

Stable isotopes and fallout radionuclides as soil erosion indicators

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Combined use of stable isotopes and fallout radionuclides as soil erosion indicators in a forested mountain site, South Korea

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Abstract

The aim of this study is to assess and to validate the suitability of the stable nitrogen and carbon isotope signature as soil erosion indicators in a mountain forest site in South Korea. Our approach is based on the comparison of the isotope signature of “stable” landscape positions (reference sites), which are neither affected by erosion nor deposition, with eroding sites. For undisturbed soils we expect that the enrichment of $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ with soil depth, due to fractionation during decomposition, goes in parallel with a decrease in nitrogen and carbon content. Soil erosion processes potentially weaken this correlation. ^{137}Cs -method and the Revised Universal Soil Loss Equation are applied for the soil erosion quantification. The erosion rates obtained with the ^{137}Cs method range from $0.9\text{ t ha}^{-1}\text{ yr}^{-1}$ to $7\text{ t ha}^{-1}\text{ yr}^{-1}$. Considering the steep slopes of up to 40° and the erosive monsoon events (R-factor of $6600\text{ MJ mm ha}^{-1}\text{ h}^{-1}\text{ yr}^{-1}$), the rates are plausible and within the magnitude of the RUSLE- modelled soil erosion rates, varying from $0.02\text{ t ha}^{-1}\text{ yr}^{-1}$ to $5.1\text{ t ha}^{-1}\text{ yr}^{-1}$. The soil profiles of the reference sites showed significant ($p < 0.0001$) correlations between nitrogen and carbon content and its corresponding $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ signatures. In contrast, for the eroding sites this relationship was weaker and for the carbon not significant. These results verify the usefulness of the stable carbon isotope signature as qualitative indicator for soil disturbance. We could show further that the $\delta^{15}\text{N}$ isotope signature can be used similarly for uncultivated sites. We thus propose that the stable $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ signature of soil profiles could serve as a tool confirming the accurate choice of the reference site in soil erosion studies using the ^{137}Cs -method.

1 Introduction

Landslides (Lee, 2004), deforestation (Woo et al., 1995) and forest fires (Park et al., 2012) threaten the resource soil in Korea (Kim et al., 2008). Compared to these mentioned processes sheet (inter-rill) erosion of soils is a gradual process which passes

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5 mostly unseen especially on non-cultivated sites but may be responsible for high sediment loss. The quantification of soil erosion rates with traditional approaches often requires labor intensive long-term monitoring programs. As an alternative, fallout radionuclides (FRNs), and more particularly ^{137}Cs , have been used worldwide to evaluate
 10 soil erosion rates (Zapata, 2003; Mabit et al., 2008). To date, artificial fallout radionuclides (e.g. ^{137}Cs , ^{90}Sr , $^{239+240}\text{Pu}$) in Korean soils or sediments have been investigated mostly through radioecological studies and surveys in terrestrial and marine environments (Lee et al., 1997, 1998, 2012; Park et al., 2004), to assess the environmental impacts of Korean nuclear power plants (NPP) (Cha et al., 2006), to date sediment
 15 and evaluate sedimentation in ponds and lakes (Nahm et al., 2010) and to evaluate the radioactive fallout associated with the recent accident of the Fukushima NPP (Hong et al., 2012; Kim et al., 2012). However the use of FRN and especially ^{137}Cs as a soil erosion tracer (Menzel et al., 1987) for assessing soil erosion rates is scarce in Korea. ^{137}Cs is an anthropogenic isotope produced from the testing of thermonuclear
 20 weapons during the 1950s and 1960s. When ^{137}Cs fallout reaches the soil surface, it is tightly adsorbed to fine soil particles (Tamura, 1964; Tamura and Jacobs, 1960) and its subsequent redistribution is associated with soil erosion (Ritchie and McHenry, 1990). The ^{137}Cs method has the major advantage to integrate different erosion processes over a timespan of several decades (Mabit et al., 2008). However, the method has also
 25 some pitfalls (Parsons and Foster, 2011; Mabit et al., 2008). Assessment of soil redistribution rates is commonly based on a comparison of the ^{137}Cs inventory (areal activity density) at individual points in the landscape with that of “stable” landscape position (also termed reference site), where neither erosion nor deposition has occurred. One of the most critical aspects of the ^{137}Cs -method is the identification of undisturbed reference sites that allows the evaluation of the initial fallout. The latter has been shown to be connected to high small scale variability (Sutherland, 1996).

