stocks in the eroded soils and are thus a CO₂ source to the atmosphere (Lal et al., 2004a; Li et al., 2007), others have presented convincing data that most of the eroded soils re-built carbon rather fast up to former carbon contents (van Oost et al., 2004; 2007). Modeling and surveying of carbon contents of alluvial sites and sediments point to carbon accumulation at the erosion sink rather than degradation during detachment and transport of the material (Stallard, 1998; van Oost et al., 2004; Renwick et al., 2004; Harden et al., 2008; Smith et al., 2001, 2007). Whether or not erosion will be a sink or a source for carbon depends on the balance between eroded SOC degradation during detachment and transport, burial of SOC at the erosion sink site and replacement of SOC at the erosion source site. Most modeling efforts were based on rather rough assumptions such as estimated carbon contents (Smith et al., 2007) or the assumption of no SOC degradation during detachment and transport (Stallard, 1998; Smith et al., 2001). Other studies are based upon SOC inventories along eroded soil catena (Beyer et al., 1993; Jacinthe et al., 2001). Experimental studies to assess whether or not SOC is degraded during soil erosion are scarce and have mostly be bound to incubatory studies and CO₂ flux measurements under laboratory conditions (Jacinthe et al., 2002; Van Hemelryck et al., 2009). A method of direct evidence whether or not eroded carbon is merely transported to alluvial sites, sediments and/or water bodies and eventually deposited in long term sinks or if it is degraded during detachment and transport is still lacking.

Caesium-137 (¹³⁷Cs) has been used widely to quantify soil erosion rates (Walling and He, 1999a,b). The measurement of ¹³⁷Cs concentrations can provide important information on the extent of soil erosion in areas where ¹³⁷Cs occurs either due to nuclear weapon testing or the Chernobyl reactor accident in 1986. After deposition, ¹³⁷Cs is rapidly and tightly bound to the fine particles in the soil. Redistribution is mainly

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caused by physical processes where ^{137}Cs moves with soil particles (e.g. Ritchie and McHenry, 1990; Bonnett et al., 1990).

Stable isotopes have been successfully used to trace soil erosion and SOC transfer (Buck and Monger, 1999; Papanicolaou et al., 2003; Fox and Papanicolaou, 2007; Alewell et al., 2008; Turnbull et al., 2008; Schaub and Alewell, 2009; Schaub et al., 2009; Jacinthe et al., 2009). Regarding stable carbon isotope signatures, most studies have used the isotopic difference between C3 and C4 plants to track soil organic carbon transfer during erosion (e.g., Buck and Monger, 1999; Turnbull et al., 2008; Jacinthe et al., 2009). Few studies assessed soil erosion qualitatively in transects with C3 plants only using additional information like C/N ratios (Papanicolaou et al., 2003) or gradients from oxic upland soils (erosion source) to anoxic wetlands (erosion sink; Fox and Papanicolaou, 2007; Alewell et al., 2008; Schaub and Alewell, 2009). Stable carbon isotope signature of oxic upland and wetland soils have been shown to be significantly different with heavier values in the uplands than in the wetlands (Fox and Papanicolaou, 2007; Schaub and Alewell, 2009; Fig. 1). At the alpine grasslands investigated in this study, the latter was not due to differences in stable isotope signature of vegetation as the input signal to the soils (Schaub and Alewell, 2009). Rather, the enrichment in upland soils is connected to SOC degradation with preferential turnover of the lighter ^{12}C leaving the remaining SOC pool enriched in $\delta^{13}\text{C}$. In contrast, stable isotope signature of carbon in wetlands soils was very similar to the vegetation signal because of suppressed SOC degradation in these anoxic environments. Wetland soils impacted by erosion from upland soils have an intermediate stable isotope signature between upland (erosion source) and reference wetland soils (Schaub and Alewell, 2009; Fig. 1).

In a new interpretation of the data set from Schaub and Alewell (2009) we combined the ^{137}Cs approach with the stable isotope approach to quantify eroded SOC and also gain information about possible degradation of eroded SOC during detachment and transport to wetland soils (see concept for details). The aim of this study was to detect whether or not eroded SOC is degraded during detachment and transport at our exper-

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imental sites in the Urseren Valley (Canton Uri, Central Swiss Alps), which is heavily affected by sheet as well as land slide erosion.