The stable isotope composition of a soil profile can provide valuable information on the disturbance or non-disturbance of a potential reference site. Stable isotopes of organic carbon (i.e. $\delta^{13}\text{C}$) and nitrogen (i.e. $\delta^{15}\text{N}$) served as a useful tool to determine

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sources and thinks of carbon, nitrogen or sediments (Alewell et al., 2008; Buck and Monger, 1999; Papanicolaou et al., 2003; Schaub and Alewell, 2009; Turnbull et al., 2008; Fox and Papanicolaou, 2007; Jacinthe et al., 2009). They are also useful for process identification and even process quantification (Conen et al., 2007, 2008). More recently, stable carbon isotopes have been used to track soil erosion and wetland degradation (Alewell et al., 2011; Schaub and Alewell, 2009). The latter studies were the first to show that isotope patterns reflect degradation processes such as soil erosion or wetland drainage. In the Swiss Central Alps a close correlation of $\delta^{13}\text{C}$ and carbon content ($r < 0.80$) was found for oxic upland soils that were not affected by soil erosion, indicating that depth profiles of $\delta^{13}\text{C}$ mainly reflect decomposition of SOC (Schaub and Alewell, 2009). Long-term disturbance of upland soils was indicated by decreasing correlation of $\delta^{13}\text{C}$ and SOC ($r > 0.80$) which went in parallel with increasing (visible) degradation at the investigated site. The degree of degradation, which in our case is the soil erosion rate will be quantified by the ^{137}Cs method and with an additional independent method; the Revised Universal Soil Loss Equation (RUSLE) (Renard et al., 1997). RUSLE is one of the most widely used soil erosion models (Kinnell, 2010). Even though it is an empirical model it proved to be useful for soil erosion quantification in very different environments as well as in Korea (Park et al., 2011, 2012; Yoon et al., 2009).

The aim of this study is to apply the method of Schaub et al. (2009) to forested mountain soils in Korea with the purpose to use in addition to the carbon stable isotopes the nitrogen stable isotope as indicator for soil disturbance. The sites indicated as undisturbed by the stable isotope approach will be subsequently used as reference site for soil erosion quantification with the ^{137}Cs method. The ^{137}Cs based erosion rates will be further compared to the modelled soil erosion rates with RUSLE.

2 Materials and method

2.1 The experimental site

The study site is located in the bowl-shaped mountain catchment of Haean (Fig. 1), 1–2 km south of the demilitarized zone (DMZ), South Korea (38°15' N, 128°15' E). The Haean basin lies in a temperate climate region with a predominantly mountainous terrain. As in many other regions of the world, this mountain environment is the major source for drinking water supply for the large cities. The Haean basin is the headwater catchment of the Lake Soyang, which provides the drinking water supply for over ten millions of people in Seoul. For more details about the study site the reader is referred to Jeong et al. (2012) and Jung et al. (2012).

The Haean basin is an intensively used headwater catchment. During the last decades, the rapid agricultural expansion and intensification transformed large areas of hill slope forests into paddies and vegetable fields, which formerly were found only at the basin bottom. Still 50 % of the Haean catchment (mainly the steep mountain ridges) are stocked with mixed deciduous forests aged about 30–40 yr and mostly dominated by *Quercus* species (Jung et al., 2010). The parent material in the basin bottom consists of highly weathered biotite granites with Anthrosols and Arenosols as predominant soil types. The surrounding mountain ridges are of metamorphic rock (Kwon et al., 1990) mainly forming acidic- and podzolic Cambisols. The intensive agriculture along with the easily erodible soils at the basin bottom constitutes the major source of suspended sediment to the drinking water reservoir Lake Soyang (Jung et al., 2012).

Mean annual precipitation recorded at the Nae-myun weather station, located 5.2 km away from the study site is 1453 ± 337 mm (1997–2009), whereof 45 % of the annual rainfall occurs in June and July, associated with the Asian monsoon (Jung et al., 2010). The mean annual temperature is 7.8°C (1997–2009).

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2.2 Soil sampling and preparation

In April 2010 soil samples were taken at steep forested slope transects and two reference sites, which were located in 0.25 km and 2 km distance of the sampling sites (Fig. 1). At each sampling point, soil cores were taken until bedrock with a soil corer (Giddings Machine Company, Windsor, CO, USA). The soil cores were used to determine the bulk densities of the samples. The selection of the reference site to estimate the initial ^{137}Cs fallout is an important step. We selected two flat forested ridge tops located in the vicinity of the sampling sites, which had not been subjected to any human disturbance. Buried land mines from the Korean War still exist in the area. Therefore, sampling was limited to sites known to be free of mines. It is generally agreed that for the estimation of the initial ^{137}Cs at least ten samples are needed (Sutherland, 1996). We took twelve reference soil profiles. For six reference profiles, samples were analyzed in 3 cm depth increments. The remaining profiles of the reference site were treated as bulk samples. Along each slope transect we took cores at five locations (approximately 30 m apart) according to slope topography. For the soil profiles of the transect samples, at least 5 depth increments were kept for stable isotope analysis. The potential erosional transects were located on slopes oriented west (transect I) and east (transect II) of a small forest stream.

Soil samples were oven-dried at 40 °C, passed through a 2 mm sieve and homogenized. The fine material (< 2 mm) was used for all further analysis. Total organic and inorganic carbon concentration (%) was measured by the RC612 Multiphase Carbon and Hydrogen/Moisture Analyzer (Leco company, MI, USA). Organic matter was calculated from organic carbon content by using the conversion factor of 1.72.

2.3 Stable isotope analysis

As even small amounts of inorganic carbon can give significant errors in $\delta^{13}\text{C}$, inorganic carbon was removed prior to stable isotope analysis by acid fumigation following the method of Harris et al. (2001). Moistened subsamples were exposed to the

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exhalation of HCl in a desiccator overnight. Afterwards, the samples were dried at 40 °C before measuring the stable isotope ratio. The N measurements were done with unacidified samples. The stable N isotope ratios and the C and N concentrations were measured by isotope ratio mass spectrometry (Thermo Finnigan Delta plus XP coupled with a Flash EA 1112 Series elemental analyzer; both instruments supplied by Thermo-Finnigan, Waltham, MA, USA). The instrumental standard deviation for $\delta^{15}\text{N}$ is 0.15 % and 0.1 % for $\delta^{13}\text{C}$.

2.4 ^{137}Cs analysis and estimation of soil erosion rates

The activities of ^{137}Cs in soil samples were determined by gamma spectrometry using a “N”-type high purity germanium detector (30 % relative efficiency; energy range 20–2000 keV) shielded with 10 cm lead and 3 mm copper. The counting time for each sample was set at 80 000 s to reach an acceptable level of detection limit and to reach a measurement error under 10 % at 2σ precision. Calibration of equipment, analysis and quality control of the measurements were performed following IAEA standard procedure (Shakhashiro and Mabit, 2009).

The areal activity (Bq m^{-2}) was calculated from ^{137}Cs activities (Bq kg^{-1}) multiplying the mass of each increment sample. The total inventory was calculated as the sum of the depth intervals areal activity over the soil profile (0 cm to bedrock) divided by the sampling device area.

The soil erosion rates were obtained from ^{137}Cs measurements using the profile distribution conversion model (Walling et al., 2002, 2011). The profile distribution model is most commonly used for natural, non-cultivated soils, where the distribution of ^{137}Cs with soil depth shows an exponential decline that can be described by the following equation (Walling and He, 1999; Zhang et al., 1990):

$$A'(x) = A_{\text{ref}}(1 - e^{x/h_0}) \quad (1)$$

where:

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- $A'(x)$ = amount of ^{137}Cs above the depth x (Bq m^{-2});
- x = depth from soil surface expressed as mass between top and actual depth (kg m^{-2});
- $A_{\text{ref}} = ^{137}\text{Cs}$ reference inventory (Bq m^{-2});
- $h_o =$ profile shape factor (kg m^{-2}), coefficient describing the rate of exponential ^{137}Cs decrease with depth.

Further we used the RUSLE model (Renard et al., 1997; USDA, 2005) to assess soil erosion rates and for comparison to soil erosion rates obtained with the ^{137}Cs method. RUSLE is an empirical model based on the following equation:

$$A = R \times K \times LS \times C \times P \quad (2)$$

where A represents the predicted long term average annual soil loss in tonnes per hectare per year; R is the rainfall factor; K is the soil erodibility factor; L and S are the slope length and steepness factors, respectively; C is the cropping-management factor; P is the support practice factor.

The R - and LS -factor were calculated according to Renard et al. (1997). The R -factor was estimated with $6600 \text{ MJ mm ha}^{-1} \text{ h}^{-1} \text{ yr}^{-1}$, based on weather data from 1999 to 2011 (Arnhold et al., 2013). K -factor was estimated for the top soil layer (upper 10 cm) using grain size- (Malvern Mastersizer 2000) and carbon data. For the C -factor we used a uniform value of C -factor = 0.001 for forested sites (Wischmeier and Smith, 1978). P -factor was neglected because no soil conservation measures have been introduced in this area.

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3 Results and discussion

3.1 Carbon and nitrogen stable isotopes as indicators of soil erosion

Measurement of bulk density showed an increase with depth as it is typically observed for forest soils (data not shown). Active bioturbation, combined with high enrichment of organic matter and thawing-freezing cycles, might have resulted in lower bulk density in the upper soil horizons.

In order to detect increased and recent erosion, mainly the upper soil horizons were examined. As sheet erosion is a surface process, a change in isotope signatures was expected to primarily occur within the top 12 cm of the soil profile. The carbon isotope signatures of the upper soil horizons tended to be higher in the reference soils than in the erosion plots, with an average of -25.8% compared to an average signature of -26.4% for the eroded upper soil horizons (Fig. 2). For the lower horizons the difference between the signature of reference soils (-24.1) and the eroding soils (-25.7) was significant ($p = 0.01$). The $\delta^{13}\text{C}$ depth profile showed an increase with depth for both transect and reference samples. Increasing $\delta^{13}\text{C}$ values with depth are typical for oxic, well drained soils (Balesdent et al., 1993; Chen et al., 2002; Novak et al., 2003), because aerobic decomposition favors selective loss of ^{12}C (Nadelhoffer and Fry, 1988; Beckerheidmann and Scharpenseel, 1989; Ågren et al., 1996).

Parallel to the depth increase of $\delta^{13}\text{C}$ in undisturbed oxic soils, we usually find a decrease in SOC content, which is also observed for both transect and reference sites. Plotting $\delta^{13}\text{C}$ against C content resulted in different patterns for the reference and erosion samples. For the both reference sites we found a highly significant ($p < 0.0001$) correlation between $\delta^{13}\text{C}$ and C content for the depth incremental samples of the 3 profiles (Fig. 3, right). Decreases in $\delta^{13}\text{C}$ concurred with decreasing C contents with increasing soil depth. For the two erosional transects (consisting of 5 depth profiles, each) we did not find a correlation ($p = 0.25$; Fig. 3, left).