2 Methods

2.1 Site description

5 The Urseren Valley (Canton Uri, Central Switzerland) is a W-E extended mountain valley from approximately 1500 m above sea level (a.s.l., valley bottom) up to 2500 m a.s.l. The climate is subalpine to alpine with a mean annual rainfall of 1400 mm and a mean annual air temperature of 4.3°C (meteorological station Andermatt, 1986–2007; Federal Office of Meteorology and Climatology (MeteoSwiss), Switzerland). Upland
10 soils in the study area mainly consist of cambisols, podsols and stagnosols. Histosols and histic to mollic gleysols are found in wetland areas. All wetland soils have histic characteristics up to a minimal depth of 20 cm. Dominating land use in the valley is cultivated pastures and hayfields. The valley is strongly affected by soil erosion. For a detailed description of the Urseren Valley see Meusburger and Alewell (2008).

15 Samples were collected at two erosive grassland sites on the lower southern slopes (between 1500 and 1600 m a.s.l.) between the villages of Hospental and Realp: Spissen (affected by landslide erosion), and Laui (sheet erosion). Additionally, two reference wetlands which are not directly connected to hill slopes were sampled (Oberes Moos and Höh). Vegetation of wetland and upland soils did not differ significantly in their $\delta^{13}\text{C}$ values ($-28.7\pm 0.4\%$ and of $-28.2\pm 0.1\%$, respectively; for a detailed description of vegetation cover see Schaub and Alewell, 2009). Sites are usually fertilized
20 with organic manure once a year. While the latter has a strong effect on $\delta^{15}\text{N}$, $\delta^{13}\text{C}$ of the sites were not affected by manuring (Schaub and Alewell, 2009). All sampling sites were at similar altitudes between 1500 and 1550 m a.s.l. to avoid altitudinal effects on $\delta^{13}\text{C}$ due to changes in mineralisation rates, decomposition or differences in
25 fractionation factors during plant uptake of CO_2 at different partial pressures (Schaub

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and Alewell, 2009).

2.2 Sample preparation and analysis

Five samples of 10 cm depth were analyzed at each site. Inorganic carbon was removed prior to stable isotope analysis by acid fumigation following the method of Harris et al. (2001). Stable carbon isotope analyses were accomplished using a continuous flow isotope ratio mass spectrometer (DELTA^{plus} XP, Thermo Finnigan, Bremen, Germany) coupled with a FLASH Elemental Analyzer 1112 (Thermo Finnigan, Milan, Italy) combined with a CONFLO III Interface (Thermo Finnigan, Bremen, Germany) following standard processing techniques. Stable isotope ratios are reported as $\delta^{13}\text{C}$ values [‰] relative to V-PDB defined in terms of NBS 19=1.95‰. The long term reproducibility for all standards is better than 0.1‰ (for a more detailed description of sample handling and analysis see Schaub and Alewell, 2009).

2.3 ¹³⁷Cs measurements and calculation of erosion rates

¹³⁷Cs was analyzed in soil samples of a depth of 10 cm as most of the ¹³⁷Cs is stored in this section (Schimack et al., 1989, Owens et al., 1996, Schoorl et al., 2004; Schaub and Alewell, 2009). Five to nine samples were taken at each hillslope section during summer seasons 2006 and 2007. Measurements were done with a Li-drifted Ge detector (GeLi; Princeton Gamma-Tech, Princeton, NJ, USA) at the Department for Physics and Astronomy, University of Basel. The resulting measurement uncertainty on ¹³⁷Cs peak area is lower than 15%.

The use of ¹³⁷Cs measurements to quantify soil redistribution rates is commonly based upon a comparison of ¹³⁷Cs inventories for individual sampling points to the local reference inventory. When ¹³⁷Cs reaches the soil surface by wet deposition, it is tightly adsorbed to fine soil particles. The subsequent lateral redistribution of adsorbed ¹³⁷Cs is associated with soil erosion (Sawhney, 1972). Erosion is indicated by lower ¹³⁷Cs values, while sedimentation is indicated by higher ¹³⁷Cs inventories compared

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to the reference site.