A deviation from the correlation between $\delta^{13}\text{C}$ and C content of oxic soils in the Swiss Alps was proposed as indicative for enhanced soil degradation (i.e. carbon loss)

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due to soil erosion (Schaub and Alewell, 2009). The analysis of the Korean mountain soils confirms the approach for non-cultivated Korean forest soils.

Further the suitability of $\delta^{15}\text{N}$ as soil erosion tracer was assessed. $\delta^{15}\text{N}$ is enriched with soil depth due to an increasing proportion of microbially altered N transported downward in the soil profile and also aerobic mineralization (Mariotti et al., 1980). Amundson et al. (2003) hypothesized that $\delta^{15}\text{N}$ of steeply sloping soils with a short residence time due to soil removal, should have $\delta^{15}\text{N}$ values similar to the atmospheric input. Thus, the $\delta^{15}\text{N}$ value is expected to be higher for reference soils with long residence times than for the eroding slope samples. For our samples we could not observe a difference of the average $\delta^{15}\text{N}$ signature in the top soil between the reference and the erosion sites (Fig. 4). The difference is mainly masked by negative $\delta^{15}\text{N}$ values (-1.37 to -0.58) of the reference site 1 with podsollic conditions. Only for the deeper soil layers (> 12 cm) the differences became larger with higher $\delta^{15}\text{N}$ for the reference site. It is also expected that $\delta^{15}\text{N}$ for the topsoil layer increases with decreasing erosion rate. However, the comparison of $\delta^{15}\text{N}$ values of the topsoil layer to soil erosion rates showed no relation (figure not shown).

Plotting total nitrogen content (%) vs. $\delta^{15}\text{N}$, we found clearly higher correlations with $r^2 = 0.79$ and $r^2 = 0.67$ for both reference sites than for the erosion sites (r^2 is 0.27, 0.31, respectively, Fig. 5). At the reference site 2, some samples plot slightly below the trend line (non-filled dots). These are samples below a soil depth of 21cm with a higher sand content resulting in a dilution of % N.

The use of $\delta^{15}\text{N}$ signature of the topsoil as soil erosion indicator as proposed by Amundson (2003) is ambiguous. Stable soils are expected to have higher $\delta^{15}\text{N}$ signatures due to a longer time interval in which decomposition could foster fractionation. However, for stable soils with organic horizons, where decomposition is stopped due to e.g. acidic conditions the $\delta^{15}\text{N}$ signature is negative (Mariotti et al., 1980) even though the soil profile is undisturbed. Hence, the solely use of $\delta^{15}\text{N}$ as indicator of soil erosion is restricted to similar soil types with similar humus types. In contrast, the correlation between $\delta^{15}\text{N}$ and % N for the soil profile appears as a more robust indicator.

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3.2 Quantification of soil redistribution

^{137}Cs activity for the reference sites exponentially decreased with depth (Fig. 6). On average, 45 % of the ^{137}Cs activity occurred in the upper 3 cm and approximately 90 % of the total ^{137}Cs was concentrated in the first 9 cm. The exponential decline of ^{137}Cs with depth could best be approximated by a shape factor h_0 , (see Eq. 1) of 26.8. The variability of ^{137}Cs was highest for the top soil samples and decreased with depth (Fig. 6).

Two reference cores were excluded for the inventory assessment, as most likely the bedrock was not reached. The ^{137}Cs baseline inventory of the remaining 10 samples was $1607 \pm 292 \text{ Bq m}^{-2}$ (cv of 18 %) with an allowable error of 10 % at the 90 % confidence level (Mabit et al., 2010, 2012). The variability of the reference inventories is within the range for forest sites reported by Sutherland (1991, 1996) and Owen and Walling (1996). In addition, using the average annual precipitation of the area and based on Walling et al. (2002, 2011) the initial ^{137}Cs atmospheric fallout can be estimated at around 1620 Bq m^{-2} . Moreover, the inventory is also within the expected range of values assessed in a previous ^{137}Cs study on soil erosion in South Korea (Menzel et al., 1987). This study, conducted in 1986, provided ^{137}Cs reference inventories for 4 sites that ranged between 2059 Bq m^{-2} and 4958 Bq m^{-2} with an average value of 3541 Bq m^{-2} . Thus, after 24 yr of decay inventories of $1133\text{--}2728 \text{ Bq m}^{-2}$ can be expected for the different regions in 2010. The site investigated by Menzel et al. (1987), which is most similar to our site with respect to land cover and annual precipitation, is the Suweon site a grassy pine grove (Location: $37^{\circ}15' \text{ N}$, $126^{\circ}50' \text{ E}$; Annual mean precipitation = 1328 mm yr^{-1}). The 2010 decay corrected value of the fallout was 1884 Bq m^{-2} (3424 Bq m^{-2} in 1986), is within the confidence interval of our estimated reference value.

The ^{137}Cs inventories of the two forested hill slope transects ranged from 516 Bq m^{-2} to 1831 Bq m^{-2} with a mean value of 1258 Bq m^{-2} ($n = 10$). Except for three points, the locations showed a reduced ^{137}Cs values relative to the background fallout (Table 1).