Erosion rates (S_{input}) were calculated as:

$$S_{\text{input}} = \frac{\rho \left({}^{137}\text{Cs}_{\text{wetland}} - {}^{137}\text{Cs}_{\text{reference}} \right)}{{}^{137}\text{Cs}_{\text{reference}} \cdot t_{\text{Tschernobyl}}}$$

5 with ρ =soil density (kg m^{-2}), ${}^{137}\text{Cs}_{\text{wetland}}$ and ${}^{137}\text{Cs}_{\text{reference}}$ are the ${}^{137}\text{Cs}$ contents in Bq kg^{-1} of the wetlands and the reference sites, respectively and $t_{\text{Tschernobyl}}$ is the time since the Tschernobyl accident (=22 years).

3 Concept

10 Deposited soil material in wetland soils was quantified with ${}^{137}\text{Cs}$ (see above). Mass of soil organic carbon (SOC) input to wetland systems can be calculated with %SOC of adjacent uplands (=erosion source).

$$\text{SOC}_{\text{input}} = S_{\text{input}} \cdot \% \text{SOC}_A \quad (1)$$

15 with $\text{SOC}_{\text{input}}$ =mass of SOC deposited at the impacted wetland per year ($\text{t ha}^{-1} \text{yr}^{-1}$), S_{input} =yearly soil erosion rates ($\text{t ha}^{-1} \text{yr}^{-1}$) and $\% \text{SOC}_A$ =SOC content of the upland soil (erosion source).

With the assumption that no or only negligible degradation of SOC during detachment and transport is occurring, stable isotope signature of erosion impacted wetland B can be calculated by a mere mixing model approach as:

$$\text{SOC}_B \cdot \delta^{13}\text{C}_{B \text{ calc}} = \left(\text{SOC}_{\text{input}} \cdot \text{years} \cdot \delta^{13}\text{C}_A \right) + \left(\text{SOC}_C \cdot \delta^{13}\text{C}_C \right) \quad (2)$$

20 with C_B =mass of SOC and $\delta^{13}\text{C}_{B \text{ calc}}$ =calculated stable carbon isotope signature of wetland B (erosion sink), $\delta^{13}\text{C}_A$ =stable carbon isotope signature of upland A (erosion

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source), SOC_C = mass of SOC and $\delta^{13}C_C$ = stable carbon isotope signature of wetland C (reference wetland with no influence of erosion). Theoretically, following a mere mixing model concept, mass of soil organic matter at wetland B (SOC_B) should be greater than the reference wetland (SOC_C), because it should be the sum of the yearly input of SOC (SOC_{input}) plus mass of SOC at the reference wetland (SOC_C). However, measured $\%C_B$ is significantly smaller than $\%SOC_C$ (data not shown, published in Schaub and Alewell, 2009), which is due to sediment input from uplands and/or increased degradation of SOC at wetlands B. However (again with the assumption that there is no degradation of SOC during detachment and transport of SOC), SOC_B can be calculated as:

$$SOC_B = SOC_{input} \cdot \text{years} + SOC_C \quad (3)$$

Replacing SOC_B in Eq. (2) with Eq. (3) results in:

$$\delta^{13}C_B = \frac{SOC_{input} \cdot \text{years} \cdot \delta^{13}C_A + SOC_C \cdot \delta^{13}C_C}{SOC_{input} \cdot \text{years} + SOC_C} \quad (4)$$

or:

$$\text{years} = \frac{SOC_C(\delta^{13}C_C - \delta^{13}C_B)}{SOC_{input}(\delta^{13}C_B - \delta^{13}C_A)} \quad (5)$$

Equations (4) and (5) can now be used in two ways:

1. We can assume a certain realistic time frame of erosion (e.g., 50 years) which might have influenced the uppermost 10 cm of wetland B and insert these numbers of years in Eq. (4). Our hypothesis here would be that the calculated stable carbon isotope signature of the wetland sink ($\delta^{13}C_{B_{calc}}$) should be equal to the measured signature ($\delta^{13}C_{B_{meas}}$) if there is no degradation of eroded SOC during detachment and transport. Small and/or negligible degradation of SOC should not be measurable.