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The average soil redistribution rates resulting from the conversion model for the different slope locations ranged between a sedimentation rate of $0.9 \text{ t ha}^{-1} \text{ yr}^{-1}$ to an erosion rate of $7 \text{ t ha}^{-1} \text{ yr}^{-1}$. The average net erosion rate for the transect I is $1.7 \text{ t ha}^{-1} \text{ yr}^{-1}$ and $2.2 \text{ t ha}^{-1} \text{ yr}^{-1}$ for transect II. The soils of transect II have an average soil depth of 25.2 cm compared to 42.2 cm for the transect I, which supports the differences observed in the net erosion rates between the two slopes.

The relatively small difference in erosion rates between the two slopes might be related to the different exposition. Highest erosion rates can be expected due to monsoon events with prevailing easterly wind directions, which would result in slightly higher erosion rates for at the lee slope, which is transect II. Another factor related to exposition is freezing-thawing or snowmelt erosion (Park and Woo, 1989), which would also predominantly occur at the southeast slope of transect II.

Soil erosion rates were also quantified using the empirical RUSLE model. The yielded soil erosion estimates are with $0.02 \text{ t ha}^{-1} \text{ yr}^{-1}$ to $5.1 \text{ t ha}^{-1} \text{ yr}^{-1}$ in a comparable magnitude. The average erosion rate for transect I is $3 \text{ t ha}^{-1} \text{ yr}^{-1}$ and $2.1 \text{ t ha}^{-1} \text{ yr}^{-1}$ for transect II. However, the correlation for the single sampling points is weak (0.26) and the difference between transect I and transect II is not reproduced, which supports the hypothesis that the different erosion potential of the two transects might be associated with small scale variability of rainfall erosivity and freezing-thawing (Park and Woo, 1989) or snowmelt erosion, which are both not considered in the RUSLE model. Both transects are subject to a relatively high soil loss for forested sites. On rather flat forested sites one can expect erosion rates of 0.001 to $2 \text{ t ha}^{-1} \text{ yr}^{-1}$ (Pimentel, 2006). However, considering the steep slopes and highly erosive monsoon events the obtained erosion rates are in a plausible magnitude and compare well to measured plot erosion rates of 0.3 – $3.4 \text{ t ha}^{-1} \text{ yr}^{-1}$ for another forested site in Korea (Woo, 1984).

4 Conclusions

The method proposed by Schaub et al. (2009) and Alewell et al. (2011) to use $\delta^{13}\text{C}$ vs. SOC as an indicator for soil erosion and disturbance in uncultivated soils has been tested and validated in the Korean Haeen catchment. This study also shows that as well the soil depth profiles of $\delta^{15}\text{N}$ against %N of the soil profile can be used as soil erosion indicator. The suitability of both stable isotopes should be further evaluated for the potential effects of different soil- and humus types and a broader range of soil erosion rates. The ^{137}Cs method yielded with $0.9\text{ t ha}^{-1}\text{ yr}^{-1}$ to $7\text{ t ha}^{-1}\text{ yr}^{-1}$ relatively high erosion rates for a forest site. But with respect to the steep slopes investigated and the erosive monsoon events, the rates are plausible and could also be confirmed by the RUSLE- modelled soil erosion rates, which ranged from $0.02\text{ t ha}^{-1}\text{ yr}^{-1}$ to $5.1\text{ t ha}^{-1}\text{ yr}^{-1}$. The parallel evaluation of stable isotopes and ^{137}Cs highlighted that stable isotope data could serve as a helpful tool to confirm the accurate choice of reference sites, which is the most critical requirement to use the ^{137}Cs method in soil erosion studies.

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Table 1. Erosion estimates for the different sampling points of transect I (TI) and transect II (TII). Erosion rates are presented as negative and sedimentation rates as positive values.

| sample | ^{137}Cs inventory (Bq m^{-2}) | soil redistri- bution rates ($\text{t ha}^{-1} \text{yr}^{-1}$) | RUSLE estimates ($\text{t ha}^{-1} \text{yr}^{-1}$) | distance to hilltop (m) | slope angle ($^{\circ}$) | soil depth (cm) | slope exposition ($^{\circ}$) |
|--------|--|---|---|-------------------------------|----------------------------------|-----------------------|---------------------------------------|
| T I.5 | 917 | −3.5 | −2.2 | 20 | 30 | 51 | southeast |
| T I.4 | 1030 | −2.7 | −3.4 | 50 | 30 | 41 | southeast |
| T I.3 | 1786 | 0.7 | −4.5 | 85 | 32 | 36 | southeast |
| T I.2 | 873 | −3.8 | −5.1 | 102 | 40 | 41 | southeast |
| T I.1 | 1814 | 0.8 | 0.0 | 135 | 0 | 42 | southeast |
| T II.5 | 516 | −7.0 | −1.8 | 20 | 35 | 26 | west |
| T II.4 | 1391 | −0.9 | −2.8 | 50 | 35 | 15 | west |
| T II.3 | 868 | −3.8 | −3.7 | 85 | 30 | 45 | west |
| T II.2 | 1831 | 0.9 | −2.0 | 105 | 15 | 16 | west |
| T II.1 | 1558 | −0.2 | 0.0 | 135 | 0 | 24 | west |

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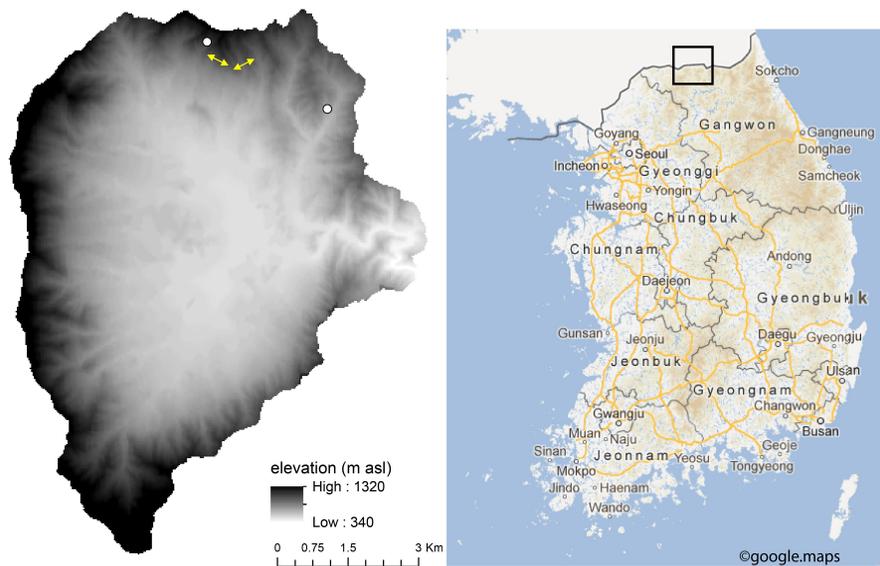


Fig. 1. Elevation model of the Haeam catchment and approximate location of the reference sites (dots) and soil erosion transects (arrows).

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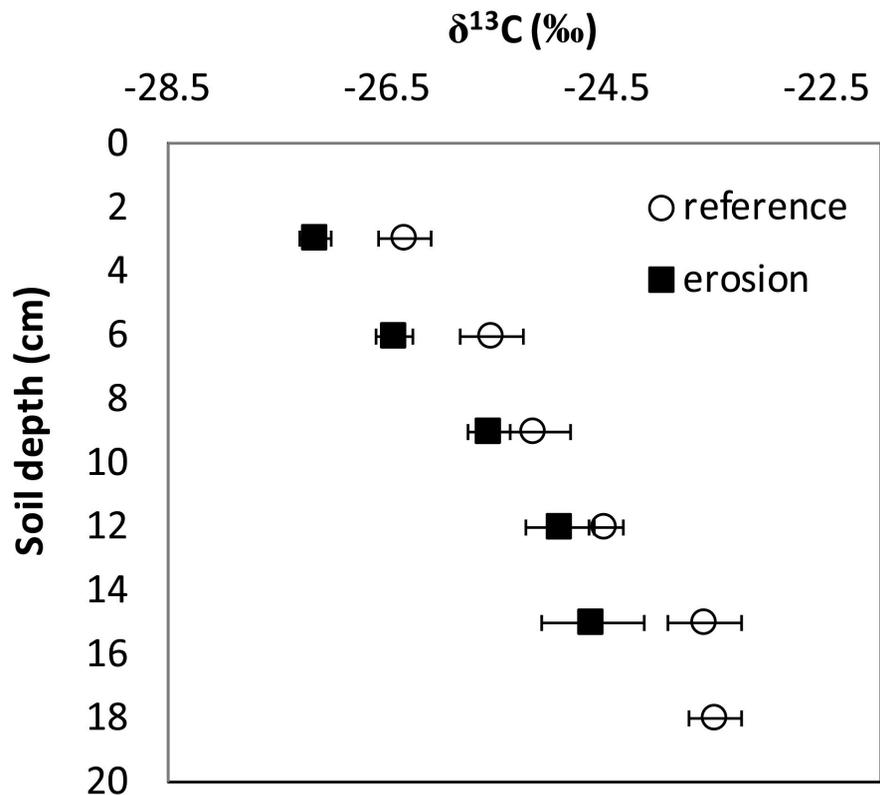


Fig. 2. Vertical distribution of $\delta^{13}\text{C}$ for reference sites and erosion transects (error bars = standard error).