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2. We can insert the measured isotope signal of erosion influenced wetland B ($\delta^{13}\text{C}_B$) into Eq. (5) and calculate the number of years of erosion to get the isotope signal we measured in wetland B.

4 Results and discussion

5 To simulate the measured $\delta^{13}\text{C}_B$ yields 800 and 419 years for the sites Spissen and Laui, respectively (Table 2). Even though the Urseren Valley was already deforested by the Romans a rough 1000 years ago and we can assume significant soil erosion ever since, this does not explain the measured stable isotope signals in the uppermost 10 cm of the wetlands. Peat accumulation rates of mires, wetlands and bogs
10 in the European Alps have been determined between 0.5 to 14 cm · 10 yr⁻¹ (Grosse-Braukmann, 1998; Jerz et al., 2000; van der Knaap and van Leeuwen, 2003; Bragazza, 2006), which results in 20–560 cm of peat accumulation in 400 years. Or, the other way round, the investigated 10 cm of peat in the Urseren Valley have an age between 7 and at the very maximum 200 years. Furthermore, sediment accumulation during erosion
15 increases “growth” rates of upper horizons which implies that the uppermost 10 cm are most likely considerably younger than 200 years. Thus, measured isotopic signatures are obviously not the result of a mere mixing process. Instead, a degradation of eroded SOC during detachment, transport and deposition can explain the shift towards heavier $\delta^{13}\text{C}$ signals in wetlands B. The latter results are congruent with experimental
20 evidence from incubation studies of Jacinthe et al. (2002) and Van Hemelryck et al. (2009) who determined up to 37% and up to 12% of eroded SOC being degraded to CO₂, respectively before SOC is deposited at depositional sites.

The underlying assumption of our approach is that SOC content and $\delta^{13}\text{C}$ would be the same in wetlands B and C if there was no erosion from upland A to wetland
25 B. An alternative explanation for observed SOC contents and $\delta^{13}\text{C}$ values would be differences in water saturation and mineralization between B and C. E.g., if water saturation at wetland C would be more permanent, the lighter $\delta^{13}\text{C}$ (less enrichment in ¹³C)

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might be due to lower decomposition rates compared to wetland B. However, wetland B is situated near the foot of the upland slopes (Fig. 1). Thus, water saturation is nearly permanent throughout the year while wetland C as the reference wetland is more in the centre of a flat valley bottom protected from erosion and from water flow from the valley slopes by smaller moraines. Furthermore, we can not exclude that wetland C has partly been drained. The latter would increase mineralization rates enriching ^{13}C in SOC of wetland C. Thus, it is more likely that we underestimate calculated years of erosion, which means we underestimate the importance of the degradation of SOC.

Our data shows for the investigated slopes in the Urseren Valley, that if eroded SOC is not degraded during detachment and transport, we have to assume 400–800 years of erosion to explain measured isotope data. The latter is highly unlikely with peat growth rates indicating that the uppermost 10 cm of the peats are not older than a maximum of 200 years. We thus conclude for the Urseren Valley that considerable degradation of eroded soil organic carbon during detachment and transport has occurred. However, our aim here is not to prove that soil organic carbon is *generally* degraded during detachment and transport of soil erosion processes. We are aware that this process will be highly site specific.

Rates of erosion (S_{input}) were calculated in this study as input of soil material to the wetland sites. However, the amount of eroded material transported is the balance between soil detachment and its retention in upland storage zones (Jacinthe et al., 2009). Konz et al. (2009) estimated approximately $30 \text{ t ha}^{-1} \text{ yr}^{-1}$ as erosion source function of the upland sites investigated in this study. However, calculated erosion input into the wetlands soils was only between $2\text{--}3 \text{ t ha}^{-1} \text{ yr}^{-1}$ (Table 1). Walling (1983) concluded that generally 70–85% of eroded material remains near the point of detachment. If the major part of the eroded material remains in the upland areas under oxic and thus favorable conditions for SOC degradation, we might assume an even higher SOC degradation.