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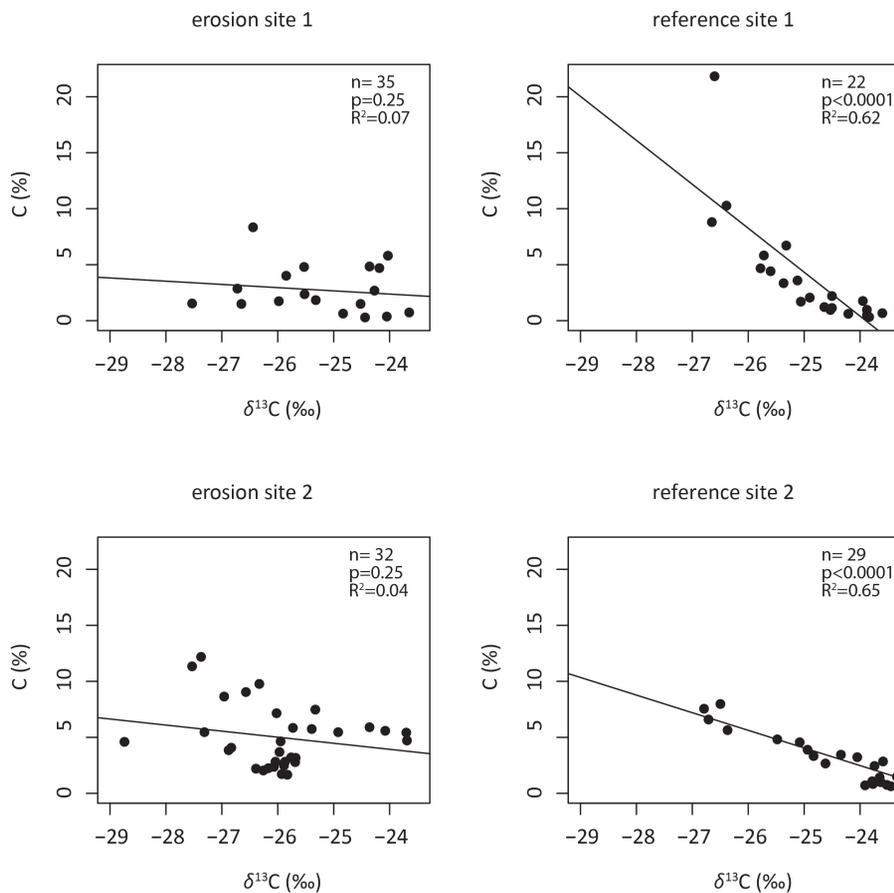


Fig. 3. Carbon content vs. $\delta^{13}\text{C}$ for reference sites and erosion transects. Sampling points of one transect are grouped in the same chart.

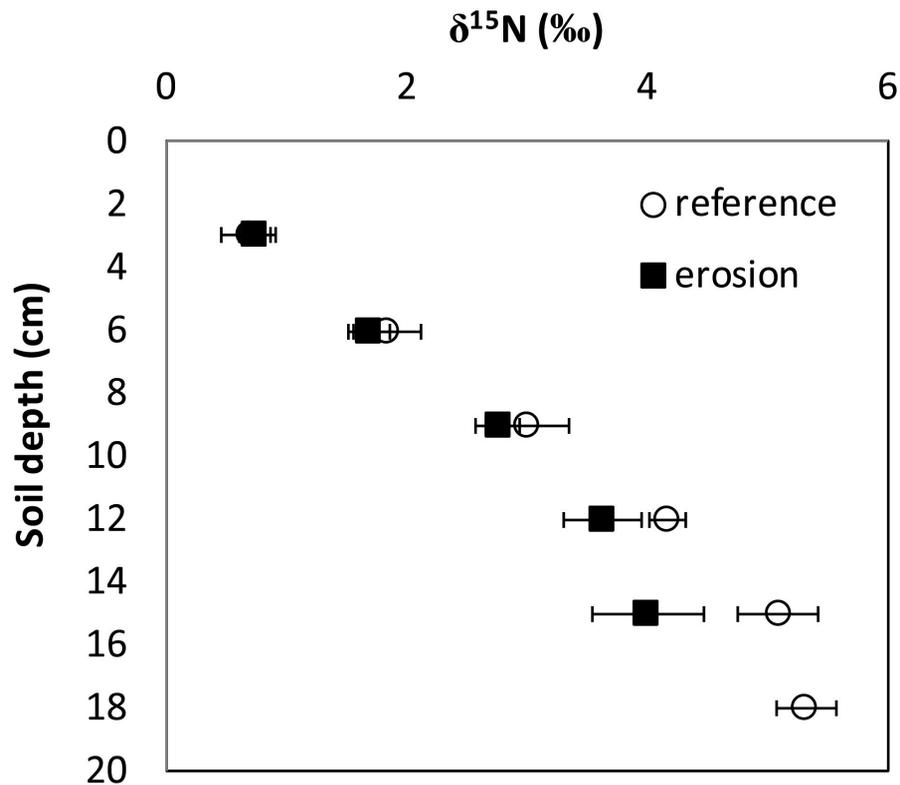


Fig. 4. Vertical distribution of $\delta^{15}\text{N}$ for reference sites and erosion transects (error bars = standard error).