Soil erosion rates determined with Cs-137 are connected to relatively high errors (Walling and He, 1999a,b; Schaub and Alewell, 2009). An assumed error in soil erosion

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rates of 100% at the sites Spissen and Laui (meaning a doubling of erosion rates) would result in calculated years of erosion of 432 and 209, respectively. Even though this brings us at least for the site Laui closer to realistic growth rates of the peats it still does not explain measured stable isotope signatures of the peats if we assume no soil organic carbon degradation during erosion.

5 Conclusions

While our data indicates for the Urseren Valley in the Central Swiss Alps that degradation of soil organic matter occurs during soil erosion, we are fully aware that conclusions will be site specific and can not be extrapolated for quantitative regional or even global conclusions without further investigation. Instead, we aim to propagate a method to close one gap in the so far rather speculative discussion about whether or not soil organic matter is degraded during detachment and transport and the resulting interpretation whether or not soil erosion is a source or sink of atmospheric carbon.

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Table 1. Yearly erosion rates, carbon contents and carbon isotope signatures of the investigated sites. ρ =soil density. A=upland (erosion source), B=wetland (erosion sink), C=reference wetland. $\delta^{13}\text{C}$ values with standard deviation ($n=5$).

	Erosion rate $\text{t ha}^{-1}\text{yr}^{-1}$	SOC_A %	$\delta^{13}\text{C}_A$ ‰	$\delta^{13}\text{C}_B$ ‰	SOC_C %	ρ_C g cm^{-3}	SOC_C $\text{t ha}^{-1}\text{10 cm}^{-1}$	$\delta^{13}\text{C}_C$ ‰
Spissen	2.0	5.9	-26.2 ± 0.6	-27.0 ± 0.6	34.0	0.15	51.0	-28.6 ± 0.5
Lauí	2.6	6.4	-26.7 ± 0.3	-27.5 ± 0.1				

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Table 2. Total carbon input ($\text{SOC}_{\text{input}}$) and calculated stable isotope values for wetland B as the erosion sink ($\delta^{13}\text{C}_{\text{Bcalc}}$) with the assumption of no carbon degradation during erosion. Note that the measured isotopic signal in wetland B is $\delta^{13}\text{C}_{\text{Bmeas}} = -27.0\text{‰}$ and -27.5‰ for Spissen and Laui, respectively.

Spissen			Lau		
years	$\text{SOC}_{\text{input}}$ t ha^{-1}	$\delta^{13}\text{C}_{\text{Bcalc}}$ ‰	years	$\text{SOC}_{\text{input}}$ t ha^{-1}	$\delta^{13}\text{C}_{\text{Bcalc}}$ ‰
0	0.0	-28.6	0	0.0	-28.6
50	5.9	-28.4	50	8.3	-28.3
100	11.8	-28.1	100	16.6	-28.1
200	23.6	-27.8	200	33.3	-27.8
419	49.4	-27.4	419	66.6	-27.5
800	94.4	-27.0	-	-	-

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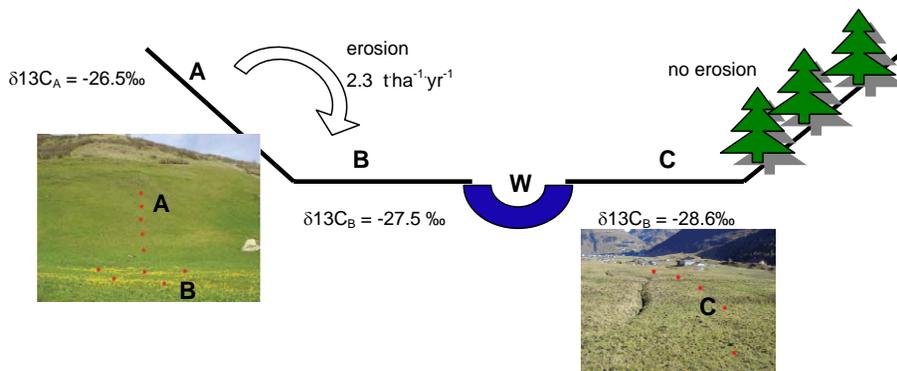


Fig. 1. Concept of tracking down soil erosion with stable isotopic signatures of carbon ($\delta^{13}\text{C}$). A=upland (erosion source), B=wetland (erosion sink), C=reference wetland, W=water body. Data from Schaub and Alewell (2009).

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