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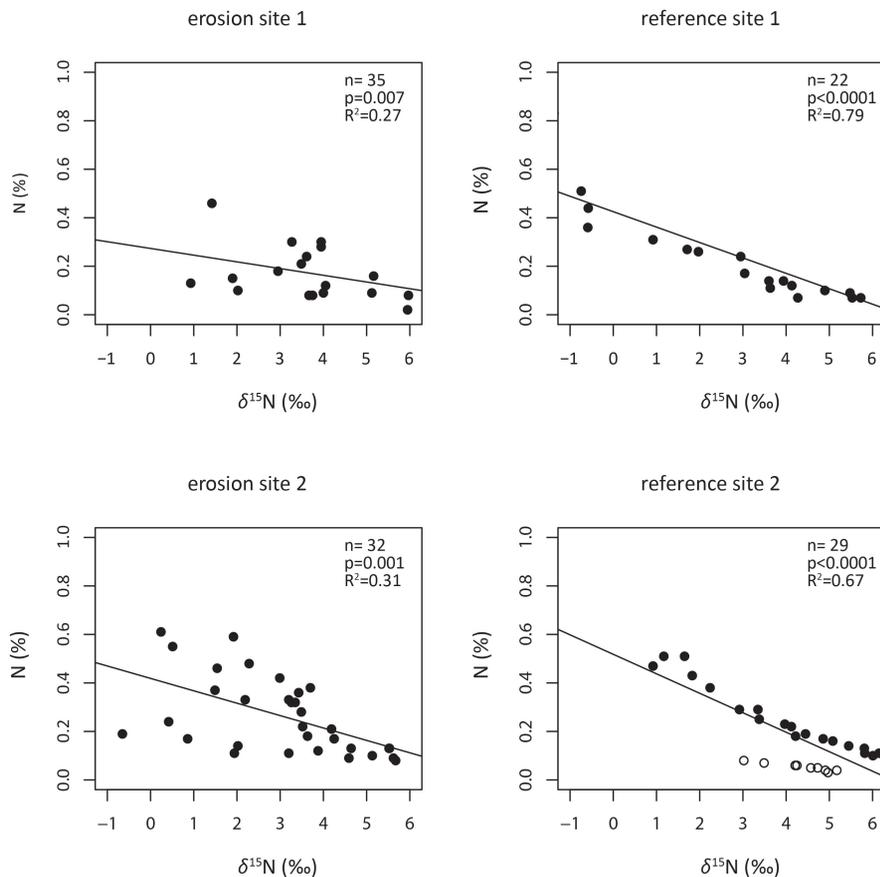


Fig. 5. Total nitrogen content (%) vs. $\delta^{15}\text{N}$ for reference sites and erosion transects. Sampling points of one transect are grouped in the same chart.

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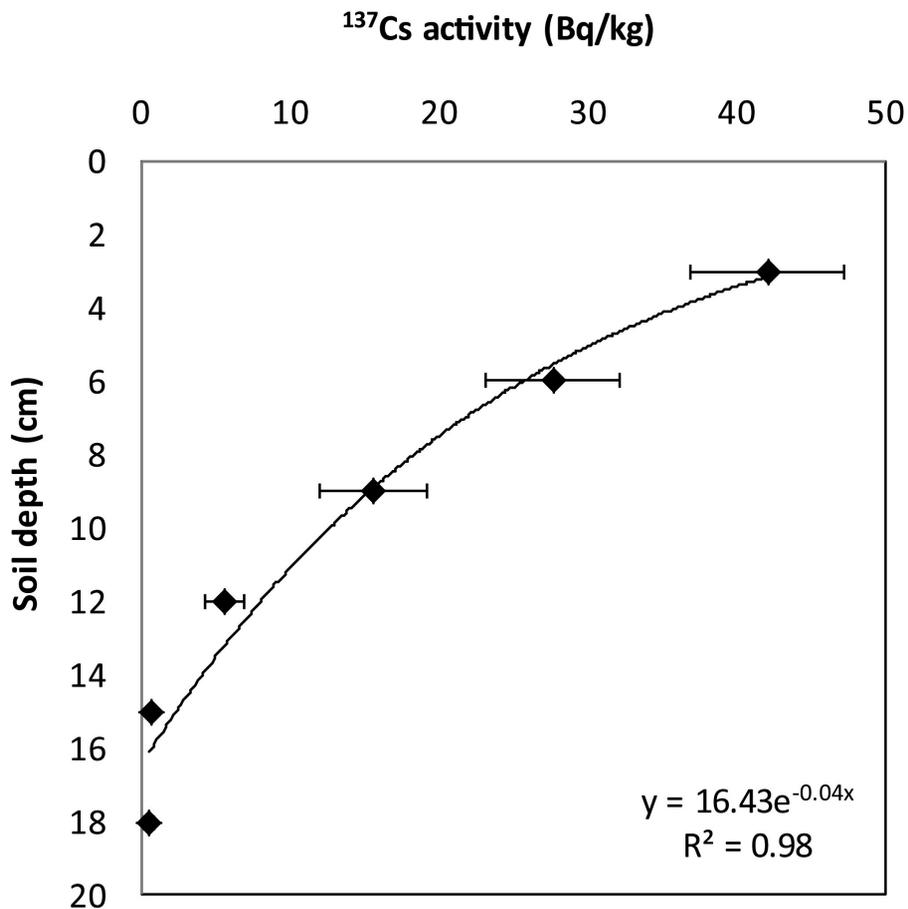


Fig. 6. Average vertical distribution of ^{137}Cs activity (Bq kg^{-1}) in the two reference sites. Error bars indicate standard error ($n = 10$), the solid line an exponential trend line.